Heating effects of the matrix of experimentally shocked Murchison CM chondrite: Comparison with micrometeorites

Naotaka TOMIOKA1, Kazushige TOMEOKA1, Keiko NAKAMURA-MESSENGER2, and Toshimori SEKINE3

1Department of Earth and Planetary Sciences, Faculty of Science, Kobe University, Kobe 657-8501, Japan
2NASA Johnson Space Center, Houston, Texas, 77508, USA
3National Institute for Materials Science, Tsukuba 305-0044, Japan
*Corresponding author. E-mail: nao@kobe-u.ac.jp
(Received 30 January 2006; revision accepted 02 November 2006)

Abstract–Micrometeorites have been significantly altered or melted by heating, which has been mainly ascribed to aerodynamic drag during atmospheric entry. However, if a major fraction of micrometeorites are produced by impacts on porous asteroids, they may have experienced shock heating before contact with the Earth’s atmosphere (Tomeoka et al. 2003). A transmission electron microscope (TEM) study of the matrix of Murchison CM chondrite experimentally shocked at pressures of 10–49 GPa shows that its mineralogy and texture change dramatically, mainly due to shock heating, with the progressive shock pressures. Tochilinite is completely decomposed to an amorphous material at 10 GPa. Fe-Mg serpentine is partially decomposed and decreases in amount with increasing pressure from 10 to 30 GPa and is completely decomposed at 36 GPa. At 49 GPa, the matrix is extensively melted and consists mostly of aggregates of equigranular grains of Fe-rich olivine and less abundant low-Ca pyroxene embedded in Si-rich glass. The mineralogy and texture of the shocked samples are similar to those of some types of micrometeorites. In particular, the samples shocked at 10 and 21 GPa are similar to the phyllosilicate (serpentine)-rich micrometeorites, and the sample shocked at 49 GPa is similar to the olivine-rich micrometeorites. The shock heating effects also resemble the effects of pulse-heating experiments on the CI and CM chondrite matrices that were conducted to simulate atmospheric entry heating. We suggest that micrometeorites derived from porous asteroids are likely to go through both shock and atmospheric-entry heating processes.

INTRODUCTION

Dust particles in the size range of 20–400 μm are accreted by the Earth with an estimated flux of ~30,000 tons per year (Love and Brownlee 1993; Taylor et al. 1998). Those particles collected on the Earth’s surface are termed micrometeorites. Previous studies (e.g., Kurat et al. 1994; Genge et al. 1997; Engrand and Maurette 1998; Nakamura et al. 2001) showed that most unmelted micrometeorites are similar in chemistry and mineralogy to the matrices of CI and CM carbonaceous chondrites, which are porous aggregates composed mainly of hydrous phyllosilicates. However, micrometeorites have been significantly altered by heating, and thus those containing phyllosilicates are rare. The effects of heating have previously been primarily explained by aerodynamic drag during atmospheric entry. The degree of heating has been thought to be related to the entry velocities and angles of particles, and thus used to infer their orbits and sources, that is, whether they were derived from asteroids or comets (e.g., Flynn 1989; Love and Brownlee 1991; Greshake et al. 1998; Toppani et al. 2001).

Tomeoka et al. (2003) recently performed shock-recovery experiments of hydrated and anhydrous carbonaceous chondrites (Murchison CM and Allende CV chondrites, respectively), and found that the application of shock to the hydrated chondrite reduces it to minute particles through a much broader range of pressures than for the anhydrous chondrite. In porous material such as carbonaceous chondrites, the amount of postshock heat is much larger than in nonporous material (e.g., Bauer 1979; Stöffler et al. 1991). The heating must cause dehydration of hydrous minerals, evaporation of H2O, and increase in gas volume, and as a result, generates great expansive force on pressure release. The experimental results therefore suggest that hydrated asteroids will produce dust particles during collisions at a much higher rate than anhydrous asteroids. If
this model is valid, an important possibility arises: the effects of heating in micrometeorites could also be ascribed to impact-induced shock on their parent bodies.

To test this hypothesis, we undertook a detailed mineralogical investigation of the matrix of the experimentally shocked Murchison CM chondrite (Tomeoka et al. 1999) using a transmission electron microscope (TEM). Murchison can be regarded to be an analogue of the hydrated micrometeorite parent bodies. In the previous study (Tomeoka et al. 1999), the Murchison samples were experimentally shocked at peak pressures from 4 to 49 GPa, and the recovered samples were studied using an optical microscope, a scanning electron microscope (SEM), and an electron probe micro-analyzer. However, details of mineralogy and texture inside the matrix remain unknown. Our goals were to determine how the matrix of Murchison changes by the progressive shock pressures and to compare the heating effects of the shocked samples with those of micrometeorites and also of the pulse-heating experiments on the CI and CM matrices that were conducted to simulate atmospheric entry heating. Thereby, we intended to determine the heating history of micrometeorites.

**MATERIALS AND METHODS**

The Murchison samples studied here were recovered from the previous shock experiments (Tomeoka et al. 1999) that were performed by using a single-stage propellant gun at the National Institute for Materials Science (formerly NIRIM). The Murchison samples were shocked in nine experiments at peak pressures of 4, 10, 21, 26, 28, 30, 34, 36, and 49 GPa. The target samples were disks of Murchison that are 5.7–10.0 mm in diameter and 1.0–3.5 mm in thickness, encapsulated in stainless steel (SUS304) containers. The velocities of projectiles range from 0.50 to 1.80 km/s. The experiments at peak pressures of 4, 10, 21, 26, 28, 30, 34, 36, and 49 GPa were subjected to the present TEM investigation.

Small pieces of matrix (~50 μm in size) were removed from polished sections of the shocked Murchison samples, embedded in epoxy, and processed to thin foils of 55–70 nm in thickness using an ultramicrotome. The thin foils were mounted on Cu grids, each of which was covered with a microgrid and a formvar/carbon film. They were studied by using a TEM (JEOL JEM-2010) operated at an accelerating voltage of 200 kV and equipped with an energy dispersive X-ray spectrometer. For quantitative TEM analyses, k-factors for the major elements were determined using standards of San Carlos olivine, San Carlos clinopyroxene, and K-feldspar from the Koryu mine, Hokkaido, Japan.

In this paper, the pressure value designated for each recovered sample is that of the equilibrium peak shock pressure calculated from the measured velocity, which follows most of the past shock-recovery experiments. However, it should be noted that shock-loading paths followed by a sample in the stainless steel container (present case) differ considerably from paths that are expected to be followed by natural targets, which are, in our case, surface materials on the meteorite parent bodies. Therefore, even if the pressure values for the experimental case and the natural case are the same, there may be a considerable difference in internal energy (thus, postshock heat) produced in the target between the two cases. This matter is discussed in detail in Tomeoka et al. (1999).

**RESULTS**

**Matrix of Natural Murchison (Before Shock Loading)**

Murchison contains chondrules of 100–500 μm in diameter embedded in a high-volume fraction (~64%) (McSween 1979) of matrix. The porosity of bulk Murchison has been reported to be 23% (Corrigan et al. 1997); the porosity of matrix has not been determined, but must be significantly higher than this value. The matrix consists mainly of fine-grained hydrous minerals, mostly Fe-Mg serpentine and tochilinite (Barber 1981; Tomeoka and Buseck 1985), that account for most of the water (~12 wt%) (Fuchs et al. 1973) in this meteorite. Fe-Mg serpentine shows a wide range of Fe/Mg ratios, from magnesian cromstedite [(Fe4Mg2)(Fe3+Si3)O10(OH)8] to ferroan serpentine [(Fe3Mg5)Si4O10(OH)8], and contains small amounts of Al (Tomeoka et al. 1989). Tochilinite is composed of interstratified Fe sulfide sheets and Fe hydroxide sheets, having an approximate chemical formula of 6FeS · 5Fe(OH)2. Tochilinite and cromstedite commonly form a unit cell-scale alternating intergrowth. Minor constituents in matrix are magnetite, troilite, pentlandite, chromite, calcite, and organic compounds.

**Matrix Shocked at 10 GPa**

The texture and mineralogy of the matrix have changed considerably from those in unshocked Murchison. The matrix consists mainly of an intermixture of an amorphous material and Fe-Mg serpentine. Tochilinite has completely decomposed. Similarly, tochilinite was not observed in any of the samples shocked at higher pressures described below. The amorphous material commonly has a fibrous texture similar to serpentine (Fig. 1a). EDS analyses reveal that the amorphous material contains major amounts of Si, O, Fe, and Mg and minor amounts of Al, S, and Ca (Fig. 1b); among these elements, Fe, Mg, and S are highly variable in contents. From these results, the amorphous material can be interpreted as a mixture of decomposed products of tochilinite and Fe-Mg serpentine. We hereafter call this amorphous material Si-rich glass. Fe-Mg serpentine occurs mainly as two types of grains. One occurs as relatively coarse grains 1–2 μm in size...
Heating effects of the matrix of experimentally shocked Murchison CM chondrite

(Fig. 1c) like those in natural Murchison (e.g., Fig. 2 in Barber [1981]; Fig. 2 in Tomeoka et al. [1989]), and the other occurs as fine anhedral grains 20–300 nm in size (Fig. 1d). Selected-area electron diffraction (SAED) patterns from both types of grains show a distinct reflection at ~0.7 nm (Figs. 1c and 1d) that corresponds to a (001) interlayer spacing of serpentine. Finer-grained serpentine (<20 nm), which is abundant in matrix of natural Murchison (e.g., Fig. 6b in Barber [1981]), is absent. Trace amounts of olivine occur as grains 1–2 μm in size. Magnetite and Fe-Ni sulfide are rare.

Matrix Shocked at 21 GPa

The matrix also consists mainly of Si-rich glass and Fe-Mg serpentine. However, compared with the sample shocked at 10 GPa, the amount of Si-rich glass increases, whereas the amount of Fe-Mg serpentine decreases. Fe-Mg serpentine having an interlayer spacing of ~0.7 nm that occurs as relatively coarse grains 0.7–5.5 μm in size (Figs. 2a and 2b). Fine-grained serpentine (20–300 nm in size), which is abundant in the sample shocked at 10 GPa (see Fig. 1d), is absent. Olivine and pyroxene were not observed.

Si-rich glass contains numerous rounded Fe-rich grains 10–200 nm in diameter (Fig. 2c). Because of the small grain size and the occurrence within Si-rich glass, it was difficult to obtain high-quality electron diffraction and compositional data from individual Fe-rich grains. SAED patterns from areas with high concentration of Fe-rich grains show several powder diffraction rings (Fig. 2c). Two of the diffraction rings at d-spacings of 0.25 and 0.15 nm can be interpreted as corresponding to the strongest (311) and second strongest (440) reflections of magnetite, respectively. However, the rest of the rings cannot be identified by any possible Fe-rich minerals. Broad-beam (0.5 μm in diameter) EDS analyses...
Fig. 2. a) A TEM image of the matrix of Murchison shocked at 21 GPa, showing a coarse grain of Fe-Mg serpentine (Serp) embedded in Si-rich glass (Glass). b) A high-resolution TEM image of the serpentine grain showing stacking of layers with an interlayer spacing of ~0.7 nm. c) Numerous Fe-rich grains (dark) embedded in Si-rich glass (grayish background). In the inset is an SAED pattern from the area of this image. d) and e) EDS spectra of Si-rich glass with Fe-rich grains obtained from two different locations, using a defocused (0.5 μm in diameter) electron beam. Note that S peaks show a large difference in intensity between the two spectra. The Cu peaks are from TEM grids.
Heating effects of the matrix of experimentally shocked Murchison CM chondrite

indicate that the Fe-rich grains show highly variable S contents (Figs. 2d and 2e), suggesting that part of the Fe-rich grains are Fe-sulfide. The Si-rich glass with magnetite grains was probably formed by decomposition of Fe-Mg serpentine, whereas the Si-rich glass with magnetite and Fe-sulfide grains was formed by decomposition of Fe-Mg serpentine and tochilinite.

Matrix Shocked at 30 GPa

The matrix mineralogy differs considerably from that in the samples shocked at 10 and 21 GPa. The matrix contains abundant Mg-rich olivine grains (Fo_{98-99}, 0.2–1.0 μm in size) embedded in Si-rich glass (Fig. 3a). Minor amounts of low-Ca pyroxene (En_{92-98}) also occur as grains 0.2–1.0 μm in size. The olivine and low-Ca pyroxene grains are commonly anhedral to angular in shape. Fe-Mg serpentine occurs in very minor abundance as relatively coarse grains (0.1–0.5 μm) (Fig. 3b).

Si-rich glass contains a higher abundance of small, rounded Fe-rich grains (10–200 nm in diameter) than in the sample shocked at 21 GPa. SAED patterns from areas with high concentration of Fe-rich grains show six distinct powder diffraction rings at d-spacings of 0.30, 0.25, 0.21, 0.17, 0.16, and 0.15 nm (Fig. 3c). These diffraction rings can be ascribed to the six strong reflections (220), (311), (400), (422), (511), and (440) of magnetite, respectively. Part of the diffraction rings can also be ascribed to some strong reflections of pyrrhotite and troilite (see Fig. 3c). Because of the presence of significant amounts of S, part of the Fe-rich grains are likely to be either pyrrhotite or troilite.
Although the Si-rich glass shows highly variable Fe and Mg contents in the individual samples, there is an apparent systematic tendency that Mg contents increase with increasing shock pressure; an EDS spectrum of a relatively Mg-rich portion of Si-rich glass in the sample shocked at 30 GPa is shown in Fig. 3d (compare with the spectrum shown in Fig. 1b). This is probably explained by increasing amounts of Fe-Mg serpentine (a major Mg source in the matrix) decomposing to amorphous material with increasing shock pressure, whereas tochilinite (a major Fe source) was completely decomposed at 10 GPa.

**Matrix Shocked at 36 GPa**

Si-rich glass with numerous small, rounded Fe-rich grains (10–200 nm in diameter) is the dominant phase in the matrix (Fig. 4). The Si-rich glass also commonly contains numerous vesicles (<0.5 μm in diameter), which suggests that the vaporization of H₂O from hydrous minerals took place by shock-heating (Fig. 4). Serpentine has completely decomposed. The Fe-rich grains show almost the same SAED patterns as those in the sample shocked at 30 GPa, indicating that they are mainly magnetite and Fe sulfide. Olivine occurs in minor amounts as grains 0.5–1.5 μm in size. Trace amounts of low-Ca pyroxene also occur as grains 1.0–2.0 μm in size. Both olivine and low-Ca pyroxene are commonly anhedral to angular in shape and slightly higher in Fe contents (Fo 90–98 and En 72–97, respectively) than those in the sample shocked at 30 GPa.

**Matrix Shocked at 49 GPa**

The matrix is extensively melted and the matrix mineralogy and texture have drastically changed compared with the samples shocked at lower pressures. The dominant phase is olivine, which occurs as subrounded to euhedral grains 0.3–2.5 μm in size (Figs. 5a–c). Low-Ca pyroxene also occurs in minor amounts as subrounded to euhedral grains 0.2–0.5 μm in size (Fig. 5a). Olivine and low-Ca pyroxene grains form aggregates, in which Si-rich glass fills interspaces between them (Figs. 5a and 5b). Both olivine and low-Ca pyroxene are distinctly more Fe-rich and have wider ranges of Mg/(Fe + Mg) ratio (Fo 40–69 and En 22–77) than those in the samples described above; olivine compositions are compared in Fig. 6. Olivine grains commonly contain small Fe-Cr-rich grains (<0.1 μm in diameter) (Fig. 5c), which may have been derived from chromite in natural Murchison or from the stainless steel container.

**DISCUSSION**

**Effects of Shock Heating on Mineralogy and Texture**

Our study has revealed that the mineralogy and texture of the Murchison matrix changed dramatically with increasing shock pressure. The major minerals in the individual samples are listed in Table 1, and the changes are summarized as a function of shock pressure in Fig. 7.

**Serpentine and Tochilinite**

The matrix of natural Murchison consists mainly of Fe-Mg serpentine and tochilinite. In the sample shocked at 10 GPa, tochilinite has been completely decomposed to an amorphous material. Serpentine has also been partially decomposed and decreases in amount with increasing pressure from 10 to 30 GPa, and has been completely decomposed at 36 GPa. These changes indicate that the threshold of complete decomposition of serpentine exists between 30 and 36 GPa. This is consistent with the results of shock experiments of Murchison done by Akai and Sekine (1994), who reported that nearly complete decomposition of serpentine occurs at 32 GPa.

Previous heating experiments showed that tochilinite decomposes at 245 °C (Fuchs et al. 1973), whereas serpentine decomposes at 500–600 °C (e.g., Brindley and Zussman 1957; Akai 1992). In order to compare the effects of shock heating with those of normal (long-duration) heating, we calculated shock temperatures (temperatures at peak shock pressures) of the present shock experiments using the
Fig. 5. a) A TEM image of Murchison matrix shocked at 49 GPa, showing that it consists mostly of aggregates of equigranular grains of Fe-rich olivine (Ol) and less abundant low-Ca pyroxene (Px) with interstitial Si-rich glass (Glass). In the inset is an SAED pattern from Si-rich glass exhibiting no particular diffraction spots. b) Euhedral grains of olivine. In the inset is an SAED pattern from an olivine grain. c) Olivine grains containing small Fe-Cr-rich grains (indicated by arrows). d) An EDS spectrum of interstitial Si-rich glass without Fe-rich grains showing an especially high Ca peak. e) Numerous vesicles (indicated by arrows) in Si-rich glass. Large dark circles in the background are holes of a TEM microgrid. Dark grains are magnetite.
computer code AUTODYN 2D (Table 2). The details of our calculations are shown in the Appendix. The calculated shock temperature at 10 GPa (200–330 °C) is nearly consistent with the decomposition temperature of tochilinite (245 °C). However, the calculated shock temperatures at 30 GPa (830–1010 °C) and 36 GPa (1170–1200 °C) are much higher than the decomposition temperature of serpentine (500–600 °C). These results indicate that the decomposition degree of individual minerals is not simply dependent on shock temperature and suggest that the kinetic processes cannot be neglected when we compare the effects of shock heating and normal heating.

**Olivine and Pyroxene**

In the Murchison matrix shocked at 10 and 21 GPa, olivine is nearly absent, like in the unshocked Murchison sample. However, at 30 and 36 GPa, olivine occurs in significant amounts as anhedral-to-angular grains. The olivine grains at 30 and 36 GPa are poor in Fe and homogeneous in composition (Figs. 6b and 6c), resembling those in most chondrules in Murchison (Fuchs et al. 1973). It has been known that olivine that is formed by thermal transformation of Fe-Mg phyllosilicate is much higher in Fe content (e.g., Greshake et al. 1998) (also see Fig. 6e). As indicated by SEM observations (Tomeoka et al. 1999), at 30 GPa, narrow fractures form at high density throughout the Murchison matrix, and all coarse-grained olivines and pyroxenes in chondrules are densely fractured with subgrains 1 to 5 μm in size. These changes probably resulted from extensive comminution and explosive expansion. Therefore, we interpret that the olivine and pyroxene grains in the sample shocked at 30 and 36 GPa are fragments of coarse-grained olivines and pyroxenes in chondrules.

At 49 GPa, the matrix is extensively melted and consists mostly of aggregates of equigranular, subrounded to euhedral grains of olivine and low-Ca pyroxene with interstitial Si-rich glass. Both olivine and pyroxene are much higher in Fe contents and have wider ranges of Mg/(Mg + Fe) ratios than those in the samples shocked at 30 and 36 GPa (Fig. 6). These characteristics suggest that the olivine and pyroxene in this sample have a formation process different from those in the samples shocked at the lower pressures. We presume that most of the Fe-poor olivine and pyroxene grains are fragments derived from chondrules, whereas the Fe-rich olivine and pyroxene grains were crystallized from the partially melted matrix material.

**Comparison with Pulse-Heated Carbonaceous Chondrite Matrices**

Previous workers have conducted pulse-heating experiments of CI and CM chondrite matrices in order to

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<th>Pressure (GPa)</th>
<th>Major minerals</th>
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<tr>
<td>10</td>
<td>Si-rich glass</td>
</tr>
<tr>
<td></td>
<td>Serpentine</td>
</tr>
<tr>
<td>21</td>
<td>Si-rich glass</td>
</tr>
<tr>
<td></td>
<td>Magnetite</td>
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<tr>
<td></td>
<td>Serpentine</td>
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<tr>
<td>30</td>
<td>Si-rich glass</td>
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<tr>
<td></td>
<td>Magnetite, Fe-sulfide</td>
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<tr>
<td></td>
<td>Olivine (Fo98.99)</td>
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<td></td>
<td>Serpentine</td>
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<tr>
<td></td>
<td>Low-Ca pyroxene (En92.98)</td>
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<tr>
<td>36</td>
<td>Si-rich glass</td>
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<tr>
<td></td>
<td>Magnetite, Fe-sulfide</td>
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<tr>
<td></td>
<td>Olivine (Fo90.98)</td>
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<tr>
<td></td>
<td>Low-Ca pyroxene (En72.97)</td>
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<tr>
<td>49</td>
<td>Olivine (Fo40.60)</td>
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<tr>
<td></td>
<td>Low-Ca pyroxene (En22.77)</td>
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<td></td>
<td>Si-rich glass</td>
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<td>Magnetite</td>
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![Figure 6](image-url)
simulate atmospheric entry heating of micrometeorites (e.g.,
Greshake et al. 1998; Toppani et al. 2001). Because the
duration of shock heating is very short, their results are
helpful in deducing the peak temperature of the individual
shock experiments. In the following, we compare the results
of our study with those of pulse-heating experiments done by
Greshake et al. (1998) and Toppani et al. (2001). However, it
should be noted that the pulse-heating experiments were
mostly conducted at temperatures of 800–1500 °C, under
which most hydrous phases are dehydrated and partially or
completely melted. Thus, the effects of heating at temperature
lower than 800 °C are still poorly known. It should also be
noted that the durations of the pulse-heating experiments
(mostly 20–120 s) are considered to be still significantly
longer than those of peak temperature of the shock
experiments. Therefore, the temperature of the pulse-heating
experiments implies minimum peak temperature for the
corresponding shock experiments.

Using a TEM, Greshake et al. (1998) studied the matrix
of the Orgueil CI chondrite heated at 700 °C for 20 s and
found that the phyllosilicates are completely decomposed to
an amorphous material. This corresponds to the Murchison
matrix shocked at 36 GPa. Thus, all the Murchison samples
shocked at <36 GPa may have experienced a lower degree of
heating than this pulse-heating experiment. On the other
hand, by heating at 800 °C for 20 s, the phyllosilicates begin
to transform to Fe-rich olivine and pyroxene, and by heating
at 1100–1200 °C for 20 s, those silicates form equigranular
aggregates with grain size of ~300 nm, in which minor
amounts of Si-rich glass fill their interspaces. The
composition of olivine in the sample heated at 1100 °C for
20 s ranges from 35 to 75 mol% Fo. These characteristics
closely resemble those of the sample shocked at 49 GPa
(Figs. 5a and 5b; compare with Figs. 7b and 7c in Greshake
et al. [1998]; also see Fig. 6 for comparison of compositions).

Toppani et al. (2001) reported that heating of Murchison
fragments at 500–1000 °C for 20 s shows no significant
textural changes; however, their observations are based on an
SEM, and so detailed mineralogy inside the fragments is
unavailable. On the other hand, the Murchison fragment
heated at 1350 °C for 20 s is extensively melted and has
numerous olivine grains embedded in a glassy mesostasis.
The vesicles in the melts are large (up to ~100 µm in
diameter) and ovoid. The mineralogy and texture are similar
to those of the sample shocked at 49 GPa (compare Fig. 6b in
Toppani et al. [2001] with Figs. 12a and 12b in Tomeoka et al.
[1999]). Because the duration of shock heating is much
shorter than 20 s, the similarity suggests that the peak
temperature reached by the sample shocked at 49 GPa is
much higher than 1350 °C. The calculated shock temperature
at 49 GPa (~1930 °C) (Table 2) is consistent with this
estimate.

Comparison with Micrometeorites

SEM Observations

We compare the characteristics of the shocked
Murchison matrix observed by an SEM (Tomeoka et al. 1999)
with those of micrometeorites reported by previous
workers. Based on SEM observations, previous workers (e.g.,
Kurat et al. 1994; Engrand and Maurette 1998) classified
micrometeorites into four main types: 1) fine-grained
dominated by partially or completely dehydrated
phyllosilicate; relative abundance = 32%) (Engrand and
Maurette 1998); 2) crystalline (dominated by coarse-grained
crystals; 8%); 3) scoriaceous (partially melted, with
numerous vesicles; 40%); and (4) cosmic spherules
(completely melted; 20%).

The matrix of Murchison shocked at 10 and 21 GPa has a
fine-grained texture with a smooth surface and exhibits no
distinct differences in appearance on SEM images from the
unshocked Murchison matrix except having a more
compacted texture and a higher abundance of fractures. Thus,
they are similar to the fine-grained (phyllosilicate-rich)
micrometeorites (e.g., Fig. 1b in Kurat et al. [1994]; Figs. 6c
and 6d in Nakamura et al. [2001]; Figs. 2a–d in Noguchi et al.
Table 2. Calculated shock temperatures at individual pressures of the shock experiments of Murchison.

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<thead>
<tr>
<th>Pressure (GPa)</th>
<th>Shock temperature (°C)</th>
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<tr>
<td>10</td>
<td>200–330</td>
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<tr>
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<td>580–930</td>
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<td>30</td>
<td>830–1010</td>
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<td>36</td>
<td>1170–1200</td>
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<tr>
<td>49</td>
<td>~1930</td>
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The sample shocked at 30 GPa has a high density of narrow fractures throughout the matrix (Fig. 10 in Tomeoka et al. [1999]). No micrometeorites corresponding to this sample have been reported. The sample shocked at 36 GPa has a rugged surface with abundant small fragments of olivine and pyroxene, and in many places, melts containing numerous vesicles of 1–2 μm in diameter (Fig. 11 in Tomeoka et al. [1999]). This sample resembles the vesicular, fine-grained micrometeorite reported by Toppani et al. (2001) (their Figs. 10c and 10d) that appears to be transitional from the fine-grained to the scoriaceous micrometeorites. The sample shocked at 49 GPa is composed largely of melts containing large, ovoid vesicles (<300 μm in diameter) (Fig. 12 in Tomeoka et al. [1999]), which corresponds to the scoriaceous micrometeorites or that transitional to the cosmic spherules (Figs. 3b–d in Kurat et al. [1994]; Figs. 9c and 9d in Toppani et al. [2001]).

TEM Observations

Previous TEM studies of unmelted, fine-grained micrometeorites (e.g., Nakamura et al. 2001; Noguchi and Nakamura 2000; Noguchi et al. 2002) revealed that they are typically a mixture of anhydrous minerals and a Si-rich amorphous material in various proportions. Major anhydrous minerals include olivine, low-Ca pyroxene, magnetite, and Fe-sulfide. Hydrous phyllosilicates occur, but in very minor abundances. Based on the relative abundances of the minerals, the micrometeorites can be classified into three types: olivine-rich (~75%), pyroxene-rich (~20%), and phyllosilicate-rich (~5%) (Nakamura et al. 2001).

In olivine-rich micrometeorites, olivine typically forms aggregates of equigranular, rounded to subhedral grains with diameters of 20–100 nm (e.g., Fig. 7d in Greshake et al. [1998]; Fig. 7a in Nakamura et al. [2001]). An Si-rich amorphous material, which is commonly intermixed with fine-grained magnetite and Fe sulfide, fills interspaces between the olivine grains. The texture and mineralogy are interpreted as having resulted from atmospheric entry heating of the micrometeorites that had been originally hydrous phyllosilicate-rich (e.g., Greshake et al. 1998; Nakamura et al. 2001). However, our study revealed that their texture and mineralogy are very similar to those of the Murchison matrix shocked at 49 GPa (Fig. 5). The range of Mg/(Mg + Fe) ratios of olivines in the micrometeorites (Nakamura et al. 2001) is also comparable to that of the olivines in the Murchison matrix shocked at 49 GPa (Fig. 6). These findings imply that the olivine-rich micrometeorites could also have been produced by shock heating that occurred on micrometeorite parent bodies.

None of the Murchison samples recovered from the present shock experiments correspond to the pyroxene-rich micrometeorites. This result is reasonable, because the phyllosilicate in CM chondrite matrix is serpentine and the thermal transformation product of serpentine is mostly olivine (Akai 1992). However, in the matrices of CI and Tagish Lake chondrites, major amounts of smectite (saponite) occur in addition to serpentine (Tomeoka and Buseck 1988; Zolensky et al. 2002). Saponite has a much smaller (Mg + Fe)/Si ratio than serpentine (0.75 versus 1.5 in atomic ratio), and thus tends to thermally transform to pyroxene rather than olivine (e.g., Akai 1992). The pulse-heating experiments of the matrix of Orgueil CI chondrite showed that both olivine and low-Ca pyroxene are indeed formed by thermal transformation of serpentine and saponite (Greshake et al. 1998). These suggest that pyroxene-rich micrometeorites were formed by heating of smectite-rich type micrometeorites or meteorites.

Phyllosilicate-rich micrometeorites can be further divided into two types, serpentine-rich and smectite-rich (Nakamura et al. 2001). Both types also contain olivine, magnetite, and Fe-sulfide in various proportions. Among the two types, serpentine-rich micrometeorites contain Fe-rich serpentine that is similar to the cronstedite in CM chondrites. An important difference is, however, that tochilinite coexisting with serpentine in CM chondrites has not been found in any of the micrometeorites. This raises the possibility that there is no direct relationship between the serpentine-rich micrometeorites and CM chondrites (e.g., Nakamura et al. 2001). However, our study showed that tochilinite is completely decomposed to an amorphous material by shock heating even at a relatively mild shock pressure (10 GPa), while most coarse-grained serpentine survives the heating. This suggests that tochilinite preferentially breaks down at the time of dust production by impacts on CM parent bodies. Thus, the serpentine-rich micrometeorites may be CM chondrite fragments that were shocked at pressures lower than 36 GPa.

A characteristic phase that has recently been found from many micrometeorites is magnesiowüstite (Nakamura et al. 2001; Noguchi et al. 2002). This phase is regarded to have been formed by decomposition of magnesian siderite (Fe-Mg carbonate) during atmospheric entry heating of micrometeorites (Noguchi et al. 2002). In the present study, no magnesiowüstite has been found from any of the shocked Murchison samples. This can be explained by the fact that carbonate in Murchison is exclusively calcite (Ca carbonate). In the shocked Murchison samples studied here, neither calcite nor its decomposed products were observed. However,
the Si-rich glass in the Murchison sample shocked at 49 GPa showed a much higher content of Ca than any of the samples shocked at lower pressures (Fig. 5d). The Ca in the Si-rich glass may have been derived from calcite that was decarbonated by shock heating.

Another characteristic product of heating in micrometeorites is magnetite, which occurs typically as thin shells surrounding micrometeorites as well as recrystallized grains within granular aggregates of anhydrous silicates (e.g., Toppani et al. 2001). The magnetite shell is generally thought to form through mineral reactions induced by heating from outside of a micrometeorite to inside during atmospheric entry. In our shock experiments of Murchison, magnetite is produced as small particles embedded in Si-rich glass at 21, 30, and 36 GPa, but its total amounts are very small. The formation of magnetite in the shock experiments was presumably restrained due to insufficient availability of oxygen within the sealed sample container and insufficient duration of heating. Shock heating occurs heterogeneously in the shocked material and does not cause such thermal gradient from core to rim in individual ejected particles, which is responsible for the formation of magnetite shells. Therefore, the magnetite shell of micrometeorites can be regarded as rare mineralogical evidence indicating its formation to have resulted from atmospheric entry heating.

CONCLUSIONS

The results of our study revealed that the Murchison matrix shocked at pressures from 10 to 49 GPa shows various mineralogical and textural effects of shock heating, and that they bear a close resemblance to the effects of heating observed in the unmelted and partially melted micrometeorites. In particular, the Murchison matrix shocked at relatively low pressures (10 and 21 GPa) is similar to the phyllosilicate- (serpentine) rich micrometeorites, and the matrix shocked at 49 GPa is similar to the olivine-rich micrometeorites. Our study further revealed that the effects of shock heating on the Murchison matrix resemble the effects of pulse-heating experiments on the CI and CM chondrite matrices that were conducted to simulate atmospheric entry heating.

The effects of heating on micrometeorites have previously been primarily ascribed to aerodynamic drag during atmospheric entry. However, most of the effects of heating in micrometeorites can also be explained by impact-induced shock on the micrometeorite parent bodies. We suggest that micrometeorites that are produced by impacts on porous asteroids go through both shock heating (on asteroids) and atmospheric-entry heating processes.

Acknowledgments—We thank Drs. A. Greshake and H. Nagahara for helpful reviews. ATEM analysis was performed at the Instrumentation Analysis Division, Kobe University. This work was supported by the Grant-in-Aid for Scientific Research (no. 15740316 to N. T., No. 16204042 to K. T., and no. 18204051 to T. S.) and The 21st Century COE Program of Origin and Evolution of Planetary Systems of the Ministry of Education, Culture, Sports, Science, and Technology.

Editorial Handling—Dr. Hiroko Nagahara

REFERENCES


### APPENDIX

**Shock Temperature Calculations**

Shock temperatures (temperatures at peak shock pressures) at individual pressures of the shock experiments of Murchison have been thermodynamically calculated using the computer code AUTODYN 2D (Century Dynamics Inc.). We used the Hugoniot for Murchison (Anderson and Ahrens 1998) and a heat capacity of 0.50 J/g/deg at constant volume that is typical of chondrites (Matsui and Osako 1979). The Grüneisen parameter was set to be 1.0 (Anderson and Ahrens 1998). Shock conditions were taken from Table 1 in Tomeoka et al. (1999). The variation of temperature at each shock pressure reflects differences in the sample thickness and the number and location of gauges that are set to give fine-scale temperature profiles within the sample. These calculated shock temperatures are regarded to be averages of bulk samples. However, it should be noted that the actual temperature rise in the samples is probably very heterogeneous on micron-to-submicron scales due to the heterogeneous distribution of pores and the fine-grained, intermixed nature of different minerals.