Shock compression of a recrystallized anorthositic rock from Apollo 15

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Shock compression of a recrystallized anorthositic rock from Apollo 15  

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Abstract—High-pressure experiments on 15,281, a recrystallized and brecciated galactic anorthosite, yield a value of the Hugoniot elastic limit (HEL) ranging from 30 to 40 kilobars at the final shock pressure in excess of 20 to 30 kilobars above the HEL and to 40 kilobars at the plasticity density limit, a density slightly above 3.7 g/cm³, at a shock pressure of 50 to 60 kilobars. The Hugoniot state indicates that a volume of one to two percent of the high-pressure phase change is occurring in the plagioclase and pyroxene. From Hugoniot data, for both the single-crystal minerals and the Federovite shatter, we infer that the shock-induced high-pressure phases in 15,281 probably consist of 3.35 g/cm³ density, high-pressure spinel for plagioclase (An₄₁), and a 4.35 g/cm³ pyroxene-spinel mixture (En₃₅). The shocked material, except for the pyroxene component, is expected to completely melt upon release from 700 and 100 kbar. The onset of shock-induced vaporization will occur upon release from 1,500 kbar and would require an impact of an ice meteoroid traveling at a velocity of 4000 kmps.

Introduction

This is the first report of shock wave data taken on a rock sample returned from the moon. The major motivation for determining the shock wave equation of state of such rocks stems from the widespread occurrence of impact effects on both large and small scales on the lunar surface. Although the large scale cratering features on the surface of the moon were recognized well before the U.S. and Soviet programs of sample collection, the wide variety of effects resulting from the passage of an intense pressure pulse associated with impact have only been recognized upon analysis of returned rock and soil samples over the last four years. Among the shock effects observed in these samples are melting (e.g., Chao et al., 1972; von Engelscheidt et al., 1971; King et al., 1972; Chao et al., 1976), cataclastic crushing (von Engelscheidt et al., 1972; SPEET, 1973), intense crushing, twinning and cold solid state reactions (von Engelscheidt, 1970, 1971, 1972). Also, geochemical evidence points to shock vaporization as a likely explanation for the apparent mobility of Pb, K, Rb, Th, and U in the regolith (Silver, 1972; Don and Taburet, 1972). Undoubtedly, the most important shock effect on the moon is melting.
The shock measurements on lunar materials which are important in relating impact parameters (shock impedance and impact velocity of the meteoroid) to their resulting effects on the moon include: (a) the Hugoniot elastic limit, or, the maximum stress achievable in the rock under one-dimensional compression without internal rearrangement taking place at the shock front. Although some twinning and other planar deformation features result from shock stresses lower than this critical stress, large-scale shock-produced deformation and flow takes place only above the HEL, and (b) the high pressure equation of state along the defor- mational Hugoniot. Knowledge of the equation of state of the rock in terms of its component minerals permits correlation of the shock pressure with the onset and complexion of melting and vaporization of the constituent minerals, in the case of many of the lunar surface materials, the unique mineralogic and physical characteristics, as compared to terrestrial materials, e.g., high ferroilolite content of pyroxene combined with appreciable porosity, have made the high pressure Hugoniot difficult to predict theoretically.

In this paper we summarize our experimental results for lunar gabbronor- tholite and describe its Hugoniot data and that of a somewhat similar terrestrial gabbror in terms of equations of state of the major constituents, plagioclase and pyroxene.

**Samples and Experimental Method**

Rock 15418, collected from site 7 (LSEP, 1972) on the Hadley delta, appears to be a previously shocked and recrystallized rock of gabbronor- tholite composition. It is quite possible that this rock was in fact ejected via one or more impacts from an initial position in the lunar highlands. Petrographic analysis yields the following mineralogic composition: plagioclase 54 ± 3%, pyroxene plus a small fraction olivine, 20 ± 3%, opaque minerals (non-magnetic, totally) 0.3 ± 0.003, and schreibers 3.0 ± 0.3%. Micropaque analysis of the plagioclase (An), the ortho- and clino- pyroxenes, and olivine are given in Table 1.

Twelve individual samples, approximately 1 cm square and 3.5 mm thick, were machined from 15.418. Upon carefully measuring the weighing, we determined their average density to 1.92 ± 0.05 g/cm³. The density of each sample did not differ from the mean by more than 0.05 g/cm³. The above density includes the volume of vesicles.

After internal impurities using X-ray radiography, the samples were mounted with flat mirrors and an inclined mirror to measure shock and free surface velocities on 1.5 mm thick polyethylene tere- phthalate or aluminum alloy (6061) plates (Table 2). These samples were impacted with 2.5 mm thick pyrolytic graphite plates of the same material at speeds ranging from 0.8 to 2.2 km/sec. A typical record resulting from the streak photography is shown in Fig. 1. In all cases an elastic precursor, whose amplitude is taken to represent the Hugoniot elastic limit (HEL), is detected. The final shock state was determined using the impact method (Kice et al., 1958). The detailed analysis of the resulting streak camera film was similar to that used by Ahrens (1960).

**Experimental Results**

The velocity of the elastic precursor varies from 5.9 to 6.3 km/sec for the different samples. The variation undoubtedly reflects slight variations in crack porosity and crack geometry along the 3.5 mm shock propagation path. These val-
<table>
<thead>
<tr>
<th>Phlogopite (wt %)</th>
<th>Chloropiroene (wt %)</th>
<th>Orthopyroxene (wt %)</th>
<th>Olivine (wt %)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO2</td>
<td>44.7 ± 1.1</td>
<td>52.1 ± 0.9</td>
<td>51.7</td>
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<tr>
<td>Al2O3</td>
<td>35.5 ± 0.8</td>
<td>23.2 ± 0.2</td>
<td>21.1</td>
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<tr>
<td>CaO</td>
<td>79.6 ± 0.8</td>
<td>21.4 ± 0.7</td>
<td>14.2</td>
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<tr>
<td>Na2O</td>
<td>0.99 ± 0.02</td>
<td>1.2 ± 0.1</td>
<td>10.2</td>
</tr>
<tr>
<td>Fe2O3</td>
<td>0.02 ± 0.08</td>
<td>1.5 ± 0.6</td>
<td>2.0</td>
</tr>
<tr>
<td>MgO</td>
<td>0.06 ± 0.04</td>
<td>0.54 ± 0.05</td>
<td>1.1</td>
</tr>
<tr>
<td>K2O</td>
<td>0.05 ± 0.01</td>
<td>0.39 ± 0.01</td>
<td>0.8</td>
</tr>
<tr>
<td>Na2O</td>
<td>0.02 ± 0.007</td>
<td>0.38 ± 0.01</td>
<td>0.02</td>
</tr>
<tr>
<td>MgO</td>
<td>0.02 ± 0.007</td>
<td>0.54 ± 0.05</td>
<td>1.1</td>
</tr>
<tr>
<td>K2O</td>
<td>0.05 ± 0.01</td>
<td>0.39 ± 0.01</td>
<td>0.8</td>
</tr>
</tbody>
</table>

*A. Chabris, California Institute of Technology, Astrobiol. The amplitude of the elastic precursor (HEL) (varying from 42 to 71 kbar) is also higher than somewhat similar terrestrial materials, where values between 40 and 50 kbar are reported. Although we have not carried out detailed analysis, we attribute these differences to the lack of hydrous mineral alteration products and the higher CaO content of the lunar material. In addition, there appears to be a moderate strain-rate effect on the value of the HEL (Fig. 2). For the 3 shots in which a final amplitude below 50 kbar was achieved, the average HEL value is 48 kbar, while the seven values of the HEL measured for experiments in which final shock states were above 120 kbar had an average value of 66 kbar. Similar effects have been noted in studies of the shock compression of quartz (Wackerle, 1962; Ahrens and Davall, 1966) and feldspar (Ahrens and Liu, 1970).

Above the HEL to 350 kbar, the achieved deformational shock states agree closely with those reported earlier for Vacaville basalt (Ahrens and Gregson, 1964) and Centreville and Fredericks diabase (McQueen et al., 1967). (See Fig. 3.) In the next section we show that these states are also in good agreement with the theoretical Hugoniots predicted from the mineralogic constraint equations of state. Above ~150 kbar the data suggests that one or more phase changes take place in the lunar (and terrestrial) rocks. Since above a pressure of ~335 and ~140 kbar the data for pyroxene (Ahrens and Gaffney, 1971; McQueen et al., 1967) and plagioclase feldspars (Ahrens et al., 1960) demonstrate the occurrence of phase changes, this result is expected. The possible nature of these phase changes is discussed below.

A Theoretical Description of the Equation of State

In order to theoretically describe the equation of state of a rock such as 15.418, which is composed of essentially two minerals, plagioclase and pyroxene, a
Table 2: Hagygot Data. Mean sample 15.41b.

<table>
<thead>
<tr>
<th>Shot No.</th>
<th>Initial Density (g/cc)</th>
<th>Flyer Plate Velocity (km/sec)</th>
<th>Elastic Shock Velocity (km/sec)</th>
<th>Static Pore Velocity (km/sec)</th>
<th>Haygrot Elastic Limit (kbar)</th>
<th>Final Shock Pressure (kbar)</th>
<th>Final Density (g/cc)</th>
</tr>
</thead>
<tbody>
<tr>
<td>270</td>
<td>2.821 ± 0.005</td>
<td>1.600 ± 0.09</td>
<td>0.88 ± 0.09</td>
<td>0.40 ± 0.03</td>
<td>70 ± 1</td>
<td>204 ± 4</td>
<td>3.02 ± 0.07</td>
</tr>
<tr>
<td>276</td>
<td>2.834 ± 0.005</td>
<td>1.38 ± 0.02</td>
<td>0.62 ± 0.12</td>
<td></td>
<td></td>
<td>155 ± 4</td>
<td>3.69 ± 0.08</td>
</tr>
<tr>
<td>288</td>
<td>2.813 ± 0.005</td>
<td>2.166 ± 0.01</td>
<td>6.30 ± 0.06</td>
<td>0.67 ± 0.10</td>
<td>282 ± 6</td>
<td>4.25 ± 0.04</td>
<td></td>
</tr>
<tr>
<td>269</td>
<td>2.822 ± 0.005</td>
<td>1.992 ± 0.10</td>
<td>6.10 ± 0.05</td>
<td></td>
<td></td>
<td>261 ± 5</td>
<td>4.97 ± 0.04</td>
</tr>
<tr>
<td>270</td>
<td>2.816 ± 0.005</td>
<td>1.139 ± 0.04</td>
<td>5.94 ± 0.10</td>
<td>0.42 ± 0.03</td>
<td>88 ± 5</td>
<td>3.22 ± 0.06</td>
<td></td>
</tr>
<tr>
<td>290</td>
<td>2.823 ± 0.005</td>
<td>0.890 ± 0.04</td>
<td>6.04 ± 0.06</td>
<td>49 ± 5</td>
<td>65 ± 2</td>
<td>3.06 ± 0.05</td>
<td></td>
</tr>
<tr>
<td>281</td>
<td>2.812 ± 0.005</td>
<td>0.802 ± 0.02</td>
<td>6.18 ± 0.04</td>
<td>55 ± 2</td>
<td>63 ± 1</td>
<td>3.01 ± 0.01</td>
<td></td>
</tr>
<tr>
<td>277</td>
<td>2.821 ± 0.005</td>
<td>1.029 ± 0.04</td>
<td>5.99 ± 0.14</td>
<td>71 ± 10</td>
<td>129 ± 7</td>
<td>3.75 ± 0.06</td>
<td></td>
</tr>
<tr>
<td>241</td>
<td>2.823 ± 0.005</td>
<td>1.17 ± 0.04</td>
<td>6.24 ± 0.05</td>
<td>57 ± 6</td>
<td>148 ± 8</td>
<td>3.46 ± 0.08</td>
<td></td>
</tr>
<tr>
<td>206</td>
<td>2.806 ± 0.005</td>
<td>1.082 ± 0.02</td>
<td>6.14 ± 0.04</td>
<td>69 ± 11</td>
<td>145 ± 7</td>
<td>3.25 ± 0.04</td>
<td></td>
</tr>
</tbody>
</table>

Poly-crystalline W, 10.3 g/cc.

*Aluminum foil, 2024.

*Not measured.

Fig. 1: Shock wave photograph, after 206, Apollo 15.21b. (a) Still photograph of specimen viewed through streak camera. 500 image sweep across film. Mirror refection is destroyed by shock wave. (b) Resulting streak photograph showing streak cutoffs of similar reflectivities. Shock pressures at 60 ± 10 and 282 ± 6 kbar were determined for the elastic and plastic deformation shock wave from this record.
Shock compression of a matrixxiated anorthosite rock

Fig. 1. Hugoniot Elastic Limit (HEL) as a function of final shock pressure in 15.419.

Theoretical framework was chosen which uses an appropriate mixture of the equations of state of the constituents. The extension of the mixture theory of Truesdell (1957) by Kelly (1964) has been used here to synthesize the Hugoniots of Fredrick diabase and hornblende anorthosite and elsewhere to synthesize the Hugoniots of a number of other rocks (O'Keefe and Ahrens, 1973). Similar analyses have been carried out on water-saturated tuff (Cagnac and Kinosh, 1971) and composite materials (Tsou and Chiot, 1969, 1975; Torvik, 1969; Davis and Wu,
1972). In using this theory, each of the mineralogical constituents of a rock is modeled as a separate continuum and the interactions between continua accounted for. The pertinent theory and accompanying assumptions are outlined below. Application to a terrestrial analog of the lunar sample and the lunar sample is also presented.

We assume that the total mass density of the rock is expressed as the sum of the mass densities of each mineral per unit volume of total mixture,

$$\rho = \sum \rho_m$$  \hspace{1cm} (1)

where \( N \) is the number of minerals. Upon being encompassed by a shock, the mean velocity of the rock is given by the mass weighted average

$$V' = \frac{\sum C_m V_m}{\rho}$$  \hspace{1cm} (2)

where \( V_m \) is the velocity component of the \( S_m \) mineral and \( C_m \) is the mass fraction defined by

$$C_m = \frac{\rho_m}{\rho}$$  \hspace{1cm} (3)

The vector component of diffusion velocity \( U' \) of a single mineral relative to the mean rock particle velocity is

$$U' = V_m - V'$$  \hspace{1cm} (4)

The component of the total stress tensor is

$$\tau_{ij} = \sum C_m (\tau_{ij}^m - \rho_m U_m U_m)$$  \hspace{1cm} (5)

where \( \tau_{ij}^m \) is called the partial stress tensor for the \( S_m \) constituent. The internal energy, \( e \), is defined as

$$e = \sum C_m (e_m + \frac{1}{2} U_m U_m)$$  \hspace{1cm} (6)

where \( e_m \) is the internal energy per unit mass within a given mineral.

In the case of a shock front in a rock, the balance of mass for each mineral is

$$[\rho_m U_m] = \beta_m$$  \hspace{1cm} (7)

where \( \beta \) is the surface supply of mass, and the notation \( [\cdots] = \cdots \) \( \Psi_m - \Psi_m \) implies that \( \Psi \) and \( \Psi \) are the limiting values of the function in brackets at two points on either side of the shock propagating at velocity \( U_m \), and

$$U_m = U - V_m$$  \hspace{1cm} (8)

where \( V_m \) is the particle velocity of the mineral, \( S_m \) normal to the shock surface.

The balance of momentum for each constituent is

$$[\rho_m U_m V_m] + [\beta_m N_m] = \gamma_m$$  \hspace{1cm} (9)
where \( y_{1} \) is the surface supply of momentum and \( N_{t} \) is the component of a unit vector normal to the shock surface.

The balance of energy is given by

\[
(U_{xx}, U_{yz}, U_{zy}) + (1, 1, 1, 1) N_{t} = \epsilon_{xx},
\]

(10)

where \( \epsilon_{xx} \) is the surface supply energy for material \( S_{x} \), and the assumption has been made that no heat transfer occurs within a given mineral.

The above equations are in terms of mixture constituent mass densities and partial stresses. We will now relate these quantities to the mineral properties.

The mass density in terms of the crystal density \( \rho_{c} \) is defined as

\[
\rho_{x} = n_{c} \rho_{c},
\]

(11)

where \( n_{c} \) is the ratio of the volume of a constituent to the volume of the mixture.

The volume fraction has the following relationship

\[
\sum_{i} n_{c} = 1
\]

(12)

The partial stress tensor in terms of the crystal stress tensor \( \sigma_{y} \) is given by

\[
\sigma_{y} = n_{c} \sigma_{c}
\]

(13)

The crystal internal energy \( \epsilon_{c} \) is the same as the mixture internal energy since they are both defined per unit mass, i.e.,

\[
\epsilon_{c} = \epsilon_{x}
\]

(14)

We assume that the stress state behind the shock is hydrodynamic

\[
\sigma_{y} = -P_{c} I - n_{c} P I
\]

(15)

where \( P_{c} \) is the unit tensor, \( P_{c} \) is the partial pressure and \( P \) is the total pressure which is the same for all minerals. In addition to the above equations, the equations of state of each constituent are required. The Mie–Grüneisen equation of state was used for each constituent

\[
P_{c} = \rho_{c} \left( T - 300^\circ K \right) + \rho_{c} \gamma_{c} \left( \epsilon_{p}, \rho_{c}, \rho_{c}, T_{c} \right) - \epsilon_{c, x} - \epsilon_{c, x}
\]

(16)

where \( P_{c} \left( T - 300^\circ K \right) \) is the 300K isotherm, \( \epsilon_{c, x} \) is the transition energy in the case of complete phase changes, and \( \gamma_{c} \) is the Grüneisen parameter (Abaren, et al., 1960b). We assume that the 300K isotherm is described by the Birch–Murnaghan equation (Birch, 1952) in which \( K_{0} \) is the zero pressure isotherm bulk modulus and \( \delta \) is

\[
3 - \frac{5 K_{0}}{4 P}
\]

The internal energy as a function of density and temperature was calculated from Debye theory

\[
\epsilon_{c, x}(\rho_{c}, T_{c}) = \int_{1, 4 K} C_{c, x}(\rho_{c}, T_{c}) dT
\]

(17)
where $C_0$ is the Debye form of specific heat at constant volume. The dependence of the Debye temperature, $\theta_0$, and the Gruneisen parameter on volume is given by Davies' (1973) extension of Eulerian finite strain theory as

$$\theta_{\text{Debye}} = \theta_{\text{Debye}}^0 (1 + \gamma_{\text{Debye}})$$

$$\gamma_{\text{Debye}} = \left( \frac{1 + f_a \left( \theta_{\text{Debye}}^0 + f_a \rho_{\text{Debye}}^0 \right)}{\theta_{\text{Debye}}^0} \right)$$

where we assume for the high pressure phases that

$$\theta_{\text{Debye}}^0 = 164 \rho_{\text{Debye}}^0$$

(Anderson, 1965)

and $f_a$ is defined as a strain parameter linear in atomic displacements

$$f_a = \left( \frac{\Delta \mathbf{u}}{\rho_{\text{Debye}}^0} \right)$$

and

$$\rho_a = \rho_{\text{Debye}}^0$$

$$\Delta \mathbf{u} = \frac{d}{d V_{\text{Debye}}} \left( \frac{d \ln \gamma_{\text{Debye}}}{d \ln V_{\text{Debye}}} + \rho_a + 2 \right)$$

We also assume that (Davies and Guffney, 1973)

$$\left( \frac{d \ln \gamma_{\text{Debye}}}{d \ln V_{\text{Debye}}} \right) = 1$$

The individual constituent balance equations when summed over all constituents are required to give the well known Rankine-Hugoniot equations for a single continuum. This requires the additional set of constraints

$$\sum_j x_j R_{ij} = 0$$

$$\sum_j x_j C_{ij} = 0$$

$$\sum_j x_j \dot{C}_{ij} = 0$$

Because individual minerals have different equations of state, the shock temperatures are, in general, different. If the shock wave has a duration much greater than the relaxation time for thermal transport between minerals then the constituent will be in thermal equilibrium, that is, they would all have the same temperature. However, meteoroid impacts involving centimeter-sized objects and laboratory experiments have shock wave durations of the order of a microsecond, which precludes significant thermal transport between minerals. With the goal in mind of examining the present data, we will assume that there is no thermal transport between constituents.

$$\dot{E}_{\text{const}} = 0$$
In addition, we have assumed that the rock is in dynamic equilibrium,

$$U^* = 0$$  \hspace{1cm} (27)$$

that is, all of the particle velocities behind the shock wave are the same, and the diffusion velocities of Eq. 4 are zero. The assumption that the rock is in dynamic equilibrium is very reasonable for rocks of low porosity. A violation of this assumption would imply that in a shocked rock disymmetric minerals separate relative to their unshocked positions. This condition has not been observed in this section analysis of shocked specimens. The balance equations (Eqs. 7-10) along with the equations of state (Eq. 18) and the assumptions (Eqs. 26, 27) can be shown to result in 2N + 2 equations in 2N + 2 unknowns. In calculating a mixture Hugoniot, we have chosen to fix the pressure, P, and calculate the 2N + 2 unknowns, $p_{e0}$, $T_{e0}$, $U_{e0}$, and $V$. The use of the above synthesis technique requires a knowledge of the equation of state parameters ($K_e$, $K''$, $y_e$, and $\delta$) of each mineral and, implicitly, the crystal structure. The crystal structures and equation of state parameters and the uncertainties in their values are discussed by Ahrens et al. (1969b) and Davies and Gaffney (1973). In the following analysis, we make calculations using several possible hcp crystal structures and compare these to the experimental data. However, non-uniqueness with regard to solving the hcp crystal structures of multi-mineralic rocks exists with the present treatment.

**Theoretical Equations of State and Predicted Shock Conditions for Melting and Vaporization**

We have used the mixture theory described in the previous section to synthesize the Hugoniots of Frederick diabase and the lunar gabbric orthosphilithe. Previous studies of terrestrial basalts, diabases, and other plagioclase-bearing rocks (Ahrens and Gregory, 1964; McQueen et al., 1967; Ahrens et al., 1969b) have demonstrated that the Hugoniot can be considered as describing states in essentially three regimes, a low pressure (untransformed) regime, a mixed phase regime, and a fully transformed or high pressure phase regime (Ahrens et al., 1969b).

Because our experimental Hugoniot data for 15.418 extend only over the low pressure and mixed phase regimes, a mixture theory calculation was carried out for the high pressure phase regime. In order to have confidence in this calculation, we computed the Hugoniot of a similar but less plagioclase-rich rock, Frederick diabase. For this rock, which is well-characterized mineralogically, Hugoniot data are available in the mixed and high pressure phase regime (McQueen et al., 1967).

The crystal density of 15.418 was calculated by using both (a) the results of our microprobe analysis (Table 1) of plagioclase, clinopyroxene and olivine and the normative compositions from the LSPET (1972) analysis and (b) the LSPET normative compositions and the whole rock analysis upon which these are based. Using the molar volumes in Clark (1967), zero-pressure crystal densities of 2.92 g/cm$^3$ and 2.53 g/cm$^3$ are obtained for cases (a) and (b), respectively. The porosity was determined to be 3.25% (m = 1.0325). The 2.92 g/cm$^3$ value for the density was adopted in the construction of the equation of state model.
described below. The observed and modeled volume fractions are listed in Table 3. Frederick Diabase is approximately 50% plagioclase and 50% pyroxene (by volume). For simplicity, we adopted these values and a non-porous zero-pressure density of 3.01 g/cm$^3$. The plagioclase components of the rocks were modeled as the anorthosite shocked by McQueen et al. (1967). The pyroxene components were modeled as a solid solution of enstatite and ferrosilite and the fraction of these chosen so that the zero-pressure value of the density of rock was matched. The observed volume fraction of plagioclase of both rocks was exactly matched and the Fe to Mg ratio of the model pyroxene was adjusted to exactly match the zero-pressure density, i.e., pyroxene was substituted for the minor quantities of other minerals (mica and olivine) present. The equation of state parameters for the plagioclase and pyroxene components of both rocks are listed in Table 4. (The equations of state parameters for the pyroxene were calculated by taking a mass fraction average of the properties of enstatite and ferrosilite.)

The calculated Hugoniot for Frederick diabase along with the experimental data are shown in Fig. 4. In contrast to the case for 15,418, there are no

<table>
<thead>
<tr>
<th>Rock</th>
<th>Mineral</th>
<th>Phase</th>
<th>$m$</th>
<th>$\rho_0$ (g/cm$^3$)</th>
<th>$K_0$ (Mbar)</th>
<th>$\varepsilon$</th>
<th>$\tau$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Frederick</td>
<td>Plagioclase</td>
<td>lpp</td>
<td>1.0</td>
<td>2.73</td>
<td>0.74</td>
<td>-0.3</td>
<td>0.45</td>
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<tr>
<td>Diabase</td>
<td>Pyroxene</td>
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<td>3.29</td>
<td>1.2</td>
<td>-0.6</td>
<td>1.0</td>
</tr>
<tr>
<td>(n = 3.0 g/cm$^3$)</td>
<td>Plagioclase</td>
<td>lpp</td>
<td>1.0</td>
<td>3.71</td>
<td>1.2</td>
<td>0.7</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Pyroxene</td>
<td>Majorite</td>
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<td>3.68</td>
<td>-0.27</td>
<td>1.5</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Enstatite</td>
<td>1.0</td>
<td>3.89</td>
<td>2.3</td>
<td>-0.26</td>
<td>1.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Ferrosilite</td>
<td>1.0</td>
<td>4.49</td>
<td>3.78</td>
<td>1.3</td>
<td>1.5</td>
</tr>
<tr>
<td>Apollo 15,418</td>
<td>Plagioclase</td>
<td>lpp</td>
<td>1.0325</td>
<td>2.73</td>
<td>0.34</td>
<td>-0.5</td>
<td>0.45</td>
</tr>
<tr>
<td></td>
<td>Pyroxene</td>
<td>lpp</td>
<td>1.0325</td>
<td>3.45</td>
<td>1.2</td>
<td>-0.6</td>
<td>1.0</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Majorite</td>
<td>1.0325</td>
<td>3.83</td>
<td>1.9</td>
<td>-0.33</td>
<td>1.5</td>
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<tr>
<td></td>
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<td>Enstatite</td>
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<td>2.38</td>
<td>-0.19</td>
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<tr>
<td></td>
<td></td>
<td>Ferrosilite</td>
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<td>4.7</td>
<td>3.73</td>
<td>-1.9</td>
<td>1.5</td>
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</table>
Fig. 4. Experimental and theoretical Hugoniot for Frederick diabase. The bars on the abscissa indicate the melting ranges of the minerals. The Ipp curves are calculated assuming a 3.71 g/cm³ plagioclase equilibrated and the bpp components indicated for the pyroxene.

Experimental shock data for the low pressure phase (Ipp). The intersection of the trend of the data in the mixed phase regime with the theoretical Hugoniot for the Ipp indicates that there is an onset of a phase change(s) at pressures of ~140 kbar.

A series of high-pressure phase Hugoniots (Ipp) were calculated assuming that the phase changes were complete. The plagioclase in all cases was assumed to go to an unknown crystal structure having the zero-pressure density of 3.71 g/cm³ inferred by Davies and Anderson (1971). However, we feel that this assumption is difficult to defend on the basis of evidence, other than the good fit to the data of McQueen et al. (1967), since no Ipp corresponding to this density has yet been recovered from either static or dynamic experiments. Also this assumption was made only after we were unsuccessful in fitting Hugoniot data above 600 kbar for the gabbroic rocks with the hollandite-type structure zero-pressure density of 3.84 g/cm³. The response of pyroxene is quite complex as it is thought to undergo a series of transformations with increasing shock pressure (Abrams et al., 1969b; Davies and Gaffney, 1973). The bpp crystal structures assumed in the calculations were the majorite, ilmenite, and perovskite structures. Referring to Fig. 4, the high-pressure Hugoniot calculated by assuming the perovskite structure for the high-
pressure phase agrees with the experimental data quite well. The Hugoniot calculated assuming the hpp's had majorite and ilmenite structures are not directly related to the shape of the Hugoniot in the mixed phase regime, but are indicative of when phase transformations occur and go to completion. The description of the Hugoniot in the mixed phase regime is complicated by the fact that the phases are probably not in thermodynamic equilibrium, and the degree of reaction is probably controlled by the kinetics of the phase transformation.

The shock pressures required to melt the individual minerals were determined by calculating the entropy increase along the shock and comparing these values to the experimental values at one atmosphere (Ahrens and O'Keefe, 1972). This procedure has the implied assumption that the release path from the shock state is isentropic. The shock temperatures and entropies of each of the minerals are listed in Table 5 and the melting regimes plotted in Fig. 5. Referring to Fig. 4 for Frederick diabase, the shock pressure for incipient and complete melting of plagioclase is 0.6 and 0.67 Mb respectively, and the pressure for incipient and complete melting of pyroxene is 0.625 and 0.775 Mb respectively, and overlaps the melting range of plagioclase.

The calculated Hugoniot of 15.418 is shown in Fig. 5. The theoretical Hugoniot for the hpp correlates well with the experimental data shown, and passes through

![Fig. 5. Experimental and theoretical Hugoniots of 15.418 and Frederick diabase.](image-url)
Table 5. Calculated temperatures and entropies of minerals in shocked Fredericidabase and Apollo 15,418.

<table>
<thead>
<tr>
<th>Fredericidabase</th>
<th>Apollo 15,418</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mineral</td>
<td>P (GPa)</td>
</tr>
<tr>
<td>Orthopyroxene</td>
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</tr>
<tr>
<td>Clinoptyroxene</td>
<td>0.7</td>
</tr>
<tr>
<td>Enstatite</td>
<td>0.9</td>
</tr>
</tbody>
</table>

The table shows the calculated temperatures and entropies of minerals in shocked Fredericidabase and Apollo 15,418. The temperatures are in degrees Celsius (°C) and the entropies are in joules per kelvin per mole (J/K·mol).

The conclusions and discussions section begins with: The dynamic yield stress under shock compression (the HEL) is observed to vary from ~45 to ~70 kbar, apparently depending on final shock pressure. These values, which probably give a lower bound to the shock pressure range above which major shock-induced deformations occur, are somewhat higher than the ~45–50 kbar HEL values previously reported for terrestrial basalts. The reason for this may be the lack of a small component of usually hydrous alteration minerals often present in terrestrial rocks. The higher than terrestrial value of the HEL implies that a welded rock such as 15,418 may require more energy to achieve comminution than a terrestrial counterpart. Above the HEL and to ~150 kbar our
data agree closely with the theoretical pressure-density hydrostat inferred from ultrasomice data on pyroxene and plagioclase single crystals. The Hugoniot points are also in approximate agreement with earlier data for a pyroxene and amphibole-rich basalt (Vacaville basalt). Since the data for both single-crystal pyroxene (Abreu and Gaffney, 1971; McQueen et al., 1967) and various plagioclases (McQueen et al., 1967; Abreu et al., 1969) demonstrate that shock-induced phase changes take place in these materials above ~135 and ~140 kbar, respectively, it is not surprising that a mixed phase regime occurs above ~140 kbar in the 15.418. However, above 150 kbar and extending to 282 kbar the present pressure-density data lie at significantly greater densities (by about 0.12 g/cm$^3$) than the earlier results for Vacaville basalt and Frederick and Centreville diabases (McQueen et al., 1967).

We suggest that the higher apparent compression of 15.418 in the mixed phase region arises from the higher iron content of its pyroxene in contrast to the terrestrial basalts. Although we have not yet been able to describe the transformation kinetics in detail, we suspect that the higher Fe$^2+$ content allows the extent of reaction of the shock-induced phase change to be greater for a Fe$^2+$-rich pyroxene, for a given shock pressure, than for a more Mg$^2+$-rich pyroxene. (For Mg-Fe solid-solution silicates transformation to the high-pressure denser polymorph always occurs at a significantly lower pressure for the Fe member.)

We have examined the Hugoniot data for Frederick diabase with the Kelly-Truesdell theory of mixtures using mineral equations of states obtained from the mineral properties' systematics of Davids and Gaffney (1973), with the result that the high pressure phase data could be closely reproduced in the range of 0.6-1.0 Mbar. The plagioclase component was assumed to be equivalent to the anorthosite shocked by McQueen et al. (1967), which is assumed to transform into an unknown structure at high pressures (inferred zero-pressure density of ~3.71 g/cm$^3$). Although the equilibrium phase assemblages resulting from the application of high static pressures and high temperatures to the Ca- and Na-rich plagioclase are quite different, similar behavior of these minerals under shock condition is observed, i.e., the Hugoniot of albite and anorthite are observed to nearly coincide (Abreu et al., 1969). The pyroxene component of the diabase was assumed to be in the perovskite structure (zero-pressure density of 4.7 g/cm$^3$).

We did not carry out calculations in the mixed phase regime (150 to 600 kbar), however, the calculations we performed do suggest that the "incipient," high-pressure phases having a garnet structure (majorite) and ilmenite structure might form at pressures less than 600 kbar. A theoretical hydrostatic was calculated for the low-pressure phases and from the intersection of the trend in the data in the mixed phase regime a phase change at 140 kbar is inferred. As far as we know, all the phase changes are reversible. They are, however, of considerable importance in calculating the shock conditions necessary for impact-induced melting or vaporization.

A calculation similar to that for the Frederick diabase was carried out for the low- and high-pressure regimes of 15.418. The results predict the shock-induced phase transformation implied by the data should go to completion by ~400 kbar.
We note that although the pyroxene in 15,418 is more iron-rich than that in terrestrial rocks, because of the higher plagioclase content of this rock, the zero-pressure density was slightly less than that of previously studied terrestrial diabases.

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REFERENCES


