SHOCK WAVE PROPERTIES OF ANORTHOSITE AND GABBRO

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Abstract. Shock wave experiments have been conducted on San Gabriel anorthosite and San Marcos gabbro to peak stresses between 5 and 11 GPa using a 40-mm-bore propellant gun. Particle velocity wave profiles were measured directly at several points in each target by means of electromagnetic gauges, and Hugoniot states were calculated by determining shock transit times from the gauge records. The particle velocity profiles yielded sound velocities along the release adiabats which indicate a retention of shear strength upon shock compression for anorthosite, with a loss of strength upon release to nearly zero stress. Sound velocities of anorthosite shocked to peak stresses between 6 and 10 GPa were measured to be between 5.1 and 5.3 km/s upon release to nearly zero stress as compared to ~6.9 and 5.4 km/s for the expected longitudinal and bulk wave speeds. Stress density release paths in the anorthosite indicate possible transformation of albite to jadeite + (quartz or coesite), with the amount of albite transformed ranging from as low as 0.05 to as much as 0.19 mass fraction in the 6-10 GPa shock stress range. Electrical interference effects precluded the determination of accurate release paths for San Marcos gabbro. Because of the apparent loss of shear strength during unloading from the shock peak stresses, the fluidlike rheology of anorthosite which is indicated implies that calculations of energy partitioning due to impact upon planetary surfaces based on elastic-plastic models will underestimate the amount of internal energy deposited in the impacted surface material.

Introduction

The plagioclase feldspar-bearing rocks, anorthosite and gabbro, are important components of the lunar and terrestrial crusts. It is necessary to understand the behavior of such rocks under high dynamic stress in order to model cratering processes which result from hypervelocity impacts and to characterize the stress history of rocks which have been subjected to shock loading on planetary surfaces (such as the moon) and meteorites. Shock wave studies of these and similar materials have been conducted in the past [Ahrens et al., 1969; McQueen et al., 1967; Jeanloz and Ahrens, 1980; Boslough et al., 1985a], but in these studies the data are limited to the Hugoniot state and in some cases a single state on the release isentrope. By employing particle velocity gauges, a complete stress-strain history subsequent to shock compression can be determined, along with sound velocity information [Fowles and Williams, 1970; Cowperthwaite and Williams, 1971; Seaman, 1974]. Particle velocity experiments supply detailed release paths, which provide better constraint for mechanical properties and polymorphism than is available with Hugoniot experiments alone.

Peterson et al. [1970] used particle velocity gauges to determine release paths of plays alluvium, tonalite, and novaculite shocked to stresses up to 5 GPa. They attributed high rarefraction velocities and steep release paths in the stress-density plane to irreversible compaction. Grady et al. [1974] carried out experiments on polycrystalline quartz (novaculite) to 40 GPa, using a combination of particle velocity and manganin stress gauges to determine release adiabats. They concluded that a partial quartz-stishovite transformation takes place above 15 GPa, with the quantity of material transformed an increasing function of peak stress, and that the Hugoniot states are not on the quartz-stishovite coexistence curve. Similar experiments were conducted on polycrystalline quartz and perthitic feldspar by Grady et al. [1975] and Grady and Murri [1976], who used manganin stress gauges to determine Hugoniot sound velocities and found that these rocks lose shear strength when shocked to pressures above 20 GPa. Larson and Anderson [1979] used particle velocity gauges to study limestone and tuff at lower stress levels (4 GPa) and attributed the observed time-dependent behavior to the closing of pores in these rocks.

In this paper we present new Hugoniot data on San Gabriel anorthosite and San Marcos gabbro to 11 GPa. Release paths in the stress-density plane and sound velocities are reported, as determined from particle velocity data.

Experimental Methods

The particle velocity experimental design is similar to those of Grady et al. [1974], Larson and Anderson [1979], and Kondo et al. [1980] and makes use of electromagnetic particle velocity gauges [Dremin and Shvedov, 1974]. Gauges are oriented in a steady, uniform magnetic field such that the active element of the gauge, the magnetic field lines, and the direction of motion are all mutually perpendicular and the gauge leads are parallel to the magnetic field. An electromotive force is induced along the length of the gauge element, proportional to the velocity of the gauge. For a gauge with effective length L in a magnetic field B, the potential measured across the gauge leads is

$$V(t) = B L u_p(t)$$

(1)

where $u_p$ is the particle velocity of the material in which the gauge is embedded. These signals are recorded by an array of cathode ray oscilloscopes.

Gauges were photoetched from 10.0-μm-thick copper foil with a 12.5-μm-thick polyamide (Kapton) film backing. The active elements of the gauges were 0.9-1.0 cm long. The precise effective gauge length (used in equation (1)) is dependent on the geometry of the gauge due to electromagnetic edge effects from the leads and the finite gauge width. It is about equal to the center-to-center lead distance of 0.95 cm.

Rock samples were constructed by sawing rectangular slabs about 3.2 cm by 4.5 cm and grinding them to a uniform thickness of about 1.5 mm. Archimedean densities were determined for each slab individually. Four slabs were bound together with epoxy, with three gauges at the interfaces and one at the free surface. The mean thickness of the glue at each interface was ~9 μm. All four sets of gauge leads extended from the sample in the same direction along the long axis of the rock slabs. The gauge leads were connected to oscilloscopes via coaxial cable.

In Figure 1 the particle velocity experimental assembly is shown schematically, with a rock target mounted on the...
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Figure 1. Schematic drawing of particle velocity experiment, with major components indicated: A, polycarbonate projectile; B, 40-mm bore gun barrel; C, timing laser; D, photodetector; E, high-power switch (ignitron); F, capacitor bank; G, Helmholtz coils; H, rock target; I, self-shorting trigger pins; J, fiducial pulse generator; K, copper foil particle velocity gauge elements.

center of the axis of a set of Helmholtz coils, which supply the magnetic field. To ensure mutual perpendicularity between the gauge, field, and particle velocity, the target is aligned with the gun barrel by using a laser beam.

The Helmholtz coils have a radius and separation of 14.1 cm and are wound with four turns of gauge 10 copper wire. The magnetic field at the center point of the coil axis is constant to third order and equal to

$$B = \frac{8\mu_0 I N}{(3)^{3/2}r^3}$$

where $\mu_0$ is the permeability constant, $I$ is the current, $N$ is the number of turns, and $r$ is the radius and distance between coils. The field is typically about 1.8 kG. The current is supplied by a bank of five 15-μF capacitors charged to 5.0 kV and reaches its peak value of about 6.7 kA in about 60 μs. Because the time scale of the experiment, which is determined by the shock and rarefaction transit times through the sample (≈ 2 μs), is short compared to the period of the capacitor-coil circuit (≈ 250 μs), the timing can be controlled so that impact occurs at the peak current, when the field is effectively constant.

Targets were shock-loaded by impact of flat-faced polycarbonate (Lexan) projectiles fired from a 40-mm bore propellant gun at velocities from 1.4 to 2.4 km/s. Projectile velocities were determined using the time intervals between the projectile obscuring a series of laser beams [Ahrens et al., 1971]. The geometry and time history of a typical experiment is illustrated by means of an x-t diagram in Figure 2. The stationary rock target lies to the right of the origin, with four particle velocity gauges initially at intervals of 1.5 mm. The polycarbonate projectile approaches

TABLE 1. San Gabriel Anorthosite Mineralogy

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Volume %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Plagioclase *</td>
<td>91</td>
</tr>
<tr>
<td>White Mica</td>
<td>6</td>
</tr>
<tr>
<td>Epidote</td>
<td>3</td>
</tr>
<tr>
<td>Opaques</td>
<td>trace</td>
</tr>
<tr>
<td>Quartz</td>
<td>trace</td>
</tr>
<tr>
<td>Apatite</td>
<td>trace</td>
</tr>
</tbody>
</table>

Analysis by S. Rigden.

* $An_{40}$

Figure 2. Particle velocity experiment represented by x-t diagram. Projectile approaches stationary target from left and impacts at $t=0$. 
from the left and strikes the x=0 surface of the target at
time t=0, driving a shock wave to the right into the rock
and to the left into the projectile. Each gauge is stationary
until overtaken from the left by the shock wave, at which
time it begins moving with the particle velocity associated
with the Hugoniot state. The shock wave reflects from the
free surface as a rarefaction wave, and each gauge again
accelerates to the right as this wave passes through it from
the right.

San Gabriel anorthosite samples were collected in the
San Gabriel Mountains near Pasadena, California. This
rock type is highly variable in composition and texture and
has been studied in detail by Carter [1982]. The particular
specimen used in these experiments had randomly oriented
plagioclase crystals with a mean grain size of 1-2 mm.
Significant alteration was observed at grain boundaries and
the mineralogy, given in Table 1, was determined with a
petrographic microscope. The plagioclase was found to
have a mean composition of An_{40} as determined from
extinction angle measurements. An electron microprobe
analysis indicated a composition of An_{35}. The San Marcos
gabbro was obtained near Escondido, California. This
intrusion has been studied petrologically by Miller [1937].

Samples from the same specimen used in the present study
were used in impact and in spall strength experiments by
Lange et al. [1984]. The mineralogy is tabulated in Table
2.

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Volume %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Plagioclase</td>
<td>67.9</td>
</tr>
<tr>
<td>Amphibole</td>
<td>22.5</td>
</tr>
<tr>
<td>Clinopyroxene</td>
<td>1.5</td>
</tr>
<tr>
<td>Orthopyroxene</td>
<td>1.1</td>
</tr>
<tr>
<td>Quartz</td>
<td>1.4</td>
</tr>
<tr>
<td>Biotite</td>
<td>0.9</td>
</tr>
<tr>
<td>Opaques</td>
<td>4.3</td>
</tr>
<tr>
<td>Alkali feldspar</td>
<td>trace</td>
</tr>
<tr>
<td>Calcite</td>
<td>trace</td>
</tr>
<tr>
<td>Chlorite</td>
<td>trace</td>
</tr>
<tr>
<td>Apatite</td>
<td>trace</td>
</tr>
</tbody>
</table>

Analysis by R. Hill.

Figure 3. Oscillograms of particle velocity-time profiles in (a) anorthosite and (b) gabbro. Gabbro signals are significantly noisier, presumably due to presence of quartz grains.
corresponds to shock wave arrival, and the second increase corresponds to the acceleration from free surface rarefaction. The gabbro records were found to be significantly noisier than the anorthosite records in all cases. The deviations in particle velocity appear to be too large to be caused by differential grain motion. Instead, they are probably electrical effects which result from the presence of piezoelectric quartz grains in the gabbro, whereas the anorthosite was relatively free of quartz.

Digitized oscilloscope records for anorthosite shocked to 10 GPa are shown in Figure 4. Synchronization of the four signals in time was achieved by means of a fiducial pulse received simultaneously at each oscilloscope. Shock transit times were taken from the interval between arrival at different gauges and were used to determine shock velocity. The known projectile velocity and polycarbonate Hugoniot states for both rocks are given in Tables 3 and 4.

It is evident from Figure 3 that the rise time of the shock records is as great as 200 ns. The rise time of a gauge itself is limited by its deviation from planarity to the planar shock (tilt) and by the time it takes the gauge to reverberate up to the particle velocity of the surrounding medium. The thickness of the gauge-backing-glue layer is about 30 μm. Several shock transit times through this layer should accelerate it in less than 50 ns. Other experiments carried out on the same gun demonstrate that shock pulses are no greater than ~ 50 ns across a 1-cm element [Watt and Ahrens, 1983].

The observed release waves are nonsteady simple waves and can be inverted to stress density release paths by numerically integrating the equations for conservation of mass and linear momentum [Cowperthwaite and Williams, 1971]:

\[
\begin{align*}
\frac{\partial \rho}{\partial x} \bigg|_b &= \frac{\rho}{\rho_0} C(u_p) \\
\frac{\partial \sigma}{\partial x} \bigg|_b &= \frac{\rho}{\rho_0} C(u_p) 
\end{align*}
\]

where \( \rho \) is the density, \( \rho_0 \) is the initial density, \( \sigma \) is the stress, \( u_p \) is the particle velocity, and \( h \) is the Lagrangian space coordinate along the direction of wave propagation. The Lagrangian sound velocity is determined by the finite difference approximation

\[ C(u_p) \approx \frac{\Delta h}{\Delta t} \]  

where \( \Delta h \) is the initial distance between gauges and \( \Delta t \) is the transit time for a disturbance with particle velocity \( u_p \).

In the case of anorthosite the release paths are smooth and single-valued (Figure 3a). The oscilloscope records were digitized and integrated directly. The resulting release paths are consistent and are plotted in Figure 5. Owing to the noisiness of the gabbro records (Figure 3b), it was necessary to approximate the rarefaction waves by smooth curves before integrating. Complete release paths were not obtained, and the partial release paths for San Marcos gabbro are less consistent than those for anorthosite (Figure 6). Eulerian sound speeds, equal to \( (\rho_0/\rho) C(u_p) \), were also calculated for the release paths. These are plotted for anorthosite as a function of stress in Figure 7.
relative to the data and was subtracted out by a linear extrapolation of the observed baseline drift prior to the shock arrival.

Discussion

Sound velocities of anorthosite from this study can be compared to those measured ultrasonically for anorthosite rocks of similar composition [Birch, 1960, 1961; Simmons, 1964; Anderson and Lieberman, 1968; Lieberman and Ringwood, 1976]. At $T = 25^\circ C$ and $P > 0.4$ to $1$ GPa, longitudinal velocities have been measured in anorthosite with composition between An$_{40}$ and An$_{65}$ in the range $6.76 < V_p < 7.47$ km/s and shear velocities in the range $3.87 < V_s < 4.09$ km/s. Bulk sound velocities $V_s^2 = V_p^2 - \frac{4}{3} V_s^2$ (6) are therefore in the range $5.04 < V_s < 5.99$ km/s.

We have adapted values of 6.85, 3.60, and 5.44 ± 0.07 km/s for values of $V_{p0}$, $V_{s0}$, and $V_{so}$, respectively, the zero pressure sound speeds of An$_{40}$ using the collected data of Lieberman and Ringwood [1976]. In order to compare our results to those expected for the longitudinal wave speeds at high pressure, we have used the Eulerian finite strain formulation of Sammis et al. [1970] (see also Burdick and Anderson [1975]) to predict values of $V_p$ along an isentrope:

$$V_p(P)^2 = V_p^2(1-2\epsilon)(1-2\epsilon(3K_0 D_p - 1))$$ (7)

$$P = -3K_0(1-2\epsilon)^{5/2}(1+2\epsilon\xi)$$ (8)

where $K_0 = 79$ GPa is the zero pressure isentropic bulk modulus. Here the Eulerian volumetric strain is

$$\epsilon = [1-(p/p_0)^{2/3}]$$ (9)

the finite strain parameter $\xi$ is

$$\xi = \frac{3}{4}[4-(dK_0/dP)_0]$$ (10) and

$$D_p = d\ln V_p/dP$$ (11)

No well-constrained values of $\xi$ and $D_2$ for any feldspar have yet been measured. Using data for other minerals

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**TABLE 4. San Marcos Gabbro Shock Wave Data**

<table>
<thead>
<tr>
<th>Shot</th>
<th>Projectile Velocity, km/s</th>
<th>Initial Density, Mg/m$^3$</th>
<th>Shock Velocity, km/s</th>
<th>Particle Sound Velocity, km/s</th>
<th>Pressure, GPa</th>
<th>Density, Mg/m$^3$</th>
<th>Sound Velocity, km/s</th>
</tr>
</thead>
<tbody>
<tr>
<td>40-569</td>
<td>1.394</td>
<td>2.907</td>
<td>6.487</td>
<td>0.286</td>
<td>5.39</td>
<td>3.041</td>
<td>6.5</td>
</tr>
<tr>
<td>40-573</td>
<td>+0.009</td>
<td>+0.008</td>
<td>+0.035</td>
<td>+0.011</td>
<td>+0.20</td>
<td>+0.004</td>
<td>+1</td>
</tr>
<tr>
<td>40-555</td>
<td>1.968</td>
<td>2.886</td>
<td>6.11</td>
<td>0.474</td>
<td>8.36</td>
<td>3.129</td>
<td>8.5</td>
</tr>
<tr>
<td>40-556</td>
<td>+0.006</td>
<td>+0.001</td>
<td>+0.30</td>
<td>+0.016</td>
<td>+0.14</td>
<td>+0.022</td>
<td>+0.4</td>
</tr>
<tr>
<td>40-574</td>
<td>+0.050</td>
<td>+0.007</td>
<td>+0.19</td>
<td>+0.020</td>
<td>+0.34</td>
<td>+0.015</td>
<td>---</td>
</tr>
<tr>
<td>40-556</td>
<td>2.242</td>
<td>2.929</td>
<td>6.898</td>
<td>0.515</td>
<td>10.41</td>
<td>3.165</td>
<td>7.0</td>
</tr>
<tr>
<td>40-574</td>
<td>-0.039</td>
<td>-0.039</td>
<td>-0.042</td>
<td>-0.048</td>
<td>-0.84</td>
<td>-0.020</td>
<td>-0.5</td>
</tr>
<tr>
<td>40-574</td>
<td>+0.025</td>
<td>+0.049</td>
<td>+0.016</td>
<td>+0.052</td>
<td>+0.77</td>
<td>+0.026</td>
<td>+0.5</td>
</tr>
</tbody>
</table>

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Figure 5. Hugoniot states and release paths of San Gabriel anorthosite. Included are two Hugoniot states of anorthite glass [Boslough et al., 1985b], with respective partial release states. Hydrostat assumes $K_{so} = 79$ GPa.

Figure 6. Hugoniot states and partial release paths of San Marcos gabbro. Noisy particle velocity records precluded determination of release to zero stress. Two interpretations are given for the release data for shot 573.
Figures 5 and 7 demonstrate that in all three shock wave states, would be expected to achieve Hugoniot states having pressure. This implies a subsequent loss of shear strength, indicating elasto-plastic or elasto-hydrodynamic unloading the bulk sound speed \( v \) as the rock was released to zero pressure. This implies a subsequent loss of shear strength, indicating elasto-plastic behavior upon release from shock compression [Grady et al., 1974] for higher pressures that the release path of shocked polycrystalline quartz is controlled by the quartz \( \rightarrow \) stishovite transition. A clear case cannot be made for the retention or loss of shear strength or existence of phase transitions in gabbro due to the lower quality of the data. However, in most of the experiments the measured sound velocities are consistently lower for the gabbro than for the anorthosite. The release data are not of good enough quality to resolve whether the release paths are above the Hugoniot.

Conclusions

Since the unloading wave speed approaches the bulk sound speed, we conclude that San Gabriel anorthosite loses shear strength upon release from shock compression to pressures greater than 6 GPa. This material's release behavior can be partially attributed to the phase transformation of albite to jadeite and quartz or coesite. The amount of material transformed is likely to be controlled by kinetics and appears approximately to be an increasing function of peak stress. This may be due to the localization of shock heating into shear bands, in which the temperature is high enough for the phase transition to occur. The mechanical result of the shear banding is hypothesized to be fluidlike rheological behavior upon release [Grady et al., 1975; Grady, 1980]. The hydrodynamic behavior of shocked anorthosite will result in less rapid attenuation of a decaying shock wave than obtained in the usual elastic-plastic rheological model.

In calculations by O'Keefe and Ahrens [1977] of energy partitioning of a hypervelocity impact onto the anorthosite surface of a planet, the release behavior of the rock was assumed to be elastic-plastic [Fowles, 1960; Davison and Graham, 1979]. This assumption would tend to overestimate the attenuation of the shock wave due to catch-up of rarefaction waves. Thus a larger quantity of surface material is shocked above a given pressure, and estimates by O'Keefe and Ahrens [1977] of the fraction of meteoroid kinetic energy deposited in the planetary surface material will be too low.

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\[
\text{NaAlSi}_{3}\text{O}_{8} \rightarrow \text{NaAlSi}_{2}\text{O}_{6} + \text{SiO}_{2}
\]
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