SHOCK WAVES PROPERTIES OF ANOORTHOSITE AND GABRO

Mark B. Bostough and Thomas J. Ahrens

Abstract. Shock wave experiments have been conducted on San Gabriel anorthosite and San Marco gabbro to peak stresses between 5 and 11 GPa using a 40-mm-bore propellant gun. Particle velocity 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 adiabat 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 8 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 velocities. Stress-release release paths in the anorthosite indicate possible transformation of albite to jadeite + quartz + other, with amounts of albite transformed ranging from 0.2% to 8% as much as 0.19 mass fraction in the 9.7 GPa shock stress range. Electri-

cal interference effects precluded the determination of accurate release paths for San Marco gabro. Because of the apparent loss of shear strength during unloading from the shocked state, the fluidlike behavior of anorthosite which is indicated implies that calculations of energy partitioning due to impact onto 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 gabbrod, are important components of the lunar and terrestri-

al crust. It is necessary to understand the behavior of such rocks under high dynamic stresses in order to model cratering processes which result from hypervelocity impacts and to understand their role in the formation of such rocks which have been subjected to shock loading on planetary surfaces (such as the Moon). Stock waves studies of these and similar materials have been conducted in the past (Ahrens et al., 1989; McQueen et al., 1981; Jenkis; and Ahrens, 1989; Bostough et al., 1985a), but in those studies the data are limited to the Hugoniot state at 10 mm/s shock velocity and in some cases shock states. 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; Swanson, 1974). Particle velocity experiments supply detailed release paths, which provide better constraint for mechanical properties and polymorphism than is available with Hugoniot experi-

ments alone. 

Petersen et al. (1970) used particle velocity gauges to determine release paths of plagioclase, anorthosite, and sodalite shocked to stresses up to 5 GPa. They attri

buted high rarefaction velocities and steep release paths in the anorthosite phases to unstable compression. Grady et al. (1974) carried out experiments on polycrystalline quartz (nominally) to 80 GPa, and an examination of partic-

le velocity and mass average stress gauges to determine release adiabats. They concluded that a partial quartz-mica transformation takes place above 15 GPa, with the quantity of material transformed as increasing function of the quartz-mica transition temperature. Similar experiments were conducted on polycrystalline quartz and polycrystalline feldspar by Grady et al. (1975) and Grady and Wurz (1976), who used manometer stress gauges to determine Hugoniot sound velocities and found that these rocks lose shear strength when shocked to pressures above 20 GPa. Lame and Anderson (1979) used particle velocity gauges to study stishovite and tuff at lower stress levels (4 GPa) and attributed the observed time-dependent behavior to the slowing of waves in these rocks.

In this paper we present new Hugoniot data on San Gabriel anorthosite and San Marco 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), Lame and Anderson (1979), and Koseo et al. (1986) and makes use of electromagnetic particle velocity gauges (Dremin and Shevov, 1974). Gauges are oriented in a steady, uniform magnetic field such that the active element of the gauge is the magnetic field lines and the direction of motion is all nearly per-

pendicular and the gauge plane is parallel to the magnetic field. An electromagnetic signal of the gauge element, proportional to the velocity of the gauge. For a gauge of effective length L is in a magnetic field B, the potential measured across the gauge leads is

\[ V = \mu B L |v| \]

where \( v \) is the particle velocity of the material in which the gauge is embedded. These signals are recorded by an array of coaxial thin-foil oscilloscopes. Gauges were phosphorized from 10.0-mm-thick copper foil with a 12-Yttria-stabilized zirconia (Kapton) film backing. The active elements of the gauges were 0.6-1.0 cm long. The particle effective gauge length (used in equation [1]) is dependent on the geometry of the gauge due to electromagnetic edge effects from the gauge plane. \( L \) is about equal to the center-to-center lead distance of 0.6 cm.

Rock samples were constructed by sawing rectangular slabs about 3.2 cm by 4.5 cm and grinding them to a uni-

form thickness of about 1.5 mm. Accuracy dimensions were required for each slab individually. Four sides were ground together with epoxy, with three gauges at the interface and one at the free surface. The mean thickness of the glass at each interface was 0.8 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 cables.

In Figure 1 the particle velocity experimental assembly is shown schematically, with a rock target mounted on the
center of the axis of a set of Helmholtz coils, which supply the magnetic field. To ensure minimal 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 141 cm and are wound with four turns of gauge 10 copper wire. The magnetic field at the outer point of the coil axis is constant to third order and equal to

$$B = \frac{E_{in}}{2I_{141}}$$

or

$$B = \frac{E_{in}}{2I_{141}}$$

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 10-mC capacitors charged to 5.5 kV and reaches its peak value of about 5.7 kA in about 60 \( \mu \)s. Because the time scale of the experiment, which is determined by the shock wave fraction transit times through the sample (2 \( \mu \)s), is shorter compared to the period of the capacitor-coil circuit (15 \( \mu \)s), 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 charring a series of laser beams (Abram et al., 1971). The geometry and time history of a typical experiment is illustrated by means of an n-t diagram in Figure 2. The stationary rock target lies to the right of the origin, well, four particle velocity gauges initially at intervals of 1.5 mm. The polycarbonate projectile approaches

![Schematic drawing of particle velocity experiment, with major components indicated. A, polycarbonate projectile; B, double-bore gun barrel; C, timing laser; D, photodetector; E, high-power switch (ignition); F, capacitor bank; G, Helmholtz coil; H, rock target; I, self-igniting trigger pipe; J, induction pulse generator; K, copper foil particle velocity gauge elements.](image-url)

Figure 1. Schematic drawing of particle velocity experiment, with major components indicated. A, polycarbonate projectile; B, double-bore gun barrel; C, timing laser; D, photodetector; E, high-power switch (ignition); F, capacitor bank; G, Helmholtz coil; H, rock target; I, self-igniting trigger pipe; J, induction pulse generator; K, copper foil particle velocity gauge elements.

![Particle velocity experiment represented by n-t diagram. Projectile approaches stationary target from left and impacts at 90°.](image-url)

Figure 2. Particle velocity experiment represented by n-t diagram. Projectile approaches stationary target from left and impacts at 90°.

<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>Opaque</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.

**TABLE 1. San Gabriel Anorthosite: Mineralogy**
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's stationary until overtaken from the left by the shock wave, at which time it begins moving with the particle velocity associated with the Lagrangian state. The shock wave reflects from the free surface as a reflection 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 [1965]. 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 petrography, given in Table 1, was determined with a petrographic microscope. The plagioclase was found to have a mean composition of Ab50, as determined from extinction angle measurements. An electron microprobe analysis indicated a composition of Ab35. The San Marcos gabbro was collected near Escondido, California. This intrusion has been studied petrologically by Miller [1977].

TABLE 2. San Marcos Gabbro Mineralogy

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Volume %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Plagioclase</td>
<td>57.9</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>Opaque</td>
<td>4.3</td>
</tr>
<tr>
<td>Alaskan olivine</td>
<td>trace</td>
</tr>
<tr>
<td>Calcareous</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.

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

Figure 2. Oscillograms of particle velocity-time profiles in (a) anorthosite and (b) gabbro. Gabbro signals are significantly noisier, presumably due to presence of quartz grains.
Typical particle velocity records for shocked anorthosites and gabбро are shown in Figure 2. The sudden increase corresponds to shock wave arrival, and the second increase corresponds to the acceleration from free surface reflection. The gabбро 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 gabбро, whereas the anorthosite was relatively free of quartz.

Digested oscillographic 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 polycrystalline Hugoniot (March, 1980) were used with an impulsive match solution [McQueen et al., 1970] to determine the Hugoniot states achieved in the rocks. In principal, the Hugoniot state could be determined by measuring the particle velocity of the shocked state directly. This would require a series of gauge calibration shots and precise measurement of the precise moment of the reduced magnetic field. By our method each shot was self-calibrating: the particle velocity of the shock state was determined by impedance matching from which we obtained a calibration factor to get the subsequent particle velocity history. Fabrication error velocities were determined from the transit time of the free surface refractioin front and the Hugoniot density. Hugoniot states and sound velocities for both rocks are given in Tables 3 and 4.

It is evident from Figure 3 that the rise time of the

**TABLE 3**

<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 Velocity, Sound Velocity, GPa, Density, Mg/m$^3$</th>
<th>Sound Velocity, km/s</th>
</tr>
</thead>
<tbody>
<tr>
<td>40-572</td>
<td>1.57</td>
<td>2.565</td>
<td>5.568</td>
<td>0.364</td>
<td>3.91</td>
</tr>
<tr>
<td>40-570</td>
<td>1.581</td>
<td>2.133</td>
<td>5.724</td>
<td>0.829</td>
<td>5.833</td>
</tr>
<tr>
<td>40-571</td>
<td>2.381</td>
<td>2.853</td>
<td>3.985</td>
<td>0.444</td>
<td>10.83</td>
</tr>
</tbody>
</table>

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

$$\frac{\partial \rho}{\partial t} = \rho c \frac{\partial (\rho \bar{v})}{\partial z}$$

$$\frac{\partial \bar{v}}{\partial t} = \frac{\partial (\rho \bar{v})}{\partial z}$$

where $\rho$ is the density, $\bar{v}$ is the initial density, $\sigma$ is the stress, $\bar{v}_0$ is the particle velocity, and $C_{L}$ is the Lagrangian space coordinate along the direction of wave propagation. The Lagrangian sound velocity is determined by the finite difference approximation

$$C_{L} \cong \frac{d}{d\Delta t}$$

where $\Delta t$ is the initial distance between gauges and $\Delta t$ is the transit time for a disturbance with particle velocity $\bar{v}$. In the case of an accurate shock arrival time, the release paths can be integrated to obtain the stress-strain path. The stress-strain paths for shocked rocks are presented in Figure 3. Owing to the nature of the gauge records (Figure 3b), it was necessary to approximate the release wave by smooth curves before integrating. Complete release paths were not obtained, and the partial release paths for San Marcos gabбро are less consistent than those for anorthosite (Figure 6). Equivalent sound speeds, equal to $\rho c_{L}$, were also calculated for the release paths. These are plotted for anorthosite as a function of stress in Figure 7.

In practice, it was necessary to carry out some adjustment of the data before integrating. A small amount of baseline drift existed in the data due to the coercive magnetic field flux through the gauge loop. The drift was slow
## Table 4. San Marcos Gabro Shock Wave Data

<table>
<thead>
<tr>
<th>Shock Velocity, km/s</th>
<th>Initial Density, Mg/m³</th>
<th>Shock Velocity, km/s</th>
<th>Particle Velocity, GPa</th>
<th>Pressure, GPa</th>
<th>Density, Mg/m³</th>
<th>Sound Velocity, km/s</th>
</tr>
</thead>
<tbody>
<tr>
<td>49-595</td>
<td>1.984</td>
<td>7.907</td>
<td>6.497</td>
<td>0.286</td>
<td>0.060</td>
<td>0.095</td>
</tr>
<tr>
<td>49-573</td>
<td>1.984</td>
<td>7.907</td>
<td>6.497</td>
<td>0.286</td>
<td>0.060</td>
<td>0.095</td>
</tr>
<tr>
<td>49-555</td>
<td>2.187</td>
<td>2.892</td>
<td>6.65</td>
<td>0.515</td>
<td>0.100</td>
<td>0.105</td>
</tr>
<tr>
<td>49-595</td>
<td>2.342</td>
<td>2.920</td>
<td>8.898</td>
<td>0.415</td>
<td>0.100</td>
<td>0.105</td>
</tr>
<tr>
<td>49-574</td>
<td>2.418</td>
<td>2.874</td>
<td>4.904</td>
<td>0.909</td>
<td>0.100</td>
<td>0.105</td>
</tr>
</tbody>
</table>

Relative to the data and was subtracted out by a linear extrapolation of the observed baseline shifts 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 (Nee, 1960, 1961; Simmons, 1964; Anderson and Latham, 1968; Liefmann and Ringwood, 1979). At T = 27°C and P > 0.6 to 1 GPa, longitudinal velocities have been measured in anorthosite with compositions between An 60 and An 80 in the range 3.76 < v_l < 4.74 km/s and shear velocities in the range 2.87 < v_s < 3.06 km/s. Bulk sound velocities

\[
v^2 = v_l^2 - \frac{4}{3} \frac{v_s^2}{v_l^2}
\]

are therefore in the range 5.04 ≤ v_l ≤ 5.99 km/s.

We have adopted values of 6.35, 3.60, and 5.44 ± 0.07 km/s for values of v_l, v_s, and \( \sigma \), respectively, the zero pressure sound speeds of An 80 using the collected data of Liefmann and Ringwood (1979). 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 Stockmarr et al. (1970) (see also Burdick and Andersen [1975]) to predict values of \( v_l \) along an isosteric:

\[
v_l(P) = v_l(2-\sigma) [1-2(1-K_m - \sigma)] \quad (7)
\]

where \( K_m \) = 79 GPa is the zero pressure isentropic bulk modulus. The Eulerian volumetric strain is

\[
\epsilon = \frac{1}{3} \sigma \frac{1}{\kappa} \frac{v_l^2}{2 v_s^2}
\]

the finite strain parameter \( \xi \) is

\[
\xi = \frac{\epsilon}{\epsilon + D_3}
\]

and

\[ D_3 = \frac{1}{v_l^2 v_s^2} \]

No well-crystallized value of \( \xi \) and \( D_3 \) for any feldspar have yet been measured. Using data for other minerals

![Figure 5. Hugoniot states and release path of San Gabriel anorthosite. Included are two Hugoniot states of anorthosite glass (Bosworth et al., 1985), with respective partial release states. Hydrostatic stresses \( \sigma_0 = 79 \) GPa.](image1)

![Figure 6. Hugoniot states and partial release path of San Marcos gabro. Noisy particle velocity records precluded determination of release to zero stress. Two interpretations are given for the release data for shock 57.)
Figure 7. Eulerian sound speeds along release path of shocked NaAlSiO₃ anhydroite. Hatchet region indicates range of possible zero pressure bulk sound velocities.

(Appertet al., 1964) as a guide, we have adopted values in the range

\[ 1.3 \leq \varepsilon \leq 3.75 \]  

(12) and

\[ 0.012 \leq D_0 \leq 0.016 \text{ GPa}^{-1} \]  

(13)
to estimate the variation of \( \varepsilon \) with pressure (Figure 7). Figures 5 and 7 demonstrate that in all stress shock wave experiments carried out on anhydroite, the initial sound velocity immediately behind the shock front was close to the expected longitudinal velocity, indicating that the material has retained strength. In all cases the sound speed decreased markedly upon unloading and approached the bulk sound speed \( v_0 \) at the rock was released in zero pressure. This implies a subsequent loss of shear strength, indicating elastic-plastic or elasto-hydodynamic unloading behavior e.g., Lee, (201). Using equation (6) and \( K_0 = 79 \text{ GPa} \), we constructed a hydrostatic (isentropic) in the stress-density plane (Figure 9). An ideal elastic-plastic solid, retaining strength in the shock state, would be expected to achieve Hugoniot states having a perfect shock stress, at a given density, less than those of the hydrostatic by the amount \( 4/3 \varepsilon \), where \( \varepsilon \) is the shear strength of the material (Corson and Graham, 1979). Above the Hugoniot elastic limit, the hydrostatic defines the lower limit of the Hugoniot for material which sustains elasto-hydodynamic behavior, losing strength on compression.

There are two possible interpretations of the observation that the Hugoniot lies well below the calculated hydrostatic. First, it is possible that the measured bulk modules (79 GPa) is too large. A lower effective bulk modulus, \( \sim 65 \text{ GPa} \), would be required to bring the hydrostatics into coincidence with the observed Hugoniot. Moreover, if the anhydroite is considered an elastic-plastic solid, with nonzero strength, the required bulk modulus would need to be significantly less than 85 GPa. A much reduced bulk modulus \( \sim 65 \text{ GPa} \) is discordant with the extensive ultrasonic measurements which have been carried out on this type of phosphates (Liebermann and Kingwood, 1976). For this reason, we prefer a second hypothesis involving one or more phase transitions occurring below, or within, the range of our present data.

A phase transformation such as

\[
\text{NaAlSiO}_3 \text{ (albite)} \rightarrow \text{NaAlSiO}_3 \text{ (jadeite)} \]

\[ + \text{SiO}_2 \text{ (quartz or comite)} \]  

(14) could begin to occur at the stress levels of the present experiments, especially if it took place within a minor population of shear band tested zones (Grady, 1980). We suggest the occurrence of reaction (14) in light of the identification of shock-induced formation of jadeite from shocked olivines in material from the Minoan-Greek (James, 1988). The assemblage [anhydroite + jadeite + quartz] has a density of 3.01 Mg/m³, and [anhydroite + partial jadeite + quartz] has a density of 2.96 Mg/m³. At 20.3 K, NaAlSiO₃ + SiO₂ are the stable phase above 0.6 GPa (Kraatsch, 1964). The pressures calculated are well above the transformation pressure, so the only barrier to the phase transition is the shock-induced pressure density of the related anhydroite can be used to estimate the maximum amount of material transformed. For example, the final densities are in the range 2.69-2.73 Mg/m³, which requires a maximum transformation of from 10-15 % of the albite to high-pressure phase if the release path is along the hydrostatic of the high-pressure phases. The possibility of pure space crunching and irreversible compaction would bring this estimate down somewhat. The data clearly show that the fraction of albite transformed to high-pressure phases is an increasing function of peak stress. This result is analogous to the conclusion of Grady et al. (1974) for higher pressures that the release path of shocked pyroxenite quartzite is controlled by the quartz - stishovite transition. A clear case cannot be made for the retention of or loss of shear strength or existence of phase transitions in jadeite due to the lower quality of the data. However, in most of the experiments the post-shock sound velocities are consistently lower for the albite than for the anhydroite. The release data are not of good enough quality to resolve whether the release path are above the Hugoniot.

Conclusions

Since the unloading wave speed approaches the bulk sound speed, none of the shock transformation phases lose their shear strength upon release from shock compression to pressures greater than 8 GPa. The materials' release behavior can be attributed to the phase transformation of albite to jadeite and quartz or comite. The amount of material transformation appears to be controlled by kinetics and appears approximately to be an increasing fraction of shock compression. The release path of shock heating into shear bands, in which the temperature is high enough for the phase transition to occur. The mechanism of shock heating in the unloading wave is to be fluidized rheological behavior upon release (Grady et al., 1979). The shock healing behavior of shocked anhydroite will result in less rapid intensification 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 anhydroite surface of a planet, the release behavior of the rock was assumed to be elastic-plastic (Powell, 1982; Ziegler and Grady, 1979). This assumption would need to overestimate the attenuation of the shock wave due to early mixing of rarefaction waves. Thus a larger volume of surface material is shown to give a good pressure, and estimate by O'Keefe and Ahrens (1977) of the fraction of mass of kinetic energy deposited in the planetary surface material will be too low.

Acknowledgments

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