Pyroxene polymorphs in melt veins of the heavily shocked Sixiangkou L6 chondrite

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Abstract: Sixiangkou is a heavily shocked L6 chondrite that contains numerous shock-induced melt veins with varying widths. This paper focuses on pyroxene polymorphs in shock-induced melt veins in the meteorite and discusses their implication to the high pressure and temperature conditions during impact metamorphism. The major high-pressure polymorphs are majorite and majorite-pyrope solid solution. Diopside, akimotoite and jadeite were also observed in melt veins of the Sixiangkou chondrite. Diopside coexists with ringwoodite (+ akimotoite) in coarse-grained fragments of the melt veins. Two mechanisms were proposed for the lack of phase transformation of diopside: a) low local temperature and b) a very sluggish transformation rate. Majorite and majorite-pyrope occur as coarse-grained polycrystalline and fine-grained solid solution. They were formed through solid state transformation and crystallized from a melt under pressures, respectively. Akimotoite coexists with low-Ca pyroxene in coarse-grained fragments of melt veins. They have nearly identical chemical compositions to the host low-Ca pyroxene, implying a solid state transformation mechanism. Jadeite appeared as fine-grained and was originated through retrograde transformation of lingu-nite under moderate post-shock pressure and high temperature conditions. The coexistence of different high-pressure pyroxene polymorphs reflects an inhomogeneous distribution of temperature in the shock-induced melt veins and a kinetic effect of phase transformations of minerals during shock metamorphism.

Key-words: Sixiangkou L6 chondrite, heavily shocked, melt veins, akimotoite, phase transformation.

1. Introduction

Pyroxene is one of the most important components in stony meteorites and the upper mantle of the Earth. During asteroidal impact events, pyroxene would transform into various high-pressure polymorphs (e.g., majorite, akimotoite, perovskite) which have been observed in meteorites (Smith & Mason, 1970; Sharp et al., 1997; Tomioka & Fujino, 1997; Tomioka & Kimura, 2003). These high-pressure phases potentially provide information about temperature and pressure conditions prevailing in the region of Earth’s mantle. Static high-pressure experiments (Liu, 1976; Presnall, 1995) demonstrated that low-Ca pyroxene could transform to akimotoite (17.5-27.5 GPa and 600-2050°C), majorite (16-22.5 GPa and 1600-2500°C), and perovskite (> 23 GPa and ~ 2000°C). Under high pressures, high-Ca pyroxene would decompose into assemblages of high-pressure phases, such as MgSiO3-garnet (majorite) + CaSiO3-perovskite; MgSiO3-ilmenite (akimotoite) + CaSiO3-perovskite; MgSiO3-perovskite + CaSiO3-perovskite; and possible CaMgSi2O6-perovskite (Oguri et al., 1997; Irifune et al., 2000; Asahara et al., 2005). In nature samples, majorite, akimotoite, and vitrified perovskite were observed in the Tenham L6 chondrite (Tomioka & Fujino, 1997, 1999; Xie et al., 2006), and majorite and vitrified perovskite were reported in the Suizhou L6 chondrite (Xie et al., 2001; Chen et al., 2004b). Akimotoite and vitrified MgSiO3 perovskite have also been reported in the Acfer 040 L5-6 chondrite. Recently, Tomioka & Kimura (2003) reported the existence of both Ca-rich (~ 12 wt% CaO) and Ca-poor majorite in the Yamato 751000 H6 chondrite. Ca-rich majorite was thought to be a breakdown product of diopside. These chondrites belong to a high-degree stage (S5-6) of shock classification. However, in another heavily shocked L6 chondrite, the Sixiangkou chondrite, only garnet-structured pyroxene was reported (Chen et al., 1996). Here, we revisited Sixiangkou with an electron microprobe and a micro-Raman spectroscopy and observed akimotoite, diopside and jadeite in shock-induced melt veins. This paper aims to discuss their pressure and temperature stabilities and significance to shock metamorphism.
2. Sample and analytical methods

The Sixiangkou meteorite fell on August 15, 1989, in Sixiangkou Town of Taizhou city, Jiangsu Province of China. In total, 375 g sample was collected and the main mass was preserved in the Purple Mountain Observatory, Chinese Academy of Sciences. Three thin sections were consecutively cut from a single fragment. These thin sections were first observed with an optical microscope to locate shock-induced veins. Then they were coated with carbon and were studied with a scanning electron microscopy (JSM-5610LV at Nanjing Normal University and Hitachi 3400N at Purple Mountain Observatory) and an electron microprobe (JEOL JXA-8800M at Nanjing University).

Mineral chemistry was determined with the JEOL JXA-8800M electron microprobe. Analyses were performed at a 15 kV acceleration voltage with a beam current of 20 nA, except that 10 nA was used for maskelynite and jadeite. The electron beam was focused to an area of less than 1 μm for most minerals, and a beam diameter of 10 μm was used for maskelynite. Natural and synthetic standards are used for quantitative analysis. All the data were reduced with the ZAF routine. The detection limits (ppm) of some minor elements are: Na (350), K (130), Al (135), Mn (190), Ti (250), Cr (170), and P (100).

Raman spectra of minerals were taken with a Renishaw RM2000 micro-Raman spectrometer equipped with a CCD detector at Nanjing University. The operating conditions were: excitation laser wavelength: 514 nm (Ar+ laser), laser energy: 5 mW, and spectral slit: 25 μm. The 50 × objective was used on a Leica DM/LM microscope. With this objective, the lateral spot-size of the laser beam was about 1 μm. Silicon (520 cm–1 Raman shift) was used as a standard.

3. Results

3.1. Petrography of the Sixiangkou chondrite

Petrographic textures and distribution of shock-induced melt veins are very similar in the three thin sections studied (Fig. 1). The shock-induced melt veins range in width from several μm to 4 mm. These melt veins align in two directions and cross with an angle of about 45°. Some melt pockets up to several hundred microns were also observed in the thin sections. The Sixiangkou chondrite was classified as an L6 chondrite. The presence of maskelynite and ringwoodite indicated that this chondrite was heavily shocked and was classified as shock stage S6 (Chen et al., 1996, 2002).

The host portion of the chondrite is mainly composed of olivine, low-Ca pyroxene, diopside, plagioclase (completely transformed to maskelynite), FeNi metal and troilite. Minor phases include chromite, whitlockite, and chlorapatite (Chen et al., 1995, 2002). Some relic chondrules with barred olivine and maskelynite could be recognized. Olivine, low-Ca pyroxene and diopside are euhedral and display undulose extinction. Low-Ca pyroxene has low contents of Al2O3 (< 0.05 wt%) and CaO (less than 1 wt%) with a molar Fe/(Fe+Mg) ratio of 0.23, similar to that (0.22) of Chen et al. (1996). Diopside has a slightly high Al2O3 content (~ 0.50 wt%) and a Fe/(Fe+Mg) ratio of 0.17. Most maskelynite grains have typical flow textures and fit interstitially between olivine and pyroxene. Olivine and pyroxene grains surrounding maskelynite show radiating cracks outward from maskelynite (the right side of Fig. 2).

3.2. Mineralogy of the shock-induced melt veins

The shock-induced melt veins contain two distinct lithologies: fine-grained matrix and coarse-grained frag-
ments. Some thick melt veins show round to dendritic occurrences of metal-troilite assemblages in the fine-grained matrix from the edge to the center (Fig. 3), similar to previous observations (e.g., Fig. 1 of Chen et al., 1996). The matrix of melt veins is composed of fine-grained majorite-pyrope (1-5 μm) and eutectic metal-troilite (from < 1 μm to 30 μm). Majorite-pyrope has two Raman bands at 656 and 927 cm\(^{-1}\). It has higher contents of Al\(_2\)O\(_3\) (4.10 wt%) and CaO (1.92 wt%) than low-Ca pyroxene in the host portion of Sixiangkou (Table 1). Magnesiowüstite (up to 5 μm) was reported as an important component in the fine-grained matrix (Chen et al., 1996). Some jadeite and lingunite grains look darker than the main components (majorite-pyrope) of the fine-grained matrix on backscattered electron images (Fig. 2 and 4). Jadeite has typical Raman bands at 1039, 695, 574, 524, and 375 cm\(^{-1}\), whereas lingunite at 977, 771, 627, 277, and 213 cm\(^{-1}\) (Fig. 5). Both Raman spectra of jadeite and lingunite contain bands of other high-pressure phases, e.g., majorite or ringwoodite.

Coarse-grained fragments in the shock-induced melt veins are commonly ovoid in shape and vary in size from several μm to 150 μm. They are mainly composed of olivine, low-Ca pyroxene, ringwoodite, majorite, diopside, chlorapatite, and chromite. Within the shock-induced melt veins, we also found some polymineralic fragments that contain closely contacted low-pressure and high-pressure phases. Here we report three typical coarse-grained

### Table 1. EMPA analyses of minerals (wt %) in the Sixiangkou chondrite.

<table>
<thead>
<tr>
<th></th>
<th>Ol</th>
<th>Pyx</th>
<th>Di-h</th>
<th>Msk</th>
<th>Ri</th>
<th>Mj</th>
<th>Mj-Py</th>
<th>Di-v</th>
<th>Ak</th>
<th>Ak(^a)</th>
<th>Ak(^b)</th>
<th>Jd</th>
</tr>
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<tbody>
<tr>
<td>Na(_2)O</td>
<td>–</td>
<td>0.011</td>
<td>8.56</td>
<td>–</td>
<td>0.04</td>
<td>0.26</td>
<td>0.12</td>
<td>0.11</td>
<td>0.67</td>
<td>0.42</td>
<td>1.63</td>
<td></td>
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<tr>
<td>TiO(_2)</td>
<td>–</td>
<td>0.04</td>
<td>0.23</td>
<td>–</td>
<td>0.11</td>
<td>0.10</td>
<td>0.24</td>
<td>0.09</td>
<td>0.17</td>
<td>–</td>
<td>–</td>
<td></td>
</tr>
<tr>
<td>MnO</td>
<td>0.24</td>
<td>0.28</td>
<td>0.15</td>
<td>–</td>
<td>0.20</td>
<td>0.26</td>
<td>0.19</td>
<td>0.13</td>
<td>0.52</td>
<td>0.28</td>
<td>–</td>
<td></td>
</tr>
<tr>
<td>MgO</td>
<td>37.94</td>
<td>28.43</td>
<td>16.01</td>
<td>0.01</td>
<td>37.58</td>
<td>28.48</td>
<td>26.47</td>
<td>16.28</td>
<td>27.17</td>
<td>28.39</td>
<td>35.34</td>
<td>0.11</td>
</tr>
<tr>
<td>K(_2)O</td>
<td>–</td>
<td>0.017</td>
<td>1.07</td>
<td>–</td>
<td>–</td>
<td>0.04</td>
<td>0.02</td>
<td>–</td>
<td>–</td>
<td>0.08</td>
<td>–</td>
<td></td>
</tr>
<tr>
<td>FeO</td>
<td>24.13</td>
<td>13.85</td>
<td>5.97</td>
<td>0.17</td>
<td>24.58</td>
<td>15.59</td>
<td>12.24</td>
<td>5.891</td>
<td>3.391</td>
<td>3.54</td>
<td>3.72</td>
<td>0.68</td>
</tr>
<tr>
<td>Al(_2)O(_3)</td>
<td>–</td>
<td>0.13</td>
<td>0.49</td>
<td>20.91</td>
<td>0.03</td>
<td>0.13</td>
<td>4.10</td>
<td>0.50</td>
<td>0.04</td>
<td>0.07</td>
<td>4.44</td>
<td>1.67</td>
</tr>
<tr>
<td>CaO</td>
<td>0.07</td>
<td>0.89</td>
<td>23.05</td>
<td>2.15</td>
<td>–</td>
<td>0.85</td>
<td>1.922</td>
<td>3.11</td>
<td>0.61</td>
<td>0.38</td>
<td>–</td>
<td>2.13</td>
</tr>
<tr>
<td>Cr(_2)O(_3)</td>
<td>0.08</td>
<td>0.32</td>
<td>–</td>
<td>0.03</td>
<td>0.03</td>
<td>0.15</td>
<td>0.35</td>
<td>0.08</td>
<td>0.16</td>
<td>0.50</td>
<td>0.07</td>
<td></td>
</tr>
<tr>
<td>SiO(_2)</td>
<td>37.29</td>
<td>55.24</td>
<td>53.65</td>
<td>67.00</td>
<td>37.61</td>
<td>54.17</td>
<td>52.96</td>
<td>53.89</td>
<td>56.90</td>
<td>56.35</td>
<td>55.54</td>
<td>69.04</td>
</tr>
<tr>
<td>P(_2)O(_5)</td>
<td>0.60</td>
<td>–</td>
<td>0.02</td>
<td>–</td>
<td>–</td>
<td>0.05</td>
<td>0.02</td>
<td>–</td>
<td>–</td>
<td>0.06</td>
<td>–</td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>100.30</td>
<td>98.95</td>
<td>99.98</td>
<td>99.90</td>
<td>100.06</td>
<td>99.66</td>
<td>98.48</td>
<td>100.55</td>
<td>98.91</td>
<td>100.00</td>
<td>100.00</td>
<td>96.51</td>
</tr>
</tbody>
</table>

Number of cations per formula unit

| Na | – | – | 0.008 | 0.727 | – | 0.003 | 0.018 | 0.009 | 0.016 | 0.09 | 0.06 | 0.139 |
| Ti | – | 0.001 | 0.006 | – | – | 0.003 | 0.003 | 0.007 | 0.004 | 0.01 | – | – |
| Mn | 0.005 | 0.009 | 0.005 | – | 0.004 | 0.008 | 0.006 | 0.004 | 0.032 | 0.02 | – | – |
| Mg | 1.483 | 1.539 | 0.885 | 0.001 | 1.480 | 1.547 | 1.435 | 0.894 | 2.924 | 3.02 | 3.60 | 0.007 |
| K | – | – | – | – | 0.060 | – | – | 0.002 | 0.001 | – | – | – |
| Fe | 0.524 | 0.417 | 0.183 | 0.006 | 0.538 | 0.471 | 0.369 | 0.180 | 0.800 | 0.81 | 0.21 | 0.025 |
| Al | – | 0.006 | 0.021 | 1.079 | 0.001 | 0.006 | 0.174 | 0.022 | 0.004 | 0.01 | 0.36 | 1.121 |
| Ca | 0.002 | 0.034 | 0.910 | 0.101 | – | 0.033 | 0.074 | 0.907 | 0.048 | 0.03 | – | 0.002 |
| Cr | 0.002 | 0.009 | 0.009 | 0.001 | 0.001 | 0.004 | 0.010 | 0.004 | 0.01 | 0.03 | 0.03 | 0.02 |
| Si | 0.971 | 1.994 | 1.977 | 2.939 | 0.987 | 1.962 | 1.914 | 1.974 | 4.084 | 4.02 | 3.79 | 3.036 |
| P | 0.017 | – | 0.001 | – | – | 0.002 | 0.001 | – | – | – | – | 0.004 |
| Fe\(^0\) | 0.26 | 0.21 | 0.17 | 0.10 | 0.27 | 0.23 | 0.20 | 0.17 | 0.21 | 0.21 | 0.06 | 0.23 |

Oxygens 4 6 6 4 6 6 6 12 12 12 8

Fe\(^0\)=Fe/(Fe+Mg) in atom. Ol: olivine; Pyx: low-Ca pyroxene; Di: diopside; h: in the host portion of the chondrite; Ri: ringwoodite; Msk: maskelynite; Mj: majorite; Py: pyrope; v: in melt veins; Ak: akimotoite; Jd: jadeite. a and b are energy-dispersive spectroscopy data from Tomioka & Fujino (1999) and Sharp et al. (1997), respectively. –: below detection limit.
Fig. 4. A back-scattered electron image of a shock-induced melt vein containing a polymineralic fragment. The fragment is composed of diopside, pyroxene, akimotoite, and ringwoodite, which are in triple-junction contact. Ol: olivine; Ri: ringwoodite; Di: diopside; Pyx: pyroxene; Ak: akimotoite; Jd: jadeite; Mj: majorite; Msk: maskelynite; Lng: lingunite.

Fig. 5. Raman spectra of some minerals in shocked Sixiangkou chondrite. Pyx: pyroxene; Mj: majorite; Ri: ringwoodite.

Fig. 6. Optical (Reflectance) microscopic photograph of the fragment akimotoite + low-Ca pyroxene. Ak: akimotoite; Pyx: pyroxene.

Fig. 5. Raman spectra of some minerals in shocked Sixiangkou chondrite. Pyx: pyroxene; Mj: majorite; Ri: ringwoodite.
polymineralic fragments found in melt veins of Sixiangkou.

One thin lensoid fragment (Fig. 2) of diopside and ringwoodite was observed at the edge of a shock-induced melt vein. The melt vein is about 50 μm wide, and was mainly composed of fine-grained majorite-pyrope, re-crystallized FeNi metal and troilite, and possible magnesiowüstite. A subhedral chromite grain (~ 7 μm) was also observed. The diopside + ringwoodite fragment was arbitrary defined as a polymineralic fragment due to the existence of high-pressure phase. The fragment is about 20 μm long and 5 μm wide. In the fragment, diopside has the same chemical composition as that in the host portion. It has sharp Raman bands at 1013, 666, 394 and 323 cm⁻¹ (Chopoulos & Serghiou, 2002), indicating that no phase transformation occurred in diopside. Ringwoodite has a similar composition to olivine in the host portion. Ringwoodite exhibits two Raman bands at 399 and 485 cm⁻¹, and olivine at 821 and 852 cm⁻¹.

A fragment of diopside + ringwoodite + akimotoite/low-Ca pyroxene was found at the center of a shock-induced melt vein (Fig. 4). The vein varies in width from 55 to 115 μm. The matrix of the melt vein is composed of fine-grained majorite-pyrope, FeNi metal and sulfide, and possible magnesiowüstite (Chen et al., 1996). Mineral fragments of ringwoodite and majorite (10 to 35 μm in size) also occur in this melt vein. The diopside + ringwoodite + akimotoite/low-Ca pyroxene fragment appears ovoid in shape, and is about 50 μm long and 20 μm wide. Diopside, ringwoodite, akimotoite/low-Ca pyroxene are in triple-junction contact. Akimotoite coexists with low-Ca pyroxene. Diopside has similar Raman spectra and chemical composition (En13Fs8Wo1) to that in the host portion. Ringwoodite has a composition of Fo78Fs21Wo1, similar to olivine in the host portion. Ringwoodite is characterized with two sharp Raman bands at 795 and 843 cm⁻¹. Akimotoite is identified with two typical Raman bands at 798 and 476 cm⁻¹ that could not be assigned to other high-pressure polymorphs of low-Ca pyroxene. Akimotoite (En78Fs13Wo1) and coexisting low-Ca pyroxene (En77Fs21Wo1) have the same chemical composition, identical to the result of Chen et al. (1996). The chemical composition of akimotoite is similar to akimotoite reported in Tenham (En78Fs21Wo1: Tomioka & Fujino, 1997) and Yamato 791384 (En78Fs21Wo1: Ohtani et al., 2004), but differs significantly from those in Acfer 040 (En84Fs6: Sharp et al., 1997), Umbarger (En88Fs11Wo1: Xie & Sharp, 2004), and Tanhem (En88Fs12: Xie et al., 2006).

A fragment of akimotoite and low-Ca pyroxene was also observed in a shock-induced melt vein (Fig. 6). This melt vein is about 50 μm in width. The matrix is also composed of fine-grained majorite-pyrope, dendritic or round metal-troilite, and possible magnesiowüstite. No large mineral fragments were observed in this melt vein. The fragment is lensoid and about 15 × 25 μm in size. Under reflectance optical microscope, akimotoite and low-Ca pyroxene exhibit different reflectance (Fig. 6). Akimotoite is mainly set at the center of the fragment. Akimotoite and low-Ca pyroxene have very similar chemical compositions. Akimotoite was characterized with three bands at 797, 617, and 479 cm⁻¹ (Fig. 5), which is in good agreement with the result of Ohtani et al. (2004). A broad band was observed in the range of 820-1200 cm⁻¹, which indicates the presence of amorphous substance.

4. Discussion

4.1. Stability of diopside during shock metamorphism

The pressure and temperature of the shock-induced melt veins of the Sixiangkou chondrite could be constrained by high-pressure mineral assemblages (Chen et al., 1996, 2002). By comparing the assemblage of majorite-pyrope + magnesiowüstite with static high-pressure experimental data of Zhang & Herzberg (1994) and Agee et al. (1995), Chen et al. (1996) concluded that the shock-induced melt veins should have crystallized at pressures of 20-24 GPa and temperatures of above 1000°C. Based on the fact that the occurrence of metal-troilite-magnesiowüstite assemblage in the shock-induced melt veins and ringwoodite lamellae in olivine, a long duration of high pressure up to several minutes was suggested (Chen et al., 2002, 2004a, and 2006).

Diopside has not been observed from melt veins and the host portion of the Sixiangkou chondrite in previous investigations (e.g., Chen et al., 1996), but was reported in other chondritic meteorites, such as the Suizhou L6 chondrite (Chen et al., 2004b), the Yamato 791384 L6 chondrite (Ohtani et al., 2004), and the Yamato 751000 H6 chondrite (Tomioka & Kimura, 2003). In this study, diopside was observed not only in some fragments of melt veins but also in the host portion of the Sixiangkou chondrite. It indicates that diopside was a relict mineral during the shock metamorphism of the chondrite rather than crystallized from the shock-induced melt veins described in other heavily shocked chondrites, e.g., Umbarger L6 chondrite (Xie & Sharp, 2004).

In the shock-induced melt veins of the Sixiangkou chondrite, diopside was closely associated with ringwoodite (+ akimotoite) within the same fragments. The occurrence of diopside in the veins suggests that either the local pressure-temperature conditions did not exceed the stability field of diopside or the rate of phase transformation of diopside is very low and the duration is too short to induce the transformation.

High-pressure experiments have been performed to understand the stability and phase transition of diopside, and most of them were carried out at high temperatures (> 1000°C) (e.g., Canil, 1994; Gasparik, 1996 and references therein; Oguri et al., 1997; Irfune et al., 2000; Asahara et al., 2005). These experimental data show that diopside would become unstable at pressures higher than 18-19 GPa and temperatures > 1000°C, and would dissoicate into high-pressure phases of CaSiO₃ and MgSiO₃ (or Mg₂SiO₄ + SiO₂) (Irfune et al., 1989; Canil, 1994; Oguri et al., 1997). Recently, Tomioka & Kimura (2003) reported in a shocked H chondrite the breakdown of diopside to Ca-rich majorite and glass, and inferred a breakdown pressure about 18-24 GPa and temperature at 1100-1900°C.
Chopelas & Serghiou (2002) measured Raman spectra of diopside with increasing pressures at room temperature, and found that the Raman spectrum of diopside at 20.2 GPa was nearly identical to that at 1 atm, suggesting no phase transformation. These investigations suggest that temperature is an important factor affecting the phase transition of diopside. At a given high pressure, diopside is more stable at low temperature than high temperature. The presence of ringwoodite rather than wadsleyite in the coarse-grained fragments would suggest a pressure greater than 20 GPa (Chen et al., 1996). At such a high pressure, a low temperature could be inferred in order to preserve diopside during the heavily shock metamorphism.

The above explanation did not exclude the kinetic effect of phase transformation of diopside during shock metamorphism. At a given high pressure and temperature, the phase transformation of diopside is controlled by the transformation rate of diopside and the duration of high pressure and high temperature. Recent observations showed that the duration of high pressure could be up to several seconds, even several minutes in shock-induced melt veins (Chen et al., 2002, 2004a; Ohtani et al., 2004; Chen et al., 2006; Xie et al., 2006). A long duration of high pressure up to several seconds was also suggested in the Sixiangkou chondrite (Chen et al., 2002, 2004a, and 2006). With such a long duration many minerals would have been partly or completely transformed into their high-pressure polymorphs. The survival of diopside suggests that phase transformation rate of diopside would be very sluggish. A possible interpretation is that the crystal structure of diopside contains the large radium cation Ca$^{2+}$ (0.99 Å), which is more difficult to modify its coordination under high pressures than Mg$^{2+}$ (0.66 Å) and Fe$^{2+}$ (0.74 Å). However, the kinetics of phase transformation of diopside remains unclear.

4.2. Transformation from pyroxene to majorite

As reported in previous investigations (e.g., Chen et al., 1996), majorite is the main high-pressure phase with a pyroxene composition in the shock-induced melt veins of the Sixiangkou chondrite. Two mechanisms have been proposed to explain the formation of coarse-grained polycrystalline majorite and fine-grained majorite-pyrope (Chen et al., 1996; Ohtani et al., 2004; Sharp & DeCarli, 2006). The coarse-grained polycrystalline majorite was believed to have formed through solid state transformation from low-Ca pyroxene. The main reason was that the coarse-grained polycrystalline majorite grains lack an idiomorphic polycrystalline appearance and have the same chemical composition as low-Ca pyroxene in the host portion (Chen et al., 1996, 2004b; Ohtani et al., 2004). Sharp & DeCarli (2006) suggested that this solid state transformation mechanism was homogenous intracrystalline nucleation followed by interface-controlled growth. The fine-grained majorite-pyrope was thought to have crystallized from melts under pressure because the solid solution contained higher contents of Ca and Al than low-Ca pyroxene in the host portion (Chen et al., 1996).

4.3. Transformation from pyroxene to akimotoite

Akimotoite has been previously found in a few of L6 shocked chondrites (Acfer 040, Umbarger, Yamato 791384, Tenham) (Sharp et al., 1997; Tomioka & Fujino, 1997, 1999; Ohtani et al., 2004; Xie & Sharp, 2004; Xie et al., 2006). Two models have been proposed for its formation: solid-state transformation (Tomioka & Fujino, 1997, 1999) and crystallization from a high-density melt under high pressures (Sharp et al., 1997; Xie & Sharp, 2004; Xie et al., 2006). Tomioka & Fujino (1997, 1999) observed chemical similarities among clinoenstatite, akimotoite, and perovskite in Tenham and suggested their formation by solid-state transformation at high pressures. Our results in Sixiangkou are consistent with those of Tomioka & Fujino (1997, 1999). On the other hand, akimotoite grains in Acfer 040 and Umbarger have different chemical compositions from their respective host pyroxenes. In particular, akimotoite in Acfer 040 and Umbarger is rich in Al$_2$O$_3$, indicating crystallization from the shock-induced melt (Sharp et al., 1997; Xie & Sharp, 2004; Xie et al., 2006).

Akimotoite in the Sixiangkou chondrite has a similar mineral chemistry as that in Tenham (Tomioka & Fujino, 1997, 1999) and Y791384 (Ohtani et al., 2004), but differs significantly in chemical composition from that in Acfer 040 (Sharp et al., 1997) and in Umbarger (Xie & Sharp, 2004). This discrepancy may indicate that akimotoite has a rather large compositional stability field. However, up to date, no compositional effect on stability of akimotoite has been known, and it remains an open question for further investigation on the effects of Fe/(Fe$^3+$Mg) ratio, Fe$^{3+}$, and Al$^{3+}$ on the high-pressure stability of akimotoite. In addition, the fact that low-Ca pyroxene was partly transformed into akimotoite may suggest that the phase transformation of low-Ca pyroxene to akimotoite was more sluggish than that of olivine to ringwoodite (Hogrefe et al., 1994; Sharp & DeCarli, 2006 and references therein).

4.4. Formation of jadeite in the shock-induced melt veins

In this study, jadeite was only observed in the shock-induced melt veins, similar to those described in other shocked chondrites (Kimura et al., 2000, 2001; Tomioka & Kimura, 2003; Ohtani et al., 2004). In these chondrites, jadeite had chemical compositions similar to plagioclase (Ohtani et al., 2004), suggesting that jadeite has originated from plagioclase or its high-pressure polymorphs. Liu (1978) reported that with increasing pressures albite would first transform to jadeite and SiO$_2$ (quartz, coesite, and stishovite) and then to hollandite-structured. A possibility has been proposed by Kimura et al. (2000) that jadeite might have formed through retrograde transformation of lingunite under moderate post-shock pressure and high temperature conditions. This explanation may also be suitable to the jadeite in this study. The undetected SiO$_2$ may be due to the overlap of the Raman bands at $\sim$ 520 cm$^{-1}$ with jadeite. More detailed studies are needed to determine the origin and formation mechanism of jadeite in the future.
4.5. Implications for shock metamorphism

The coexisting of a variety of pyroxene phases in shock-induced melt veins has an important implication for the P-T history of melt veins and phase transformation processes of minerals. Chen et al. (1996) reported that the pressure and temperature in the shock-induced melt veins of the Sixiangkou meteorite could be up to 20-24 GPa and ~2000°C, respectively. This pressure and temperature were suitable for the formation of majorite in melt veins, and the high temperature is too high for the preservation of akimotoite. Akimotoite and majorite have a similar pressure stability range (Presnell, 1995). If the temperature of the whole melt veins was up to 2000°C, no akimotoite was expected to be observed. The presence of akimotoite in melt veins must reflect a local low temperature that has induced the transformation of low-Ca pyroxene to akimotoite and could not induce the transformation of akimotoite to majorite. Chen et al. (2004b) explained that the inhomogeneity of temperature in shock-induced melt veins was related to the widths of melt veins. Relatively large amounts of shock-induced melt were produced in thick shock veins (up to several mm in widths), and they must contain larger amounts of heat and could heat all material inside the veins to a homogeneous high-temperature (Chen et al., 2004b). However, thin shock veins produced less melt and the relatively small amounts of heat could not heat all materials to a homogeneous high-temperature in a short time (Chen et al., 2004b). In this study, our observations are generally consistent with this interpretation by Chen et al. (2004b) with akimotoite only occurring in some thin melt veins (not thicker than 150 µm). The fact that diopside remained intact and coexists with ringwoodite (+akimotoite) might be another evidence of local low temperature in melt veins. The low temperature failed to overcome the large activation energies of reconstructing the crystal structure of diopside (DeCarli et al., 2002; Sharp & DeCarli, 2006).

Phase transition of minerals during shock metamorphism is a rather complex kinetic process, and different from static high-pressure experiments. For instance, in static high-pressure experiments, at a high pressure of 18-20 GPa and a temperature of above 900°C, olivine would transform into ringwoodite on an observable timescale (Sharp & DeCarli, 2006 and references therein). However, ringwoodite has never been recovered in shock recovery experiments (Stöffler et al., 1991). Thus, it was suggested that phase transition of minerals during shock metamorphism is both temperature and time dependent (Sharp & DeCarli, 2006). As discussed above, the inhomogeneous temperature in the shock-induced melt veins has lead to the phase transition of some minerals (olivine to ringwoodite, low-Ca pyroxene to akimotoite and majorite) and the preservation of some minerals (e.g., diopside). Chen et al. (2002, 2004a, and 2006) have concluded that the duration of high-pressures in melt veins was up to several seconds and even several minutes. This duration is long enough for the phase transitions of olivine to ringwoodite and low-Ca pyroxene to majorite in melt veins because most olivine and low-Ca pyroxene grains have completely transformed into ringwoodite and majorite. The incomplete transformation of low-Ca pyroxene to akimotoite observed in some fragments of melt veins must reflect a low transformation rate. This conclusion is consistent with previous observations that the transformation of enstatite to akimotoite is more sluggish than that of olivine to ringwoodite (Sharp & DeCarli, 2006).

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