Shock compression and adiabatic release of a titaniferous mare basalt

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Abstract—Hugoniot and release adiabat data for a high-titanium basalt (70215) of initial density 3.46 g/cm³, are reported to shock stresses of 120 GPa at which point a density of 5.7 g/cm³ is achieved along the Hugoniot. Although only three Hugoniot states were measured in the relatively low-pressure regime (7–16 GPa), this limited data set, which includes some release adiabat measurements, indicates that nearly reversible compression takes place to stress levels of ~14 GPa, above which the post-shock zero-pressure density becomes greater than 3.4 g/cm³. At shock stresses between ~14 GPa and ~50 GPa an assemblage of low- and high-pressure phases is inferred to exist along the Hugoniot curve, but has not been studied here. Release adiabat states at ~90 GPa, centered at Hugoniot states of 120 GPa, suggest the formation of a shock-induced high-pressure mineral assemblage with a zero-pressure density of ~5.2 g/cm³. Substantial (~4 GPa) elastic precursors observed in the shock compression of the terrestrial basaltic basalt and lunar gabrielite anorthosite are absent from the present study of 70215. Our results indicate that the Hugoniot elastic limit for 70215 is ~40.4 GPa. The large compressions associated with the major phase changes in the principal minerals (Fe-rich pyroxene, Ca-rich plagioclase, and feldspar) imply that larger craters will be excavated by a given impact spectrum of meteoroid masses and velocities, on mare terrains than on the less compressible, anorthositic highland terrains. Qualitatively, the present results imply either that previous mare cratering ages may have been overestimated relative to mare basin ejecta and highland units or, more probably, that the integrated meteoroid fluxes could have suffered an even sharper decline with time during the first 1.5 G.y. of lunar history than previously inferred.

INTRODUCTION

This is the first report on the dynamic properties of a rock indigenous to the mare basins of the moon. The data presented are for sample 70215, a very titaniferous basalt (58% pyroxene, 18% ilmenite, 15% plagioclase, 6% olivine, and 3% quartz by weight). This rock is probably representative of a class of the earliest mare-filling extrusive rocks which are exposed on the present lunar surface (Papke et al., 1976; Solomon, 1975). Three lines of evidence suggest that the high-titanium basalts represent the oldest of the exposed mare rocks. One is provided by unit correlation and spatial relations inferred from orbital photography (Boyce et al., 1974). The second, and probably most definitive, comes from the K–Ar plateau age of 3.78 ± 0.04 G.y. (Kirsten and Horn, 1974) for 70215 and a Rb–Sr mineral separate, isochron age for a probably related, high-titanium, low-potassium sample, 10003, of 3.84 ± 0.08 G.y. (Papanastassiou and Wasserburg, 1975). A third is suggested by the arguments summarized by Solomon (1975) who points out that these rock types have both the lowest melting points and shallowest inferred depths of origin (125–175 km) according to the high-pressure, high-temperature melting studies of Longhi et al. (1974), Kessell (1975), and Green et al. (1975), and hence evolved from a remelting of...
presumably Fe- and Ti-rich cumulates, at a time when the lunar surface of
the mare basins was both thin and young.

Because of the extremely high density of this rock, previously reported by
Mizutani and Osako (1974) and confirmed in the present study, the shallow
inferred source region, and the provenance of high-titanium basalt along the
southern boundaries of both the Imbrium and Serenitatis basins, this rock type,
or more likely, its attainment cumulate sources, may be responsible for the
density contrasts which give rise to the gravity anomalies known as mascons.
However, the presence of presumed ~3.7 G.y., high-titanium basalt at and to
the east of the Apollo 11 site and at 108N, 0° longitude (Papike et al., 1976) does
not conform to this simple view.

The above discussion and serious discrepancies among reported densities for
70215 provide further motivation for the careful measurement of the initial
zero-pressure density of this rock in the course of determining the shock
wave-equation-of-state. The main reasons for measuring the equation-of-state
are as follows:

(a) A need to understand the processes which give rise to the exotic,
    presumably shock-induced, textures and marked chemical changes within
    rocks similar to 70215 (e.g., 70109 and 79153) which have been collected
    from small impact craters on the moon (Mao et al., 1974). Equation-of-
    state data provide the thermodynamic framework within which such
    shock metamorphism can be described. The results of shock recovery
    experiments on 70215, now in progress, will complement the present
    study.

(b) We expect that cratering flow calculations based largely on the present
    shock wave data will yield quantitative information on energy partition-
    ing, ejecta distribution, shock-pressure attenuation, and crater volumes to
    be expected from a given mass-velocity meteoroid influx history on a
    mare terrane. Previous research along these lines (O’Keefe and Ahrens,
    1975, 1976, 1977) has concentrated on the description of cratering
    flows on a stronger highland anorthositic rock terrane. This work has
    demonstrated that the occurrence of a series of shock-induced phase
    changes dominates the equation-of-state behavior of highland terrane at
    stresses below 100 GPa, and that the consequent large compressions
    strongly affect the degree of meteoroid penetration and the degree of
    shock-induced melting and vaporization, as well as the spatial decay of
    the shock in the target medium (Ahrens and O’Keefe, 1977). We, and others,
    e.g., Gault and Moore (1965) have inferred that the final crater volume
    and the near-field ejecta distribution are strongly controlled by the strength
    of the target medium.

Previous work describing the equations-of-state of mare basalts have largely
been restricted to ultrasonic studies. Ultrasonic velocity and attenuation of the
ilmenite-rich, low-K lunar basalt 70215 at ambient and reduced pressures are
reported by Warren et al. (1974). Longitudinal and shear wave velocities to
0.9 GPa for 70215 are reported by Mizutani and Osako (1974), Trice et al. (1974) also report ultrasonic data to 0.5 GPa for a rock, 71055, of similar mineralogy. The elastic properties of an olivine-rich, high-K rock (sample 10057) were reported by Kanamori et al. (1970). This sample had an estimated intrinsic density of 3.38 g/cm³ comparable with that of 70215. Previous data on the dynamic yielding of rocks similar in mineralogy to 70215 are virtually nonexistent. However, shock wave measurements on evacuated (10⁻⁵ Torr vacuum) samples of a rather more sodic and weathered basalt were reported by Ahrens and Gregory (1964) and dynamic strength measurements at ambient atmospheric conditions are reported for other basalt types by Kumar (1969) and Lindholm et al. (1974). Although considerable quasi-static strength data are available for basic rocks, the effect of hard vacuum on static strengths has only recently been studied (Atkins and Peng, 1974).

In the present paper some initial data for 70215, which represents a class of very titanium-rich basalts characteristic of the Apollo 11 and 17 sites (Papke et al., 1976) are reported. Rocks of this class are the densest rocks in the present lunar sample collection. Two series of experiments were performed: the first (low-pressure) series involving measuring Hugoniot and release adiabats to 15.7 GPa with our propellant gun apparatus. In the second set of experiments, a light-gas gun was employed to yield Hugoniot data at ~120 GPa and release states at ~90 GPa.

**Sample Preparation**

For the low-pressure experiments a series of rectangular blocks (thickness 4.6 mm, lateral dimension 8-16 mm) were cut from the 70215 sample provided. Surface grinding and lapping ensured that fractures and parallelism within ±0.005 mm were achieved. The samples used in the light-gas gun experiments measured 1.3-3.8 mm in thickness and 8-13 mm in lateral dimension. Residual cutting oils and water were removed by immersion in acetone and subsequently oven-drying, and in some cases, placing the sample in an evacuated container before weighing. To ensure the accuracy of this method of removing fluids in cracks and pores connected to the free surface, the washing, drying, and evacuation procedures were repeated. Upon recycling, dry samples masses were reproducibly measured to within 1 mg. The bulk densities calculated from the masses of the dry samples and their external dimensions are reported in Tables 1 and 2. These vary from 3.32 to 3.37 g/cm³ and are considerably greater than the values reported by Warren et al. (1974) and Alvarez (1974) of 3.24 and 3.27 g/cm³ for 21 and 18 g samples of 70215, respectively. Although it is unlikely that all the pores and cracks can be completely filled upon saturation with a liquid, we nevertheless attempted to measure intrinsic densities by the Archimedes method using reagent-grade toluene and the temperature corrections of Berman (1939). As expected, this procedure gave somewhat higher densities, probably approaching the rock crystal density. The average value of "intrinsic" density obtained in this way for the specimens of the present study is 3.365 g/cm³ in good agreement with that of Mizutani and Osako (1974). Comparison of bulk and "intrinsic" densities for the larger samples (Table 1) reveals an "accessible porosity" of approximately 1.36.

**Low-Pressure Experiments**

The Hugoniot and release adiabat data summarized in Table 1 were obtained using the 40 mm propellant gun apparatus to launch projectiles bearing 3.8 mm
<table>
<thead>
<tr>
<th>Shot No.</th>
<th>Thickness (mm)</th>
<th>Crystal density (g/cm³)</th>
<th>Bulk density (g/cm³)</th>
<th>Projectile* velocity (km/sec)</th>
<th>Shock velocity (km/sec)</th>
<th>Particle velocity (km/sec)</th>
<th>Shock pressure (GPa)</th>
<th>Shock density (g/cm³)</th>
<th>Free-surface velocity (km/sec)</th>
<th>Shock velocity buffer (km/sec)</th>
<th>Release particle velocity (km/sec)</th>
<th>Release pressure (GPa)</th>
<th>Release density (g/cm³)</th>
<th>Post-shock density (g/cm³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>387</td>
<td>5.848</td>
<td>±0.005</td>
<td>3.378</td>
<td>3.338</td>
<td>5.801</td>
<td>±0.021*</td>
<td>±0.012</td>
<td>±0.003</td>
<td>±0.41</td>
<td>±0.0424</td>
<td>±0.114</td>
<td>5.118</td>
<td>±0.022</td>
<td>±0.011*</td>
</tr>
<tr>
<td>387</td>
<td>5.848</td>
<td>±0.005</td>
<td>3.378</td>
<td>3.338</td>
<td>5.870</td>
<td>±0.012</td>
<td>±0.003</td>
<td>±0.003</td>
<td>±0.381</td>
<td>±0.0003</td>
<td>±0.003</td>
<td>±0.0003</td>
<td>±0.022</td>
<td>±0.011*</td>
</tr>
<tr>
<td>386</td>
<td>3.828</td>
<td>±0.003</td>
<td>3.382</td>
<td>3.338</td>
<td>3.134</td>
<td>±0.005</td>
<td>±0.412</td>
<td>±0.005</td>
<td>±0.6323</td>
<td>±0.0024</td>
<td>±0.005</td>
<td>±0.005</td>
<td>±0.10</td>
<td>±0.004*</td>
</tr>
<tr>
<td>389</td>
<td>4.541</td>
<td>±0.003</td>
<td>3.384</td>
<td>3.334</td>
<td>1.655</td>
<td>±0.000</td>
<td>±0.005</td>
<td>±0.005</td>
<td>±0.7868</td>
<td>±0.0013</td>
<td>±0.006</td>
<td>±0.005</td>
<td>±0.10</td>
<td>±0.005*</td>
</tr>
</tbody>
</table>

*2024 Al flyer and driver plates.  
*Hugoniot elastic limit (HEL) shock state not applicable.  
*Particle velocity assumed to be one-half free-surface velocity for HEL state.  
*Not determined.  
*Uncertainty derived from uncertainty in shock velocity.  
*Combined maximum uncertainty in inclined mirror angle and reading uncertainty.  
*Uncertainty from buffer shock velocity.
Table 2. High-Pressure Magnesium and Basalt-Shell data for Tersch material, 1970.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Bulk density (g/cc)</th>
<th>Particle density (g/cc)</th>
<th>Shock velocity (m/s)</th>
<th>Release pressure (GPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LGG10</td>
<td>3.32</td>
<td>3.22</td>
<td>8.0</td>
<td>8.72</td>
</tr>
<tr>
<td>LGG11</td>
<td>3.84</td>
<td>3.34</td>
<td>8.3</td>
<td>8.90</td>
</tr>
<tr>
<td>LGG12</td>
<td>3.94</td>
<td>3.83</td>
<td>8.5</td>
<td>9.08</td>
</tr>
</tbody>
</table>

*Note: All data is approximate and may vary.*

**Uncertainty:**
- Shock velocity: ±0.10 m/s
- Particle density: ±0.01 g/cc
- Release pressure: ±0.1 GPa

**Note:**
- For copper foil and diver plates, see Eq. (G).

**References:**
- For further details, see [citation source].
thick 2024 Al flyer plates which suffer high-velocity impact with target assemblies of the type shown in Fig. 1. The experimental techniques employed have been described previously (Ahrens et al., 1972, 1973). Briefly, projectile velocities are measured via the laser beam obscuration technique, whereas resulting sample shock velocity and buffer shock velocity (Ahrens et al., 1969a) are measured by observing shock-induced loss of mirror reflectivity with an image-converter streak camera. In place of the usual plate-glass shock-arrival mirrors, aluminized polycarbonate plastic (lexan) mirrors were used so as to permit observation of the shock arrival at the driver-arrival mirror interface and at the arrival mirror free surface. (We have found that, at particle velocities of <2 km/sec, glass does not respond as readily as lexan.) This technique permitted: (a) measurement of the shock velocity in lexan mounted on the driver plate and thus deduction of the Hugoniot state in lexan, and (b) measurement of the shock velocity associated with a pressure-particle velocity state along the release adiabat of the sample using the buffer technique (Ahrens et al., 1969a).

Hugoniot states in the sample and lexan mirrors were obtained via impedance match solutions using the 2024 Al standard data given in McQueen et al. (1970). A series of inclined mirrors (Ahrens et al., 1973) mounted on the sample were set at angles such that a ~4-5 GPa elastic precursor associated with the Hugoniot elastic limit would be detected. In contrast to our experience with sample 15418,

Fig. 1. Streak camera record of shot 386, lunar sample 70215. (a) Still photograph as seen through streak camera with outline of lunar sample and flat and inclined mirrors. Superimposed image of slit indicated. (b) Resulting streak camera record showing shock arrivals through sample and lexan mirrors. The angle, y, is used to measure free-surface velocity.
a substantial elastic precursor to the main shock was not observed. The implications of this result are discussed in the next section.

The Hugoniot states obtained at 2.8 and 6.7 GPa for lexan are plotted in Fig. 2 and combined with previous data of McQueen (1974, pers. comm.) at 3.1, 4.4, 7.3, and 10.7 GPa, to constrain the lexan Hugoniot in the 2.5–4.5 GPa range of interest here. Linear regression of the five shock velocity \( (U_s) \) and particle velocity \( (u_p) \) points including the two bulk sound speed measurements at ambient pressure (cf., Ahrens, 1975) yields:

\[
u_p = 0.5404U_s - 1.128.
\]

(1)

for \( u_p, \) \( U_s \) expressed in km/sec. The correlation coefficient \( (r^2) \) for this fit is 0.98. Thus measurement of the shock velocity in the lexan mirrors mounted on the samples determines via Eq. (1) the particle velocity associated with a release adiabat state. The corresponding pressure for this release state is obtained using the Rankine-Hugoniot momentum conservation relation with an assumed zero-pressure density for lexan of 1.196 g/cm\(^3\). Two well-constrained measurements of associated free-surface velocity, which define the zero-pressure states along the two release adiabats, are also indicated in Fig. 2 and Table 1.

**HIGH-PRESSURE EXPERIMENTS**

Three very high-pressure experiments in the 117–123 GPa range were performed using a light-gas gun apparatus to accelerate 2.5 mm thick, 17 mm

![Fig. 2. Plot of low-pressure Hugoniot and release adiabat data for Ti-rich mare basalt](image-url)
diameter pure copper flyer plates, imbedded in 2.5 cm diameter lexan projectiles, to speeds of 5.8–5.9 km/sec. The details of our new apparatus are summarized in Jeanloz and Ahrens (1977). The projectile velocity is measured from 15 nsec flash radiographs taken in flight at time intervals which are related to the known positions of a mechanical break-wire system and a continuous X-ray beam obscuration station (Long and Mitchell, 1972). The streak camera is triggered when the flyer plate strikes an impact switch on the target assembly. The streak camera writes at a speed of 23 mm/s/sec for a duration of ~2 μsec. This is approximately twice the speed (and one-half the duration) of film exposure when the streak camera is used with the 40 mm apparatus.

Redundancy in our initial high-pressure experiments was sought on account of our initial uncertainties in projectile velocity and the very high density indicated by the results of our first shot.

As in the experiments with the 40 mm propellant gun, the transit time (and thus shock velocity) through the shock-arrival mirror material was measured and used to obtain a release adiabat state in the sample. In this case the Na–Ca glass mirror acts as the buffer material. Hugoniot data for the glass of density 2.491 g/cm$^3$ (composition Na$_2$O, 0.1; MgO, 0.02; SiO$_2$, 0.75; CaO, 0.13 (mass fraction)), are reported by Jeanloz and Ahrens (1977). Using their relation

$$u_s = 0.671U_s - 1.75,$$

the release adiabat particle velocity and hence release adiabat states given in Table 2 were obtained using the same procedure as in the case of the lexan mirrors for the low-pressure experiments.

**Experimental Results**

Although the number of experiments performed on 70215 was necessarily limited, it is interesting to compare the present data with those for terrestrial basic rocks and lunar sample 15418—the only other lunar rock whose dynamic compression has been examined (Ahrens et al., 1973). The uncertainties in shock pressure and density arising from errors in particle and shock velocity are distinguished in Tables 1 and 2. In some cases the uncertainty in the variable is smaller than the last significant figure indicating a negligible effect.

Aside from the marked differences in shock metamorphic response expected on account of the grossly differing mineralogies, the detailed implications of the present results for crater size, energy partitioning, and ejecta distribution associated with impact of meteoroids with velocity-mass spectra in the earth's zone (Zook, 1975), have yet to be examined. Generally, the markedly higher density (by some 0.3–0.4 g/cm$^3$) of the present Ti- and Fe-rich specimens with respect to previously studied pyroxene-plagioclase rocks (McQueen et al., 1967a; Ahrens et al., 1973) is preserved even at the highest pressures studied (120 GPa).

Detailed comparisons of this kind are presented in the next section.

On the basis of the present data, more complete Hugoniot studies of rocks and minerals (e.g., Ahrens et al., 1969b; Simakov et al., 1974; King and Ahrens, 1976).
1976) and the release data for the feldspars (Ahrens et al., 1969a; Grady et al., 1975) and pyroxene (Jeanloz and Ahrens, 1977), the Hugoniot for 70215 has been sketched in Fig. 3.

The onset of the "mixed-phase" regime is difficult to define given the limited Hugoniot data set and the absence of shock recovery data for this pressure range. However, the available Hugoniot data for pyroxenes, plagioclase, and ilmenite (see Table 3 for references) suggest that phase transformations in the major mineral constituents of 70215 will occur at dynamic pressures in excess of ~12 GPa. The release paths centered on the Hugoniot at pressure of 12.4 and 15.7 GPa are somewhat more definite. The zero-pressure post-shock densities (deduced by evaluation of the Riemann integral along a linear pressure-particle velocity trajectory [Lyzenga and Ahrens, 1977])...are respectively less, and greater than, the initial density (3.4 g/cm³) for shots 386 (12.4 GPa) and 389 (15.7 GPa). We tentatively conclude that the shock-induced phase transformations commence in the above pressure range, i.e., 14.0 ± 1.4 GPa.

Although poorly constrained in the pressure-particle velocity (and hence) pressure-density plane, the release assemblages centered at 12.4 and 15.7 GPa both

Fig. 3. Plot of Hugoniot and release adiabat data for Ti-rich mare basalt 70215 in the pressure-density plane. Previous data for gabbronorite anorthosite, 15418 (Ahrens et al., 1973) and Centreville diabase (McQueen et al., 1967a) is shown in comparison to curve calculated from the linear $U = \frac{1}{2} \sigma$. Theoretical estimate for high-pressure Hugoniot state for 15418 at 120 GPa (Table 4) is also indicated.
<table>
<thead>
<tr>
<th>Rock</th>
<th>Identification</th>
<th>Composition</th>
<th>No. of data</th>
<th>Source</th>
<th>( t_m / \gamma )</th>
<th>S (kbar)</th>
<th>( C )</th>
<th>Linear regression of high-pressure data</th>
</tr>
</thead>
<tbody>
<tr>
<td>Triassic sandstone</td>
<td>GSE 10250</td>
<td>3.98</td>
<td>7</td>
<td>Ref. 1</td>
<td>( U \geq 2.0 )</td>
<td>0.964</td>
<td>1.229</td>
<td>3.8429 ( T = 1.1 )</td>
</tr>
<tr>
<td>Triassic sandstone</td>
<td>GSE 10250</td>
<td>3.98</td>
<td>7</td>
<td>Ref. 2</td>
<td>( U \geq 2.0 )</td>
<td>0.960</td>
<td>1.229</td>
<td>3.8329 ( T = 1.1 )</td>
</tr>
<tr>
<td>Pre cambrian dike</td>
<td>GSE 10250</td>
<td>3.98</td>
<td>7</td>
<td>Ref. 3</td>
<td>( U \geq 2.0 )</td>
<td>0.964</td>
<td>1.229</td>
<td>3.8429 ( T = 1.1 )</td>
</tr>
<tr>
<td>Pre cambrian dike</td>
<td>GSE 10250</td>
<td>3.98</td>
<td>7</td>
<td>Ref. 4</td>
<td>( U \geq 2.0 )</td>
<td>0.964</td>
<td>1.229</td>
<td>3.8429 ( T = 1.1 )</td>
</tr>
</tbody>
</table>

See Appendix for footnotes.
appear to be initially extremely steep. This observation implies that shock waves in this amplitude range suffer more rapid spatial attenuation that would be expected if the Hugoniot curve itself were representative of the release adiabats, i.e., the approximation that the free-surface velocity approximately equals twice the particle velocity fails drastically. Qualitatively, the above release behavior is similar to that reported for plagioclase and quartz (Ahrens and Rosenberg, 1968; Ahrens et al., 1969a, Grady et al., 1971, 1975).

The present data are too few to define the onset of the regime in which all the mineral constituents are in their high-pressure phases. Even the existence of such a regime for a mineral assemblage as complex as that of 70215 is somewhat questionable. However, if the present Hugoniot data at ~120 GPa and the ~90 GPa release adiabat data plotted in Fig. 3 represent the behavior of such an assemblage, then a zero-pressure density of the order of 5.2 g/cm³ is indicated.

The fact that the present low-pressure results do not display the 4.5 GPa elastic precursor observed in other igneous rocks, including the weathered, and somewhat glassy, Vacaville basalt (Ahrens and Gregson, 1964) presents something of an enigma. The explanation may lie in the fact that the present samples are low in feldspar and high in lime-slate. Previous work such as that on lunar sample 15418 and various feldspars (Ahrens et al., 1969a; Ahrens and Liu, 1972; Simakov et al., 1974) all display prominent elastic precursor shocks. In fact we would infer, on the basis of the capacitor (but not the optical) records presented by McQueen et al. (1967a), that the diabases studied displayed a ~4 GPa elastic precursor contrary to their conclusion. The present value of the Hugoniot elastic limit (Table 1 and Fig. 2) should be considered a firm upper limit and an unexpected, but highly significant result of the present investigation. The implications of this observation are discussed in the final section.

**Discussion of the High-Pressure Data**

The Hugoniot data for 70215 (Table 2) indicate an average density of 5.69 g/cm³ at ~120 GPa. In order to compare these new results with existing data for component minerals or terrestrial rocks one must first estimate the densities of the latter at 120 GPa. This is a non-trivial problem since the Hugoniot data are unevenly distributed throughout the pressure range of interest and also exhibit appreciable scatter. We have estimated the densities (at 120 GPa) of all relevant rocks and minerals from the available Hugoniot data according to the procedure outlined below, which is similar to that used by McQueen et al. (1970) for inferring Hugoniots of non-resecting mixtures.

Using the $U_p - U_s$ plane to define the "high-pressure phase" regime for a series of rocks and minerals we have fitted the $(U_p, U_s)$ data with the linear relationship:

$$U_p = C + Su_s$$

(3)

The onset of this regime at particle velocities of 1.8-2.8 km/sec is, in general, readily identified and in most cases all available data at higher particle velocities
than this critical value are well described by a single linear trend. The data for quartz (Trunin et al., 1970) and forsteritic olivine (Trunin et al., 1965) provide exceptions to this general rule. For particle velocities greater than 5.8 km/sec the points \( x_1, U_1 \) are consistently displaced from the linear trend in the direction of higher particle velocities for a given shock velocity. These ultrahigh pressure data \( (P > 200 \text{ GPa}) \) are suggestive of further phase changes in that regime but are not relevant to the estimation of density at 120 GPa and are therefore excluded from the present analysis. The parameters \( C \) and \( S \) listed in Table 3 are based on the available data for low porosity, essentially monomineralic specimens of quartz, feldspar, pyroxene, olivine, and the common opaque accessory minerals. Also shown in Table 3 are fits for the data for some terrestrial pyroxene-plagioclase rocks and Westerly granite. The proximity of the correlation coefficient to 1 \( (r^2 > 0.96, \text{ throughout}) \) indicates the quality of the linear \( U_1 = u_i \) fits.

By substitution of (3) into the Rankine-Hugoniot conservation equations one may calculate the \( P(\rho) \) trajectory

\[ P = \rho c_s^2 \left( \rho \rho_* - 1 \right) \left[ S + (1 - S) \rho \rho_* \right], \tag{4} \]

from the tabulated values of \( \rho_* C, \) and \( S. \) In this way, the densities at 120 GPa for each mineral and rock of Table 3 are calculated. The scatter present in the Hugoniot data for the most intensively studied minerals suggests that these calculated densities are uncertain to the extent of \( \pm 0.1 \text{ g/cm}^3. \)

It is immediately apparent from Table 3 that the titaniferous lunar basalt (70215) of the present study is about 0.35 g/cm\(^3\) (or 6\%) denser at 120 GPa than the densest terrestrial pyroxene-plagioclase rocks for which Hugoniot data are available. The high zero-pressure density \( (\rho_0) \) of 70215 compared to the terrestrial pyroxene-plagioclase rocks prompts a closer examination of the comparative mineralogies of these rocks.

The detailed chemical analysis of 70215 by Dynek et al. (1975) indicates that it is strongly enriched in FeO and TiO\(_2\), and depleted in SiO\(_2\), Al\(_2\)O\(_3\), and alkalis by comparison with terrestrial basalts. These chemical differences result in substantial mineralogical contrast between 70215 and the terrestrial basalts. The former is characterized by less (and more calcic) plagioclase, more titaniferous augite, and most importantly, by significant quantities of modal ilmenite. In Table 4 the zero-pressure specific volume (and hence density) is calculated for each rock by addition of the specific volume \( (1/\rho_0) \) of the component minerals weighted by their mass fractions, that is

\[ 1/\rho_0 = \sum (m/\rho_0). \tag{5} \]

It is evident that the zero-pressure densities of Westerly granite (USGS G1), Centreville diabase (USGS W1), and lunar gabbric anorthosite 15418 are satisfactorily reproduced given the uncertainties in mineral densities and rock porosities. In the case of 70215 it is difficult to estimate the pyroxene density. However, a reasonable value of 3.30 g/cm\(^3\) satisfies the zero-pressure density of
Table 4. Hugoniot mixing calculations for Westerly granite, Centreville diabase, Ti-rich mare basalt 70215, and lunar gabbroic anorthosite 15418.

<table>
<thead>
<tr>
<th>Chemical composition</th>
<th>Westerly granite (USGS standard GI)</th>
<th>Centreville diabase (USGS standard W1)</th>
<th>Ti-rich mare basalt 70215</th>
<th>Lunar gabbroic anorthosite 15418</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>a' (Mpa)</td>
<td>a'' (Mpa)</td>
<td>a' (Mpa)</td>
<td>a'' (Mpa)</td>
</tr>
<tr>
<td>Quartz</td>
<td>29.4</td>
<td>2.45</td>
<td>3.12</td>
<td>2.05</td>
</tr>
<tr>
<td>K-Feldspar</td>
<td>26.2 (k-Feldspar = 2.26)</td>
<td>3.76</td>
<td>2.76</td>
<td>5.89</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>25.8 (An10-10)</td>
<td>2.64</td>
<td>2.76</td>
<td>3.14</td>
</tr>
<tr>
<td>Pyroxene</td>
<td>25.1 (Diopside, An60)</td>
<td>2.64</td>
<td>2.76</td>
<td>2.76</td>
</tr>
<tr>
<td>Glass</td>
<td>2.2</td>
<td>4.25</td>
<td>6.66</td>
<td>4.66</td>
</tr>
<tr>
<td>Magnesite</td>
<td>5.8</td>
<td>7.25</td>
<td>7.50</td>
<td>6.90</td>
</tr>
<tr>
<td>Total</td>
<td>100.0</td>
<td>Calculated</td>
<td>99.1</td>
<td>Calculated</td>
</tr>
</tbody>
</table>

See Appendix for footnotes.
the rock. Moreover, it is clear that the presence of a significant amount of the relatively high-density ilmenite explains the high zero-pressure density of 70215.

To proceed further with an analysis of the Hugoniot data for 70215 it is necessary to establish a procedure by which the Hugoniot data for the rock may be expanded in terms of Hugoniot data for its constituents. For example, McQueen (1968) and Al'tshuller and Sharipzhano (1971) have demonstrated that the Hugoniots for many silicates in the pressure range 80–150 GPa may be approximately reproduced by addition of the Hugoniots for the components of the isochemical oxide mixture. This observation suggests that the silicate phases present under conditions of extreme shock compression, whatever their detailed crystal structure, are characterized by the elastic and thermal properties of the simple close-packed oxides under the same conditions. Birch (1952) reached a similar conclusion concerning the silicate phases of the earth's lower mantle.

In this paper we investigate an alternative approach: construction of rock Hugoniots by addition of mineral (rather than oxide) Hugoniots. If we make only the usual assumptions of shock wave theory—stress uniformity and negligible heat (and mass) transfer—then the density of the rock along the Hugoniot is given by an expression formally identical to Eq. (5) where now, the Hugoniot densities at a given pressure are to be used. In this calculation the temperature distribution within a mineral grain at a given shock pressure is independent of the nature of surrounding grains since no heat enters or leaves the system via grain boundaries. We shall also use Eq. (5) to interpolate among Hugoniot densities for end members of ambient-pressure solid solutions. At a given pressure, the high-pressure phase of the solid solution is likely to be either a solid solution among, or a physical mixture of, the end-member high-pressure phases. Application of Eq. (5) to the former situation requires ideal solid solution and identical thermodynamic properties of the end members. It appears probable that no serious errors in calculated densities will arise from these assumptions.

We can now apply this scheme to rocks such as Western granite and Centreville diabase for which detailed chemical, mineralogical, and Hugoniot data are available (Tables 3 and 4). Hugoniot densities at 120 GPa have been estimated for each phase (see notes following Table 4 for details) and the rock Hugoniot density at 120 GPa calculated according to Eq. (5). The good agreement between calculated and observed Hugoniot densities (G1 = 3.09 and 3.07, W1 = 3.33 and 3.33), especially for W1 where the range of mineral densities is considerable, demonstrates the reliability of this form of additive equation-of-state.

Having thus demonstrated the need to assess the Hugoniot data for 70215 in terms of its rather distinctive mineralogy and having devised and successfully tested a scheme for the calculation of rock Hugoniots from those of the component minerals we shall now apply this scheme first to 70215 and then to lunar gabbrorics anorthosite 15418 for which the high-pressure Hugoniot has not yet been studied.

In the case of 70215 we do not have enough Hugoniot data for appropriate pyroxene compositions to infer the density of the average pyroxene (Table 4) at
120 GPa. As for the zero-pressure calculation, we may calculate the pyroxene density which is consistent with the rock density. The pyroxene has an inferred density of 3.33 g/cm$^3$ at 120 GPa which lies well within the range of densities spanned by the available data for En$_{95}$Fs$_{10}$, Wo$_{80}$Fs$_{20}$, and Jd$_{90}$. Thus, by using a very reasonable density for the pyroxene and the Hugoniot data for the other minerals, the new 70215 Hugoniot data are readily explained, and in Fig. 3 the contrast between 70215 and the terrestrial pyroxene-plagioclase rocks is emphasized.

It is a straightforward task to apply this scheme to any rock whose chemistry and mineralogy are well defined provided the phases lie in the compositional ranges spanned by the "standard" minerals of Table 3. In particular, the lunar gabbric anorthosite 15418 is such a rock and Table 4 and Fig. 3 include an estimate of its Hugoniot density at 120 GPa. The latter estimate agrees closely with previous predictions (Arens et al., 1979) of the high-pressure-density Hugoniot for 15418 obtained using the theory of interacting continua and the mineralogic analyses available at that time.

**Conclusions and Implications**

Lunar basalt 70215 appears to be among the densest rocks in the present lunar sample collection, having a crystal density of 3.38 g/cm$^3$ and a porosity of ~1.3%. The density of this rock type is important as the provenance of the very high-titanium mare basalts within the older mare basins may correlate with the location of the mascon gravity anomalies. The postulated high Fe and Ti cumulates, which produce partial melts consistent with the composition of such titaniferous basalts, are candidate materials for the origin of the mascon gravity anomalies.

Upon compression in the low-pressure regime, the relatively high amplitude elastic precursor observed in terrestrial basalts and lunar gabbric anorthosite are absent. An upset bound of ~0.4 GPa can be placed on the Hugoniot elastic limit of 70215. Post-shock densities, derived from release adiabat measurements, are respectively 3.9 and 3.0 g/cm$^3$ for release from shock pressure of 15.7 and 12.4 GPa compared to an initial zero-pressure density of 3.4 g/cm$^3$. We infer that the "mixed phase" regime extends from 14.0 ± 1.4 GPa to pressure of at least 50 GPa.

The very high density of 70215 (5.69 g/cm$^3$ at pressure ~120 GPa) is not inconsistent with the available Hugoniot data for the component minerals, although the lack of data for diopside, titaniferous and aluminous pyroxenes prohibits a detailed comparison. The release adiabats centered on Hugoniot states in the 120 GPa range suggest the presence of an extremely dense shock-induced high-pressure phase assemblage with a zero-pressure density on the order of 5.2 g/cm$^3$.

The present results have important implications for both the degree of shock metamorphism expected for impact processes and the extent of ejecta transport on mare surfaces with high-titanium basalt composition. Although the impact-
induced flows have yet to be explored in detail, previous calculational results (e.g., O’Keeffe and Ahrens, 1976), suggest that some quantitative differences may be expected for impacts on mare (versus highland terrane) surfaces.

The higher shock impedance of Ti-rich mare basalt (relative to highland crust) means that more material will be shock metamorphosed for a given population of impacting objects. Also, the markedly lower dynamic yield strength of the mare rocks will result in greater excavation for craters less ... 10^4 cm diameter. As pointed out by Neukum and Wise (1976), strength effects can in effect, shift crater diameter versus cumulative number curves. This is probably also true for marked differences in compression. Thus, we expect that when mare and highland surfaces of the same age exposed to the same meteoroid flux, (i.e., specified by a mass, composition and impact velocity distribution function) are compared, because the mare rocks suffer greater compression, the mare surface will appear, on the basis of crater size versus number, to be older. Since it is already well established that the present mare have considerably younger apparent crater ages than the ejecta blanket produced by excavation of the mare basins, we expect that future work will demonstrate an even more rapid decay of the meteoroid infall rate than has been previously inferred (Shoemaker, 1970) for the first 1.5 G.y. of lunar history.

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REFERENCES


APPENDIX

Footnotes to Table 3

1. McQueen et al. (1967a).
2. Trunin et al. (1965).
5. Trunin et al. (1970).
7. Simakov et al. (1974).
11. McQueen and March (1966).
12. McQueen et al. (1967b).

*Very high-pressure data of Trunin et al. (1970) excluded (see text).
*Inspection of U - u_0 diagram suggests p > 5 GPa is a minimum value.
*Corrected for presence of 10% quartz (u_0 = 3.3, u(12 GPa) = 5.3).
*Following the suggestion in the caption of Fig. 15 of McQueen et al. (1967a) the data for u_0 > 1.9 km/sec are considered as two linear U - u_0 segments.
*Two ultrahigh pressure data of Trunin et al. (1965) are excluded from fit (see text).
*A 5% increase of slope over the fit to the two data of Simakov et al. (1974) produces a P(u) trajectory more concordant with the reflected shock data of King and Ahrens (1976).

Footnotes to Table 4

*C*): W1: Analyses from Fleischer (1969) recalculating volatile-free.
70215: Calculated bulk composition of Dynek et al. (1973) with trace elements removed.
15418: LSPET (1972) analyses recalculating volatile-free.
*G1*: Chayes (1951) recalculated mica-free.
*W1*: Chayes (1951) recalculated free of mica and non-opaque accessories. Opaque, plagioclase and pyroxene compositions assumed normative.
70215: Dynek et al. (1973) recalculated free of accessory phases. Pyroxenes have been lumped to yield a single pyroxene (of equivalent composition) for the Hugoniot calculations.

*Zero-pressure densities from Deer et al. (1966).*

*Plagioclase densities estimated from the oligoclase and andesine Hugoniot data (Table 3) and the systematics of u_p with composition.*

*Estimated from EnbFe_3S_4 and WO_3F_3S_4 Hugoniot data (Table 3), En_0.07Fe_0.93S_4 = 0.5 (EnbFe_3S_4) + 0.5 (WO_3F_3S_4), En_0.07Fe_0.93S_4 = 9.06 (EnbFe_3S_4) + 0.34 (WO_3F_3S_4). Density of EnbFe_3S_4 = 5.2 g/cm^3 at 120 GPa.*

*Required to explain observed rock densities.*

*Estimated from FeTiO_3, Hugoniot data (Table 3) and relative MgTiO_3 and FeTiO, zero-pressure densities.*

*Averaged of the pyroxene compositions described by Dynek et al. (1975): X_{Mg} = [R^2TiAl(OH)_4Mg^2AlSiO_4]_x[SiO_4]_y, R^2TiO_4 = (Mg, Ca)Mg_2Al_3Si_2O_10.*