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PLANE SHOCK WAVE STUDIES OF GEOLOGIC MEDIA

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MASTER

October 31, 1977

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PLANE SHOCK WAVE STUDIES OF GEOLOGIC MEDIA

ABSTRACT

Plane shock wave experiments have been conducted on eight geologic materials in an effort to determine the importance of time-dependent mechanical behavior. Of the eight rocks studied, only Westerly granite and nugget sandstone appear to show time independence. In the slightly porous materials (1-5%), Blair dolomite and sodium chloride, and in the highly porous (15-40%) rock, Mt. Helen tuff and Indiana limestone, time-dependent behavior is associated with the time required to close the available porosity. In water-saturated rocks the time dependence arises because the water that is present shows no indication of transformation to the higher pressure ice phases, thus suggesting the possibility that a metastable form of water exists under dynamic conditions.

INTRODUCTION

Experimental descriptions of the mechanical properties of geologic materials usually include data obtained from quasi-static, acoustic and shock wave tests. These data form the basis for constitutive models used in computer calculations to test computer simulation. The Lawrence Livermore Laboratory's seismic monitoring program uses this overall approach as one means of developing relative coupling efficiencies and/or seismic source functions for a variety of geologic materials. These source functions would provide the basis for calculation of source spectra as a function of material properties and, given the proper transfer function, would allow a calculation of the ground motion at any seismic station.

The physical processes involved in the initiation of seismic waves by explosions in geologic materials are very complex. Many important nonlinear processes, such as melting, polymorphism, material failure, and crushing of pores, take place before energy is coupled into the "elastic" region. Some of these nonlinear processes are time dependent and, because the *in situ* events that we are trying to simulate are dynamic, require a measurement of the characteristic times or rate constants for the process. Therefore, as part of the effort to search for time-dependent effects, plane shock wave experi-

ments were conducted on the eight materials listed in Table 1. Additional plane shock wave data on coal and several frozen materials appear in two publications, *Shock Wave Studies of Subbituminous Coals*¹ by Anderson and Larson, and *Shock Wave Studies of Ice and Two Frozen Soils*² by Larson, Bearson and Taylor.

Plane shock wave experiments provide data at extremely high strain rates (10^5 - 10^7 sec⁻¹) and normally allow less than 10 μ s of observation time under loaded conditions, i.e., incubation time (see Fig. 1). Consequently, these tests are sensitive to some processes that are time dependent, such as phase transformations, pore collapse, etc., and thus could provide valuable insight into the mechanisms of these processes. In most seismic monitoring applications, such as nuclear or high explosive events *in situ*, the dispersive nature of the medium and the size of the energy source lead to strain rates in the range of 10 - 10^3 sec⁻¹ and incubation times on the order of milliseconds. These rates and times are much different from those observed in laboratory scale, shock wave experiments. Nevertheless, the characteristic times or rate constants observed in shock wave experiments, together with quasi-static data, should provide valuable guidelines for use in the various applications.

Table 1. Properties of the geologic materials under ambient conditions.

Geologic material	Bulk density ^a (Mg/m ³)	Total porosity ^b (%)	Dry porosity (%)	Longitudinal sound speed (km/s)
Westbury granite (dry)	2.65	1	1	4.8
Nugget sandstone (dry)	2.42-2.55	4-9	4-9	3.4 (1. to bedding plane)
Blair dolomite (dry)	2.84	1	1	5.0
Polycrystalline NaCl (dry)	2.13	1.6	1.6	4.1
Mt. Helen tuff (dry)	1.46	40	40	2.5-2.7
Mt. Helen tuff (sat'd)	1.86	40	2	2.6
Indiana limestone (dry)	2.28	16	16	4.2
Indiana limestone (sat'd)	2.41-2.48	14-20	0	4.4

^aAverage bulk densities were determined by measuring and weighing the various samples in air.

^bAverage total porosity was calculated using grain densities.

Figure 1. Schematic drawing of a particle velocity gage record. Time from C to D is incubation time.

EXPERIMENTAL TECHNIQUES

We conducted shock wave experiments using a light-gas gun so that the test material was loaded under uniaxial strain. With this gas-gun technique, flat-faced projectiles are accelerated with expanding helium gas and allowed to impact a target composed of plates of the test material, as shown schematically in Figs. 2 and 3. In this way, the shock wave produced by the projectile impact loads the sample; several microseconds later, a rarefaction from the free surface causes unloading, whereby the test material is relaxed to ambient stress conditions. Because of the geometry of these experiments, both loading and unloading processes take place under conditions of uniaxial strain.

We used diagnostic gages of thin (2.5×10^{-2} mm) brass foils. These foils were placed on the surfaces

and also sandwiched between the plates of the target material. Because the gages were thin, they equilibrated to the surrounding material velocity in nanoseconds, thus following the motion of the sample material as both loading and unloading occurred. The foils were also constrained to move perpendicular to the magnetic field produced by a large external electromagnet. Under these conditions, the particle velocity-time history, $U_p(t)$, is proportional to the recorded electromotive force, $\epsilon(t)$:

$$U_p(t) = \epsilon(t)/BL \quad (1)$$

where B is the magnetic field strength and L is the gage element length. During the fabrication of the experimental assemblies, the gages were placed on

Figure 2. (a) Schematic diagram of a gas-gun experiment. Lateral dimensions of the target are larger than the projectile, and the target is thin compared to the projectile diameter to ensure unperturbed recording at the gages until unloading has been completed. (b) Diagram showing the effective length L of the brass foil used in the particle velocity gage. The value, determined in four experiments is $0.899 \text{ cm} \pm 1\%$.

carefully machined surfaces (flat and parallel to approximately 0.0025 cm) of plates of the geologic material. In dry rocks, epoxy was used as a bonding agent to hold the plates together and to fill the voids between plates. In the water-saturated materials, water was used to fill the small air gaps between plates, and the plates were held together mechanically by the target assembly. In general, the projectile inserts were made of polymethyl methacrylate (PMMA), but for the higher impedance rock, projectile inserts of the same rock as the target material were used to provide a greater stress range. The configuration shown in Fig. 2 was used for dry target material, and an enclosed target assembly shown in Fig. 3 was used for saturated

samples to prevent water loss. In the enclosed assemblies, a void between the rear plate of the rock and the rear of the target assembly provided the free surface conditions necessary to unload the sample in uniaxial strain.

The characteristics of a typical particle velocity-time profile, recorded by a particle velocity gage, are shown schematically in Fig. 1. The ordinate is either voltage, as recorded on film from an oscilloscope trace, or the corresponding particle velocity obtained by reducing the data using Eq. (1). In either case, the abscissa is time. The pulse at A is a fiducial mark which is simultaneously recorded on each oscilloscope record, so that times can be correlated between the several gages recording a

Figure 3. Schematic for dynamic equation-of-state experiments with water-saturated samples. The PMMA buffer and vacuum plates isolate the sample from the gas-gun vacuum. The foil at the PMMA-sample interface records the input particle velocity. Other gages record the particle velocity history when the sample is first loaded and then unloaded as the shock waves rarify at the free surface.

given experiment. At point B, the shock wave arrives at the gage and imparts to it a velocity of magnitude C. From time C to time D, the gage moves with a constant velocity while the shock front itself has propagated toward the rear free surface. When it reaches the rear free surface, the shock wave is reflected as a rarefaction wave that imparts an additional particle velocity to the material and, hence, to the gage. The additional increment in particle velocity imparted by this rarefaction is in the same direction as the increment imparted by the shock wave, although the rarefaction is propagating in the opposite direction. At point D in Figure 1, the rarefaction arrives at the gage and accelerates it to velocity E. If the portion of the record BC is essentially the same from gage to gage in a given

material, the shock wave is considered to be a steady wave. In this case, the time duration between points B or C (or any corresponding points in between) on any two gage records, together with the distance between those gages, may be used to compute a shock velocity. This shock velocity, together with the particle velocity given by the plateau level CD, is sufficient to calculate a point on the Hugoniot curve for the material. If the material is dispersive under shock loading, as it might be if the shock velocity is strain-rate sensitive or the material exhibits dynamic yielding, pore collapse, or phase transitions, a steady state wave will not develop. In such cases, the particle velocity rise from points B to C will have a more complex structure and will vary from gage to gage. The portion of the profile from

point D to E, which represents a relief of the stress in the material by the rarefaction wave, is generally unsteady and flattens out with propagation distance in all materials in which the sound speed decreases with decreasing stress. However, even if the shock wave and the rarefaction wave are both unsteady, a set of particle velocity-time profiles from a single experiment can be used to deduce certain properties of the stress-strain behavior of materials under dynamic loading conditions. For example, the first arrival of the rarefaction, point D, can be compared for two or more gages, and used to calculate the longitudinal sound speed, C_1 , in the compressed state.

The data from each of the experiments were recorded on Polaroid film as voltage-time histories. These records were digitized and then converted from voltage-time histories to particle velocity-time histories using Eq. (1). The conversion of these particle velocity-time history records to equation-of-state data requires expressions relating particle velocity history to the material properties of interest (i.e., stress and specific volume). These expressions can be derived by using the conservation equations for uniaxial flow.³ The equations for conservation of mass and linear momentum are

$$\left(\frac{\partial \rho}{\partial t}\right)_h + \rho \left(\frac{\partial U_p}{\partial x}\right)_t = 0 \quad (2)$$

$$\rho \left(\frac{\partial U_p}{\partial t}\right)_h + \left(\frac{\partial \sigma}{\partial x}\right)_t = 0 \quad (3)$$

where ρ is the density, σ the axial stress, x the Eulerian space coordinate, h the Lagrangian space coordinate, and t the time. In our experiments, particle velocities were measured at a constant Lagrangian coordinate. Therefore, using the relationship

$$\left(\frac{\partial h}{\partial x}\right)_t = \rho / \rho_0 \quad (4)$$

for uniaxial flow, and the identities

$$C_{U_p} \left(\frac{\partial t}{\partial U_p}\right)_h \left(\frac{\partial U_p}{\partial h}\right)_t \equiv C_\sigma \left(\frac{\partial t}{\partial \sigma}\right)_h \left(\frac{\partial \sigma}{\partial h}\right)_t \equiv -1$$

where

$$C_{U_p} = \left(\frac{\partial h}{\partial t}\right)_{U_p} \quad \text{and} \quad C_\sigma = \left(\frac{\partial h}{\partial t}\right)_\sigma$$

Eqs. (2) and (3) can be rewritten as

$$\left(\frac{\partial V}{\partial U_p}\right)_h = -\frac{V_0}{C_{U_p}} \quad (5)$$

$$\left(\frac{\partial \sigma}{\partial U_p}\right)_h = \rho_0 C_\sigma \quad (6)$$

where $V = 1/\rho$ is the specific volume

In the case of a steady wave (i.e., $C_{U_p} = C_\sigma = a$ constant, where the constant is U_p , the shock velocity), integration of Eqs. (5) and (6) yields the jump conditions

$$V = V_0 \left(1 - \frac{U_p}{U_s}\right) \quad (7)$$

$$\sigma = \rho_0 U_p U_s \quad (8)$$

Equations (7) and (8) were used in the analysis of the shock state in several of our experiments, i.e., whenever steady loading conditions prevailed, to obtain stress and specific volume from measured shock and particle velocities. In the analysis of the remaining data, it was assumed that the nonsteady flow was that of simple waves. In the case of nonsteady simple waves, the phase velocities at constant stress and particle velocity, C_σ and C_{U_p} respectively, are equal but are functions of particle velocity rather than constants. In this case, the digitized particle velocity-time data from the experiments can be used to compute C_σ and C_{U_p} ; then numerical integration of Eqs. (5) and (6) can be used to obtain the relationship between stress and specific volume throughout the nonsteady wave. This integration was accomplished with a computer code called GANDALF, which is our interactive version of the Stanford Research Institute Lagrange Gage code (see Ref. 4).

EXPERIMENTAL RESULTS

In this section we have presented the experimental results in tabular and graphical form and have highlighted features of the data which are significant. The format in which the data and results are presented is similar for each geologic material. This format consists of the following:

1. Tabulated data — Entries in the tables refer to the Hugoniot state behind the shock. The quantity $\mu = \rho/\rho_0 - 1$ is the compression, and the quantity C_L is the longitudinal sound speed in the compressed material behind the shock.

2. Particle velocity-time profiles — Curves are examples of Lagrangian particle velocity-time histories as recorded by the gages on a single test. They are presented on a common time base and exhibit the observed characteristics of the various materials.

3. Compression curves — Curves show the Hugoniot points and, in some cases, the loading and unloading paths followed by the material in normal stress (σ)-compression (μ) space. Quasi-static compression data or hydrostatic pressure-compression data and, in some cases, quasi-static uniaxial strain data are included for comparison.

4. Shock velocity vs particle velocity plots — Hugoniot curves in which the shock velocity vs the particle velocity is plotted. Abrupt changes in slope or deviations from linear behavior of these curves are indicators of the occurrence of such phenomena as phase changes, failure phenomena, or irreversible pore collapse.

5. Sound speed vs density — Plots of the longitudinal sound speed behind the shock vs the density behind the shock. The value of the sound speed at ambient density is taken from ultrasonic data as referenced.

Westerly Granite

Westerly granite was obtained from Bonner Monument Company of Westerly, Rhode Island. It was the blue variety with an average density of 2.65 Mg/m³ and a porosity of approximately 1%.

Thirteen experiments were conducted using Westerly granite, and the results are presented in Table 2. In several of the experiments, the impacting projectile head was PMMA; in the remainder, those in which the particle velocity is half the projectile velocity, the impactor was granite. Figure 4a is a particle velocity-time history characteristic of this material. These data show no evidence of a precursor, but in some particle velocity histories, evidence of a slight precursor was observed by Swift⁵ and Larson⁶ in the vicinity of 2.5-3.0 GPa. Rosenberg⁷ also reported a precursor under plane wave loading conditions in relatively thick samples of dry granite recorded by stress gages. This precursor has never been observed to be well developed and separated from the main wave; it shows little deviation from

Table 2. Hugoniot data for dry Westerly granite.

Shot No.	Projectile velocity (km/s)	Shock velocity (km/s)	Particle velocity (km/s)	Stress (GPa)	Density (Mg/m ³)	Volume (m ³ /Mg)	μ	C_L (km/s)
GR-1	0.582	5.24	0.13	1.8	2.717	0.368	0.0254	6.64
GR-2	.495	5.37	.104	1.48	2.702	.370	.0198	6.30
GR-3	.408	5.14	.088	1.20	2.696	.3709	.0174	6.55
GR-4	.316	5.03	.067	0.89	2.685	.372	.0135	—
GR-5	.227	5.05	.048	0.64	2.675	.3737	.0096	6.07
GR-7	.485	(5.50)	.242	3.53	2.772	.360	.0460	—
GR-8	.573	5.6	.287	4.26	2.79	.358	.0540	6.95
GR-9	.76	5.23	.380	5.26	2.85	.351	.0783	—
GR-10	.79	5.60	.395	5.86	2.85	.350	.0759	—
GR-11	0.63	5.57	.315	4.59	2.77	.360	.0599	—
GR-12	1.21	5.18	.605	8.18	2.95	.338	.1322	—
GR-13	0.81	5.38	.405	5.68	2.82	.354	.0814	—
GR-14	0.98	5.56	0.490	7.12	2.86	0.349	0.0966	—

Figure 4. Experimental data for Westerly granite. (a) Particle velocity-time histories from one gas-gun experiment. (b) Hugoniot and hydrostatic data. (c) Shock velocity-particle velocity Hugoniot data. (d) Hugoniot data and quasi-static uniaxial strain data. (e) Longitudinal sound speed vs density.

steady wave behavior. Consequently, our data indicates that loading and unloading in the stress-compression (σ - μ) plane occurs essentially along a straight line, as shown in Fig. 4b. However, the plot of shock velocity vs particle velocity (Fig. 4c) exhibits a change in slope between the lower and higher stress data. With the two straight-line segments drawn as they are in Fig. 4c, the discontinuity in slope occurs at a stress of approximately 2.0 GPa. This change in slope implies that, when it reaches stresses of approximately 2.0 GPa, the material begins to change in volume with the increasing stress more than it would have if the Hugoniot had continued along the initial straight line. Inasmuch as no phase transitions, resulting in a decrease in volume, are known to occur in granite in this stress region, we believe that this change in slope may be associated with pore collapse. Although no evidence of volume decrease attributable to pore collapse has been observed by Heard et al.⁸ in the hydrostatic experiments up to stress levels of 4.0 GPa, shear stresses which enhance pore collapse may exist in the one-dimensional shock wave experiments, but are not present in the hydrostatic experiments. The stress level of 2.0 GPa is not grossly inconsistent with the 2.5-3.0 GPa level of the slight precursor. Furthermore, since Westerly granite has only 1% porosity, one third of which is associated with microcracks (see Ref. 9) which close at hundredths to tenths of a GPa and the remaining two thirds is associated with more nearly spherical pores which could close at much higher stresses, a large volume hysteresis upon loading and unloading would not be expected.*

Figure 4d compares dynamic and quasi-static uniaxial strain data to a stress of approximately 1.5 GPa. The agreement is quite good. Figure 4e plots longitudinal sound speeds for Westerly granite. The value of 4.8 km/s at a density of 2.65 Mg/m³ was measured ultrasonically by Heard et al.⁸ and the other values were determined from the current shock wave experiments. The rapid initial increase in velocity with density is consistent with the closing of microcracks.

Nugget Sandstone

The nugget sandstone was obtained from the Parley's Canyon area near Salt Lake City, Utah.

*In this paper porosity in the form of relatively long-flat cracks which can be closed by stresses of the order of 0.1-0.2 GPa is called microcrack porosity, while porosity that is more nearly spherical in shape and requires stresses of 1-10 GPa or more to close is called pore porosity.

The material was the red variety, which contains nominally 90% quartz. The initial tests, the MF series, used material with an average bulk density of 2.55 Mg/m³ and a corresponding dry porosity of about 4%. The second series of tests was conducted on samples with bulk densities between 2.42 and 2.51 Mg/m³. The corresponding range of dry porosity in these samples is 6-9%.

Table 3 records the results from 13 successful shock wave experiments in nugget sandstone loaded perpendicular to the bedding plane. Typical particle velocity-time histories for one of these experiments are shown in Fig. 5a. In this experiment, the nugget sandstone target was impacted with a PMMA projectile. The initial jump in particle velocity is due to the shock wave; it is followed by a second wave resulting from the rarefaction propagating back from the free surface toward the impact surface. Upon reaching the impact surface, the PMMA interface, this rarefaction is reflected as a second shock, back through the sandstone toward the free surface, where it is again reflected as a rarefaction. Thus, in Fig. 5a, the shock-reflected rarefaction process occurs twice, generating four steps in particle velocity.

In all the experiments on nugget sandstone, the loading waves were steady, and the resulting Hugoniot data shown in Fig. 5b suggests quasi-elastic behavior. However, unloading curves in this stress compression (σ - μ) plane showed a slight amount of irreversible volume loss at low stresses ($\sigma < 1.5$ GPa). This slight amount of compaction is assumed to be associated with the closure of thin microcracks at very low stresses. Upon release from higher stress-shocked states ($\sigma > 5.0$ GPa), the unloading curves imply an irreversible loss in specific volume, which is approximately equal to the porosity initially present in the material.

A plot of shock velocity vs particle velocity, shown in Fig. 5c, exhibits a change in slope in the vicinity of 1.5 GPa. This phenomenon is associated with the softening of a material that would occur if pores would begin to collapse. These data suggest that the nugget sandstone matrix behaves essentially elastic up to approximately 1.5 GPa, with the exception of a small loss of porosity associated with microcrack closure at very low stress. They further suggest that, at approximately 1.5 GPa, the porous matrix begins to crush, and most of the porosity is eliminated by stresses in excess of 5.0 GPa.

Figure 5d compares dynamic and quasi-static uniaxial strain data to stresses of approximately 2.2 GPa. The agreement is quite good, suggesting little time-dependent behavior in nugget sandstone, at least up to 2.2 GPa. The longitudinal sound

Figure 5. Experimental data for Nugget sandstone. (a) Particle velocity-time histories from one gas-gun experiment. (b) Hugoniot and hydrostatic compression data. (c) Shock velocity-particle velocity Hugoniot data. (d) Hugoniot and quasi-static uniaxial strain data. (e) Longitudinal sound speed vs density.

Table 3. Hugoniot data for dry nugget sandstone.

Shot No.	Initial density (Mg/m ³)	Projectile velocity (km/s)	Shock velocity (km/s)	Particle velocity (km/s)	Stress (GPa)	Density (Mg/m ³)	μ	Sound speed (km/s)
MF-55	2.55	0.227	4.11	0.0611	0.64	2.59	0.0152	4.55
MF-56	2.55	.141	3.86	.0397	0.391	2.57	.0104	4.86
MF-51	2.551	.586	4.42	.150	1.65	2.64	.0353	5.3
MF-52	2.58	.483	4.28	.120	1.31	2.633	.0285	5.2
MF-53	2.55	.402	4.43	.103	1.17	2.614	.0235	5.28
MF-54	2.55	.321	4.10	.0824	0.86	2.599	.0204	5.14
DRNU-1	2.499	.795	4.30	.397	4.27	2.75	.101	6.25
DRNU-2	2.499	0.94	4.30	.470	5.05	2.80	.120	6.11
DRNU-3	2.507	1.23	4.48	.615	7.02	2.89	.156	6.93
DRNU-6	—	1.58	4.67	.79	9.23	3.01	.204	6.97
DRNU-7	—	0.78	4.39	.39	4.26	2.73	.097	5.78
DRNU-10	2.422	.138	3.85	.069	0.645	2.47	.0182	—
DRNU-12	—	0.51	4.19	0.255	2.58	2.58	0.0648	—

speeds measured in the various shock wave experiments are shown in Fig. 5e. The measurement at a density of 2.55 Mg/m³ was determined ultrasonically by Shock et al.¹⁰ As with Westerly granite, the rapid initial increase in velocity with a comparable increase in density is consistent with the closing of microcracks.

Blair Dolomite

The Blair dolomite was obtained from the Berkeley Company in Martinsburg, West Virginia. The measured bulk density of this fine-grained rock, which is more than 98% CaMg(CO₃)₂, was 2.84 Mg/m³ and the porosity was approximately 1%.

Table 4 tabulates the results for all six shock wave experiments in Blair dolomite, and Figure 6a shows

the particle velocity-time histories for the two internal gages in experiment MF-25. In the six experiments, the loading wave was composed of two parts: (1) a fast rising, steady part whose amplitude was dependent upon stress, and (2) a more slowly rising nonsteady part that brought the sample to its final loading state. These observations disagree with those made by Grady et al.^{11,12} who report profiles of similar shape but claim that, for loading states below 2.5 GPa, the waves are steady and only increments above 2.5 GPa show nonsteady characteristics. We currently have no explanation for this difference in behavior, but our observations are consistent with similar nonsteady behavior observed by Larson⁶ in shock wave experiments that employed spherical divergent flow. In the latter experiments, a spherical high explosive charge was used to load the dolomite, and particle velocity gages measured particle velocity-time histories at

Table 4. Hugoniot data for Blair dolomite.

Shot No.	Initial density (Mg/m ³)	Projectile velocity (km/s)	Shock velocity (km/s)	Particle velocity (km/s)	Stress (GPa)	Density (Mg/m ³)	μ	Sound speed (km/s)
MF-65	2.84	0.08	5.30	0.017	0.24	2.848	0.00288	6.00
MF-58	2.84	0.153	5.92	.030	0.44	2.852	.00431	7.34
MF-57	2.84	~0.20	6.12	.040	0.63	2.856	.00575	7.68
MF-26	2.84	0.200	6.41	.100	1.70	2.885	.0159	6.95
MF-25	2.84	0.392	6.69	.196	3.40	2.929	.0314	7.33
MF-37	2.84	0.584	6.62	0.290	5.33	2.974	0.0470	7.92

Figure 6. Experimental data for Blair dolomite. (a) Particle velocity-time histories from one gas-gun experiment. (b) Hugoniot and hydrostatic compression data. (c) Hugoniot and quasi-static uniaxial strain data. (d) Longitudinal sound speed vs density.

various radial distances from the source. In these experiments, nonsteady flow was observed at stresses as low as 0.4 GPa. Figure 6b compares Hugoniot and hydrostatic compression data observed by Heard et al.¹³

In Figure 6c, a comparison of dynamic and quasi-static uniaxial strain data shows a divergence above a stress of approximately 0.4 GPa. The reason for this divergence is unknown. Longitudinal sound speeds determined from the shock wave experiments are shown graphically in Fig. 6d.

Polycrystalline Sodium Chloride (NaCl)

The polycrystalline NaCl samples were taken from Leslie Mill feed salt of 99.4-99.9% purity. The fine-grained salt was pressed isostatically for 1 hour at 120°C using 0.136 GPa pressure to produce the large cylinders (approximately 0.36 m diam and 0.36 m long) from which the samples were taken. The measured bulk density of this material was 2.13 Mg/m³, indicating a porosity of 1.6%.

Table 5 lists the Hugoniot data obtained from the three experiments in which this material was used, and Fig. 7a shows the particle velocity-time histories recorded in one experiment. In Fig. 7a, the first gage shows the input wave that developed at the interface between the NaCl and the PMMA projectile insert. The second, third, and fourth gages in this figure clearly show the development of a precursor wave in the NaCl, which we attribute to dynamic yielding of the NaCl matrix. Both precursor wave and the wave profile immediately following it are nonsteady. We attribute this to rate effects associated with the collapse of microcracks and pores. Loading and unloading paths, calculated using Eqs. (5) and (6), are shown in Fig. 7b with a quasi-static compression curve (see Ref. 14). The unloading paths, represented by the dashed lines, show considerable hysteresis and converge towards crystal density, indicated by μ_c on the abscissa of the graph. This suggests that all, or nearly all, porosity

has been eliminated at a stress of 1 GPa. The offset of the Hugoniot states above the hydrostat (0.1-0.15 GPa) suggests that rate effects are causing different effective moduli for these extremes of strain rate.

A comparison of the dynamic and quasi-static uniaxial strain results, shown in Fig. 7c, suggests reasonable agreement for an extrapolation of the quasi-static data. Longitudinal sound speed from the shock wave experiments are shown graphically in Fig. 7d.

Dry Mt. Helen Tuff

Mt. Helen tuff was obtained from a site near Mt. Helen, southeast of Tonapah in Nye County, Nevada. It is a fine-grained rock with a porosity of about 40% and a dry density of 1.46 Mg/m³.

Hugoniot data are presented in Table 6. A typical series of particle velocity-time histories for a single experiment is shown in Fig. 8a. The first gage is located on the impact surface and shows the particle velocity imparted to the target material. Subsequent gages show the development of the wave profile, including the rarefaction from the free surface. The precursor wave shown in Fig. 8a is dispersive, with the front traveling at or near longitudinal sound speed (2.5 km/s). The formation of this wave, which has an amplitude of approximately 0.25 GPa, is attributed to dynamic failure of the rock matrix; its dispersion is probably associated with the closing of microcracks and other grain boundary effects. The second or main loading wave is traveling at a speed of approximately 1.5 km/s. We attribute the highly dispersive nature of this second wave to rate effects associated with the collapse of the pores present in this extremely porous material.

Figures 8b and 8c show Hugoniot states and unloading paths computed from the data using Eqs. (5) and (6). The figures also show a quasi-statically measured hydrostat of Heard et al.¹⁵ The unloading data from the shock wave experiments show that much, if not all, of the porosity has been

Table 5. Hugoniot data for polycrystalline sodium chloride.

Shot No.	Initial density (Mg/m ³)	Projectile velocity (km/s)	Precursor velocity (km/s)	Shock wave velocity (km/s)	Particle velocity (km/s)	Precursor stress (GPa)	Shock wave stress (GPa)	Density (Mg/m ³)	μ	Sound speed (km/s)
NN-6	2.13	0.406	4.28	3.34	0.138	~0.08	0.986	2.218	0.0414	5.50
NN-7	2.13	.594	4.42	3.48	.201	~0.12	1.483	2.258	.0603	6.10
NN-8	2.13	0.491	4.30	3.38	0.170	~0.12	1.200	2.243	0.0531	6.00

Figure 7. Experimental data for sodium chloride. (a) Particle velocity-time histories from one gas-gun experiment. (b) Hugoniot and hydrostatic compression data. (c) Hugoniot and quasi-static uniaxial strain data. (d) Longitudinal sound speed vs. density.

eliminated at stresses of 3.0 GPa. The crossing and divergence of the hydrostatic and shock wave data are not completely understood but probably result from shear enhanced compaction.* It is not completely clear why the crossover of the two sets of

*The enhanced compaction of porous materials in tests involving a shear component rather than the compaction resulting from a pure hydrostatic test.

data did not occur at a stress much closer to the point of matrix failure. However, there is the possibility of competing time-dependent effects, such as an increase in strength with strain rate, which could act to "stiffen" the material. To complete the data display, Fig. 8c includes some high stress shock wave data obtained at Stanford Research Institute.¹⁶

Figure 8. Experimental data for dry Mt. Helen tuff. (a) Particle velocity-time histories from one experiment. (b) Hugoniot and hydrostatic compression data to 4 GPa. (c) Hugoniot data to approximately 21 GPa.

Water-Saturated Mt. Helen Tuff

Mt. Helen tuff was saturated by submerging samples in water and applying a partial vacuum until air bubbles no longer appeared on the sample surfaces. Measured saturated densities averaged 1.84 Mg/m^3 , indicating that an average of more than 95% of the pores were filled.

Hugoniot data and sound speeds determined from the various experiments are listed in Table 7. Figure 9a shows a particle velocity record from one experiment. The loading wave in Figure 9a is a single wave which appears to be very nearly steady. Only two experiments, MF-75 and MF-76, showed any sign of the development of a precursor wave. In these two experiments, the specimens were loaded

Table 6. Hugoniot data for dry Mt. Helen tuff.

Shot No.	Initial density (Mg/m ³)	Projectile velocity (km/s)	Shock velocity (km/s)	Particle velocity (km/s)	Stress (GPa)	Density (Mg/m ³)	μ	Sound speed (km/s)
MF-27	1.46	0.112	2.67	0.065	0.183	1.52	0.040	2.61
MF-24	1.46	.144	2.50	.087	.220	1.54	.054	2.74
MF-23	1.46	.394	2.60	.253	.590	1.75	.198	-
MF-34	1.46	.492	2.49	.319	.765	1.84	.260	2.06
MF-35	1.46	.584	2.38	.370	.877	1.92	.313	2.48
MF-22	1.46	.558	2.67	.366	.887	1.90	.299	-
MF-14	1.46	.634	2.45	.381	.975	1.89	.292	2.95
MF-20	1.46	0.634	2.48	.398	0.995	1.92	.315	2.17
DRTF4B	1.46	1.227	2.44	.72	2.05	2.289	.568	3.56
DRTF5B	1.46	1.497	2.23	0.91	2.96	2.467	0.689	3.13

to the lower stresses; we attribute the two-wave structures to the closing of the small amount of dry porosity that was still present in the sample. The absence of a precursor wave at higher stress is apparently the result of overdriving by the main wave, i.e., the main wave has a velocity greater than the precursor.

Figures 9b and 9c show the Hugoniot points for all experiments. Figure 9b also shows the unloading paths calculated using Eqs. (5) and (6). These paths show little, if any, evidence of hysteresis, suggesting that any porosity closed by compression is recovered as the material is unloaded. Figure 9c compares the Hugoniot data with the measured hydrostat. At the lower stresses, i.e., up to 1.2 GPa,

the data are essentially superimposed, suggesting that the saturated rock has lost its strength, presumably as a result of a buildup of pore pressure. Beyond 1.2 GPa, the static and dynamic experimental data shown in Fig. 9c diverge. The hydrostat shows the volume compaction associated with the transformation of water to ice VI (at 1.2 GPa) and of ice VI to ice VII (at 2.4 GPa). However, the dynamic data give no indication of volume compaction. In fact, the shock wave data seem to suggest the formation of a metastable liquid phase in the high-pressure ice regimes. Such behavior of water under shock wave loading is not unexpected, as pure water shows the same characteristics.¹⁷

Table 7. Hugoniot data for saturated Mt. Helen tuff.

Shot No.	Initial density (Mg/m ³)	Projectile velocity (km/s)	Shock velocity (km/s)	Particle velocity (km/s)	Stress (GPa)	Density (Mg/m ³)	μ	Sound speed (km/s)
MF-48	1.84	0.588	2.64	0.271	1.285	2.06	0.118	3.35
MF-63	1.84	.591	2.52	.266	1.21	2.06	.120	3.48
MF-62	1.84	.490	2.42	.220	1.00	2.03	.101	2.96
MF-61	1.84	.412	2.32	.194	0.81	2.01	.093	2.75
MF-59	1.84	.312	2.16	.162	.62	1.99	.083	3.65
MF-60	1.84	.217	1.94	.115	.44	1.95	.058	2.94
MF-75	-	.211	-	.107	.39	1.94	.057	-
Precursor	1.84	-	2.27	.040	.15	-	.020	-
MF-76	-	.142	-	.069	.255	1.91	.037	-
Precursor	1.84	-	1.80	.042	.147	-	.020	-
WTTF-1	1.83	.65	2.57	.295	1.387	2.07	.129	3.43
WTTF-3	1.83	.76	2.73	.335	1.673	2.08	.139	3.67
WTTF-4	1.83	.893	2.84	.390	2.015	2.11	.153	(3.30)
WTTF-5	1.83	0.998	2.92	.437	2.335	2.15	.176	3.95
WTTF-6	1.83	1.16	3.09	0.488	2.82	2.16	0.195	3.83

Figure 9. Experimental data for saturated Mt. Helen tuff. (a) Particle velocity-time histories from one gas-gun experiment. (b) Hugoniot states and calculated dynamic unloading paths. (c) Hugoniot and hydrostatic compression states.

Dry Indiana Limestone

Indiana limestone, obtained from the Indiana Limestone Company quarry near Bedford, Indiana, is a calcite-cemented rock formed from shells and shell fragments. The rock is more than 90% calcite (CaCO_3), and the samples had an average bulk density of 2.28 Mg/m^3 . The measured grain density was 2.70 Mg/m^3 , which gives an average porosity of 16%.

Table 8 summarizes the data for all 10 experiments, and Fig. 10a shows the particle velocity-time histories for two internal gages in experiment DRLM-3. In this experiment, unloading originated at the projectile. The design of this experiment was similar to that shown in Fig. 1, except that a thin cylinder plate of dry Indiana limestone, backed by low density foam, was used as the projectile insert, and a relatively thick plate of dry Indiana limestone was used as the downstream plate in the target. This design was followed for the dry Indiana limestone experiment because of the large velocity difference between the first and second waves. Downstream unloading would have resulted in the unloading wave arriving at the interior gages before loading was completed. The records in Fig. 10a show the development of a precursor, which probably is due to matrix failure. Following the precursor, the main loading wave shows definite nonsteady behavior, which is attributed to the crushing of pores. As the rarefaction from the projectile arrives at the gages, the particle velocity is decreased towards zero.

Figures 10b and 10c show the Hugoniot data for all 10 experiments, along with a measured hydrostat (see Ref. 18) and some high-stress shock wave data obtained at Stanford Research Institute.¹⁶ These comparisons show the shock wave data remaining above the hydrostat at all stress levels. The apparent offset is about twice that expected from a combination of quasi-static strength and pressure compression data, but it is consistent with a 10% per decade increase in strength with an increase in strain rate. However, this interpretation of the data implies the absence of any shear enhanced compaction, which is unlikely.

Water-Saturated Indiana Limestone

The Indiana limestone was saturated with water in the same manner as the Mt. Helen tuff. The samples selected for the tests had dry densities between 2.23 and 2.28 Mg/m^3 and saturated densities between 2.41 and 2.48 Mg/m^3 . Six experiments were performed; data from these appear in Table 9.

The four gage records obtained in experiment STLM-2 appear in Fig. 11a. The loading wave shows a definite spreading as it propagates. The reason for this nonsteady characteristic is unknown, but it may be the result of a small amount of unsaturated porosity or of phase transformation occurring in the calcite. Hugoniot and hydrostatic data are compared in Fig. 11b.

Table 8. Hugoniot data for dry Indiana limestone.

Shot No.	Projectile velocity (km/s)	Precursor		Shock wave				Sound speed (km/s)
		U_p (km/s)	U_s (km/s)	U_0 (km/s)	σ (GPa)	μ	ρ	
DRLM 4	0.218	4.68	2.78	0.079	0.538	0.024	2.345	4.24
DRLM 5	.319	4.55	2.28	.122	.711	.050	2.405	4.44
DRLM 6	.420	5.15	2.40	.158	1.002	.068	2.445	3.12
DRLM 7	.503	4.91	2.22	.200	1.151	.088	2.492	3.61
DRLM 3	0.603	4.94	2.25	.254	0.1478	.112	2.546	4.10
DRLM 1B	1.21	4.62	2.70	.52	3.28	.230	2.804	
DRLM 2B	1.46	4.33	2.88	.64	4.25	.275	2.908	5.00
DRLM 4B	0.37	4.73	1.82	.185	1.06	.0942	2.49	
DRLM 5B	.802	4.59	2.43	.401	2.46	.200	2.73	
DRLM 6B	0.95	4.89	2.42	0.475	2.87	0.227	2.82

Figure 10. Experimental data for dry Indiana limestone. (a) Particle velocity-time histories. (b) Hugoniot and hydrostatic compression data to 4 GPa. (c) Shock wave data to approximately 36 GPa.

Table 9. Hugoniot data for saturated Indiana limestone.

Shot No.	Initial density (Mg/m ³)	Projectile velocity (km/s)	Shock velocity (km/s)	Particle velocity (km/s)	Stress (GPa)	μ	Sound speed (km/s)
STLM-1	2.41	0.594	2.62	0.222	1.42	0.090	3.40
STLM-2	2.44	.599	2.58	.230	1.44	.096	3.85
STLM-3	2.48	.509	2.80	.182	1.24	.070	3.59
STLM-5	2.47	.411	(2.62)	.146	0.94	.059	—
STLM-6	2.48	.317	2.48	.110	0.74	.038	—
WTLM-1	2.44	0.700	2.68	0.265	1.74	0.110	—

Figure 11. Experimental data for saturated Indiana limestone. (a) Particle velocity-time histories from one experiment. (b) Hugoniot and hydrostatic compression data.

DISCUSSION

An examination of the profiles obtained from shock wave loading of these eight rocks indicates that all, except Westerly granite and nugget sandstone, show an obvious dependence of phase velocity upon strain rate or incubation time. Because of the similarity and the relative simplicity of their behavior, we will discuss nugget sandstone and Westerly granite first. The loading waves that propagate in Westerly granite and nugget sandstone are steady shocks, which means that loading occurs along a Rayleigh line. The unloading paths show some hysteresis in the stress-compression plane, but the corresponding volume of compression is consistent with that expected from the compaction or crushing of the initial dry porosity. Analysis of the shock wave data from these two materials suggests that each has a threshold stress for crushing of the rock matrix which surrounds the pores. (This threshold stress in nugget sandstone is approximately 1.5 GPa; in Westerly granite it is between 2 and 3 GPa). At stresses below this threshold, deviations from linear or elastic behavior are small and are associated with the closing of the very thin microcracks, which account for about 1/3 of the total porosity in each of these two materials. At stresses above this threshold, the remaining porosity, which is present in the form of more nearly spherical pores, is eliminated as the rock matrix yields and collapses to fill the pores. In hydrostatic experiments on Westerly granite up to a mean stress level of 4.0 GPa, no evidence of pore collapse has been observed. The reason for the discrepancy between these observations and those in the shock wave experiments may lie in the fact that shear stresses are present in the shock experiments that would initiate pore collapse at a lower stress level than in the hydrostatic experiment.

Blair dolomite and polycrystalline NaCl are similar to Westerly granite and nugget sandstone, in that both sets of rocks have small amounts of gas-filled porosity. However, unlike the two silicate rocks, Blair dolomite and polycrystalline NaCl propagate nonsteady state shocks.

In polycrystalline salt, a definite two-wave structure is formed. Neither of these waves is steady, although the second wave is more dispersive than the first. The first wave is associated with matrix yielding, i.e., plastic flow of the salt, while the dispersion of the second wave is attributed to crushing of pores.

The nonsteady wave propagation in Blair dolomite manifests itself in the form of a single

loading wave with two components, which is different from the two-wave loading seen in polycrystalline NaCl. The first part of the two-component wave in Blair dolomite is steady, but the second part spreads as it propagates.

Our observation of nonsteady behavior at stress levels between 0.24-5.33 GPa in Blair dolomite is in contradiction to the observation of Grady et al.^{11,12} According to their data, non-steady behavior occurred only at stress levels above 2.5 GPa. This difference in behavior is presently unexplained. Nevertheless, on the basis of our data, we have drawn several conclusions. The first component, i.e., the steady part, of the wave represents a threshold stress for matrix failure, which is a complicated function of the applied stress. The second part of the loading wave is nonsteady and this nonsteady behavior probably results from the compaction of available porosity but may result from some time-dependent yielding process.

Dry Mt. Helen tuff and dry Indiana limestone have greater porosity than the rocks and salt discussed. Under shock wave loading the behavior of these two highly porous rocks is very similar. Both materials exhibit nonsteady wave propagation with a precursor wave having an amplitude of approximately 0.25 GPa. The front of this precursor wave travels at longitudinal sound speed in the material and is dispersive, producing a ramp-like structure as it propagates. This precursor wave is believed to result from the hydrodynamic instability caused by dynamic yielding of the rock matrix. Once this yielding process has started, the rock undergoes a time-dependent, irreversible compaction of the existing porosity. This process gives rise to a highly dispersive, second loading wave that eventually brings the rock to its final loaded state. The effect of such compaction is also exhibited in the behavior of the subsequent unloading process. In addition, unloading paths are similar for these two highly porous rocks and, in the stress-compression plane, are displaced significantly from the loading path. Thus the loading-unloading cycles show a good deal of hysteresis, which suggests a high degree of irreversible compaction. The Mt. Helen tuff shows nearly complete compaction of the available porosity after release from 3 GPa, and the limestone shows approximately 50% compaction after release from 4 GPa. This difference in behavior may be the result of a different distribution of pore sizes, i.e., there are more spherical pores in the limestone, or may be linked, in some complex way, to polymorphism in limestone. When the shock wave

data for these two dry porous materials are compared with measured hydrostats, the similarities in behavior seem to disappear. In Mt. Helen tuff, the stress-compression Hugoniot begins above the hydrostat, then crosses into the region below it at a stress of about 1.2 GPa. However, in Indiana limestone, the Hugoniot remains above the hydrostat, and at a stress separation of about twice that expected from a combination of quasi-static strength and pressure measurements. (This approximates the separation expected if strength were increased about 10% per decade of strain rate). Considering the Mt. Helen tuff data alone, the crossing of the two sets of data could result from shear enhanced compaction in the dynamic experiments. Such interpretation, however, raises several questions. For example, why don't the curves (Hugoniot and hydrostat) cross at a lower stress, i.e., nearer to the matrix failure stress? Why don't the two curves cross in dry Indiana limestone? The answers to these questions are closely related to the time-dependent processes that are occurring in these complex rock types. Obviously, pore collapse in highly porous rock is a sensitive function of time because it involves movement of mass over relatively large distances through complex arrangements of pores. Limestone introduces an additional complexity because it can undergo polymorphism (another time-dependent process) in the stress range of this experimental investigation.

Finally, we ask why Blair dolomite, polycrystalline NaCl, dry Mt. Helen tuff, and dry Indiana limestone all propagate nonsteady waves that are partially attributed to pore collapse phenomena, while Westerly granite and nugget sandstone do not appear to do so. Actually all these rocks propagate nonsteady waves. In the silicate rocks, however, the complete pore collapse process is spread over a larger stress regime. Coupled with the small initial porosity and the greater strength of the matrix, the Westerly granite and nugget sandstone are much more nearly elastic than the other rocks. The fact that nonsteady wave propagation is observed in these rocks clearly indicates that on-going processes require time durations equal to or greater than the rise times of the initial input shock wave to establish mechanical equilibrium.

Of the two saturated porous rocks, Mt. Helen tuff is far less complicated. Comparison of shock wave

and hydrostatic data suggests that this rock loses its strength under plane shock wave loading because of pore pressure buildup. However, at stress above 1.2 GPa, where water is converted to ice VI, the behavior of this saturated rock appears to become time dependent because the shock wave data show no indication of the volume compaction expected for such transformation. This suggests the possibility of a metastable form of water existing under dynamic conditions.

In saturated Indiana limestone, a comparison of shock wave and hydrostatic data suggests that time-dependent effects are dominating the response of this material under shock wave loading at stresses both below and above the liquid water-to-ice VI transition point.

Historically, constitutive models used in computer codes to predict dynamic response have been based primarily on experimental data generated under quasi-static conditions. Shock wave experiments have been used basically as tests against which constitutive models in the codes are checked by running problems that simulate a laboratory experiment. In some materials, quasi-static laboratory data (strain rates of 10^{-4} - 10^{-3} sec $^{-1}$) and shock wave data (strain rates of 10^5 - 10^7 sec $^{-1}$) agree. In these cases, the quasi-static data can confidently be used in developing models to describe dynamic phenomena that involve strain rates of 10^{-4} - 10^5 sec $^{-1}$. However, some materials exhibit apparent rate phenomena that are not present under quasi-static conditions. Dry porosity and phase transformation apparently contribute to this rate phenomena. Therefore, extreme caution should be exercised in using data obtained quasi-statically in constitutive theory to calculate the behavior of rocks under dynamic conditions whenever any significant amount of dry porosity or a phase transformation is present. Similarly, if dynamic data were used to develop a constitutive model and calculations of static problems were performed and then tested against the results of quasi-static experiments, some discrepancies could be expected. Therefore, caution should be exercised where data obtained dynamically in constitutive theory is used in the calculation of the behavior of rocks under static conditions. In actual application, this latter case has not arisen.

CONCLUSIONS

Based on our analysis of the data obtained from plane shock wave experiments we can draw the following conclusions:

1. Of the eight rocks studied, only Westerly granite and nugget sandstone appear to propagate steady waves, i.e., are time independent in their behavior.
2. Westerly granite and nugget sandstone show evidence of a change of character at 2-3 GPa and 1.5 GPa, respectively. The stress at which this change of character is observed has been interpreted as a threshold stress for the collapse of pores.
3. In Blair dolomite and polycrystalline NaCl
- time-dependent behavior is observed and attributed to the time required to close the small amount of available porosity.
4. The highly porous dry rocks, Mt. Helen tuff and Indiana limestone, are strongly time-dependent materials. This characteristic would seem to preclude the use of quasi-static data in constitutive models that are used to calculate dynamic processes.
5. The water in water-saturated rocks shows no indication of transformation to the higher pressure ice phases. This suggests the possibility of a metastable form of water existing under dynamic conditions.

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