The pressures and temperatures of meteorite impact: Evidence from micro-Raman mapping of mineral phases in the strongly shocked Taiban ordinary chondrite

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ABSTRACT

Taiban is a heavily shocked L6 chondrite showing opaque melt veins. Raman spectroscopy was used to characterize the high-pressure silicate assemblages in a thin section crossed by a shock-created 4 mm wide melt vein. Raman spectra using different excitation wavelengths allowed identification of mineral phases such as olivine, wadsleyite, ringwoodite, high-Ca clinopyroxene, majorite-pyrope, jadeite, maskelynite, and lingunite. Olivine is Fe depleted in contact with the ringwoodite, which suggests chemical fractionation during a solid-state olivine-ringwoodite transformation. Raman imaging revealed a close correlation between the blue ringwoodite color and the peak observed at 877 cm\(^{-1}\); this signal shows strong near-resonance Raman enhancement when measured with near-IR excitation lines (785 and 830 nm) close to the optical absorption bands of the ringwoodite. We propose that the blue color of the ringwoodite is due to a small amount of iron in fourfold coordination inside the spinel structure, and that yields the observed spectral features in differently colored ringwoodite. Under the formation conditions of the studied silicate pocket, all enstatite transformed to a majorite-pyrope solid solution, whereas the high-Ca clinopyroxene likely remained unchanged. Maskelynite grains in the margins of the pocket often contain lingunite or are totally transformed to jadeite. Based on static high-pressure results, the mineral assemblages in the pocket suggest peak pressure in the 17–20 GPa range with maximum temperature (\(T_{\text{max}}\)) in the range 1850–1900 K as the formation conditions for this Taiban chondrite during shock.

Keywords: Micro-Raman, resonance Raman, ringwoodite, solid-state transformation, shock metamorphism

INTRODUCTION

High-pressure polymorphs of major minerals are commonly found in shocked meteorites. During the formation and later evolution of the Solar System (Raymond 2010; Wetherill 1980), asteroids repeatedly collided with each other and with larger bodies triggering shock waves. Meteorites that underwent such shock waves show different shock effects depending on the peak pressures and temperatures, and duration of the collision events (Stöffler et al. 1991).

Minerals commonly found in or near melt veins in highly shocked chondrites include: ringwoodite; majorite; wadsleyite; magnesiowüstite; akimotoite; lingunite; silicate–perovskite; and maskelynite (Chen et al. 1996; Chen and El Goresy 2000; Ferroir et al. 2008; Mori 1994; Price et al. 1983; Sharp et al. 1997; Tomioka and Fujino 1997; Xie et al. 2006). As the shock front triggered by a collision spreads through the irregularities of the rock, melt veins and shocked pockets form depending on pressure and temperature conditions and the properties of the rock in each point. High-pressure phases form in these areas by solid-state transformations of the original minerals or fractional crystallization from the melt. Based on results from static high-pressure and shock recovery experiments, the observed crystallization assemblages can aid in constraining the conditions of melt vein crystallization (Sharp and De Carli 2006). However, these methods present some difficulties. Shock recovery experiments probably produce lower temperatures and shorter shock durations than natural impacts. Similarly, equilibrium phase diagrams obtained from static high-pressure experiments fail to reproduce the large departures from equilibrium that mineral phases experience during shock. Furthermore, static experiments do not produce direct transformation from low-pressure phases without forming intermediate structures (e.g., olivine to ringwoodite without forming wadsleyite) (Sharp and De Carli 2006; Stöffler et al. 1991; Xie and Sharp 2007; Xie et al. 2006).

Ringwoodite \(\gamma-(\text{Fe,Mg})_2\text{SiO}_4\) is the high-pressure polymorph of olivine with spinel structure. It is found as a rare mineral in shocked meteorites and arguably in impact craters by transformation from olivine, and is considered to be the major constituent of the lower part of the mantle transition zone (Binns et al. 1969; Ringwood 1975; Rull et al. 2007). Thus, the physical properties of the ringwoodite are key to understanding the behavior of that zone in the Earth’s interior. The mineral has also been subject of abundant research since its discovery by Binns et al. (1969) in the Tenham meteorite (e.g., Gupta and Goyal 2011; Madon and Poirier 1983; Mosenfelder 2001; Price et al. 1982; Sinogeikin et al. 2003). Raman spectroscopy has been used extensively to assess the nature of the olivine polymorphs and other shock-induced phases (e.g., Chen et al. 2007; Miyahara et al. 2008; Zhang et al. 2006). However, only a few studies have taken...
advantage of the potential of Raman spectroscopy to provide compositional and structural information on the shock-induced ringwoodite phases (e.g., Feng et al. 2011).

Micro-Raman spectroscopy has successfully been used to provide unequivocal identification of minerals and glassy phases in meteorites as well as to obtain additional information on structural and compositional variations inside mineral grains (Wang et al. 2001, 2004; Cooney et al. 1999). In combination with a microscope and an automated X-Y-Z stage, Micro-Raman spectroscopy can resolve sample sizes as small as 1 μm and place the data in its spatial context. Raman mapping, or Raman imaging, is a powerful technique that combines the mineralogical information included in the Raman spectra with spatial distribution, thus providing additional information on the petrography of the studied rocks and allowing for constraints on the crystallization histories (Fries et al. 2011; Frosch et al. 2007; Miyahara et al. 2008; Steele et al. 2007).

**SAMPLE**

Taiban is a very strongly shocked, shock stage S6 ordinary L6 chondrite found in De Baca County, New Mexico, in 1975 (Lange et al. 1980). It shows multiple veins of opaque shock melt produced by shearing during the shock event that extracted the meteorite from its parent body. The melt veins form a network of complex branches surrounding pockets of highly shock-altered, mosaicized (i.e., made into a mosaic), relict, or recrystallized silicates, and their high-pressure polymorphs, along with droplets of immiscible troilite and metallic inclusions. Outside the melt vein, olivine presents only mosaicism with diaplectic plagioclase glass (maskelynite) interstitial to olivine and the different pyroxenes (Scott et al. 2004; Stöffler et al. 1991). We studied the polished thin section UNM297 shown in Figure 1, which is intersected by a 4 mm wide melt vein in which the shock silicate pockets are dominated by blue and white ringwoodite crystals that range in size from 10 to 100 μm. As ringwoodite is assembled in clasts within the vein and the surrounding white matrix, the high-pressure phase did not crystallize from the impact melt (Scott et al. 2004). Minor amounts of host olivine and the intermediate phase β-(Mg,Fe)₂SiO₄ wadsleyite also appear associated with the ringwoodite. The white matrix is composed mainly of different pyroxenes (high-Ca clinopyroxene and jadeite) and the high-pressure garnet polymorph of pyroxene, majorite. Plagioclase quenched to glass is conspicuous in the vicinity of the melt vein but was not found inside the lithic pockets or the melt areas; maskelynite grains present flow textures filling gaps between olivine and pyroxene. No lamellar intergrowth structures of ringwoodite and olivine were found so far in Taiban meteorite under optical microscope and SEM observations.

**METHODS**

Raman spectroscopy with different excitation wavelengths was used in combination with optical images. Raman spectra were taken with three different instruments at the University of Hawai‘i. Spectra with 785 nm excitation wavelength were taken with a Kaiser Optical Systems’ micro-Raman system. The system is composed of a 785 nm Invictus diode laser, a Leica microscope with imaging capabilities, a Kaiser HoloSpec spectrometer, and an Andor CCD camera. A 100 μm optical fiber transmits the laser light to the microscope and the Raman signal to the spectrometer. The laser is focused and the signal is observed through a 100x objective mounted on the microscope in backscattering geometry; the laser spot size on the sample was 2 μm. The system has a spectral range of 150–3300 cm⁻¹. Typical acquisition time was 150 s and laser output power was limited to 10 mW, after making sure no sample damage was induced at that power. To look for resonance Raman effects and allow assessment of Raman peaks vs. fluorescence peaks, additional spectra were taken with a Renishaw inVia micro-Raman system coupled with a 830 nm Invictus diode laser, a 244 nm frequency doubled Ar ion Lexus laser, and a 514.5 nm Ar ion laser. The microscope and spectrometer are coupled through optical mirrors and all the optics can be set to accommodate different wavelengths. The 514.5 and 830 nm lasers were focused on the sample through a 100x objective that also acted as a signal collector; laser spot sizes were approximately 2 μm for both wavelengths. Due to high absorption in the UV range, a 40x UV objective was used for the 244 nm line, which focuses the beam down to 5 μm spot size. Laser
powers were kept under 5 mW for 830 nm, 1 mW for 514.5 nm, and 0.2 mW for 244 nm to avoid laser-induced heating or sample photo-damage. Raman spectral ranges were 120–1600 cm$^{-1}$ for 830 nm, and from 300–4000 and 500–4000 cm$^{-1}$ for 514.5 and 244 nm, respectively.

Spectra with 532 nm excitation were collected using a WiTec Alpha300R confocal Raman microscope with a frequency doubled Nd-YAG laser (Coherent Compass) beam at 532 nm. The power of the 532 nm laser beam at the sample was 3 mW. All the spectra were acquired through a 100× objective with 300 s exposure time and a laser spot of around 2 μm. The Raman systems used for this study have a peak resolution of 6 cm$^{-1}$ for the 514, 785, and 830 nm Raman systems, and 10 and 12 cm$^{-1}$ for the UV 244 and 532 nm Raman systems, respectively (using the FWHM of the 1085 cm$^{-1}$ calcite line measured with the various Raman systems). The mineral assemblages in the thin section of the Taiban meteorite were observed in transmitted, reflected, and cross-polarized light. Different areas were examined inside and outside the ~4 mm melt vein. The main focus of the study was a silicate pocket inside the melt vein, surrounded by quenched glass. Outside the melt vein silicates present milder shock effects with the conspicuous presence of maskelynite.

For detection of minor phases, line and area scans were performed with the aid of an X-Y-Z stage with submicrometer accuracy in addition to point Raman measurements for 244, 785, and 830 nm. The sample was scanned under the laser beam in 2D for a Raman chemical image. Typically, step distances were chosen to match the 2 μm laser spot size and exposure times were set to 150 s. Data obtained during mapping runs were examined to locate the minor phases and obtain typical spectra for each phase and generate the Raman images. Typical images cover an area of approximately 600 μm$^2$ over 12 × 12 pixels and each phase was identified by its Raman signature peaks. Map intensity scales were generated by integrating the area or the peak height of each of the phases: the 856 cm$^{-1}$ peak for olivine; 1013 cm$^{-1}$ peak for high-Ca pyroxenes; 929 cm$^{-1}$ peak for blue ringwoodite. Background subtraction and cosmic ray removal software processing was applied for each acquisition. Additional processing was needed to remove the small contribution of the microscope slide substrate and epoxy resin to the Raman signal. This was necessary for mineral phases transparent to the laser light at each wavelength. Representative spectra from the substrate were obtained at the edges of the slide and then subtracted from each of the spectra. Spectra operations and curve fittings were performed with the Grams/AI 8.0 software package (Thermo-Fisher Scientific, Inc.).

After the Raman measurements, the thin section was coated with carbon and studied with an electron microprobe (JEOL JXA-8500F) equipped with 5 tunable wavelength-dispersive spectrometers at the University of Hawai‘i. Mineral chemical analyses were performed with a beam energy of 15 keV, a beam current of 15 nA, and an electron beam diameter of typically 5 μm. A combination of natural and synthetic standards was used for the calibration. Oxygen was calculated by cation stoichiometry and included in the matrix correction. Typical detection limits were 0.010 wt% for AlKα, 0.014 wt% for NaKα, 0.019 wt% for TiKα, and 0.035 wt% for MnKα. The matrix correction method was ZAF or Phi-Rho-Z calculations and the mass absorption coefficients data set was LINEMAT.

## Results and Discussion

Raman spectra imaging and single point measurements were taken in and around the biggest ringwoodite crystal; the results from the mineral characterization are shown in Figure 2. Some additional spectra were recorded in other areas of the white matrix, other silicate pockets, and in several points in the mineral assemblages outside the melt vein.

**Olivine, wadsleyite, and ringwoodite**

Raman signature peaks of olivine are found at several points in and around the ringwoodite. The spectra of the grains within the ringwoodite and between grains show the characteristic peaks of olivine at 824 and 856 cm$^{-1}$, referred to as DB1 and DB2 bands, respectively (after Kuebler et al. 2006), in good agreement with the shifts found in forsterite (Chopelas 1991a). This pair originates from the Si-O$_2$ bond stretching in the [SiO$_4$]$^{4-}$ tetrahedral and have a mixed character of $\delta_1$ and $\delta_2$ modes (Lam et al. 1990). An additional olivine peak appears at 962 cm$^{-1}$ within the ringwoodite and in the olivine filling the 1 μm crack between the ringwoodite grains. Olivine and pyroxene-majorite assemblages or the melt vein are in direct contact with the ringwoodite around the rest of the perimeter. According to a Raman composition calibration of olivine developed by Kuebler et al. (2006), olivine intermingled with the ringwoodite inside the lithic pocket in the melt vein is Fo$_{90}$, while olivine filling the 1 μm crack is Fo$_{94}$. In the host rock, olivine is Fo$_{89}$ (Fig. 3, Tables 1 and 2). The olivine inside the pocket is intermingled with the ringwoodite formed after shock and found in small grains showing dendritic textures or filling the 1 μm crack between ringwoodite grains (Figs. 2 and 4). The discrepancy in the calculated Fo of the olivine located within the ringwoodite is most likely due to the difference in sampling depth between the two techniques and the Raman
spectra of that area showing a contribution of both olivine with dendritic shapes and ringwoodite, thus inducing a bigger error in the estimation of the Raman peak positions. The olivine in the host rock is predominantly fine grained (however, crystals up to 100 µm are found in some areas) and is also found in inclusions and among high-Ca and low-Ca pyroxenes, and maskelynite.

Spectra taken in the inner points of the crystal show Raman fingerprints of pure ringwoodite. The vibrations located at 301, 589, 799 (referred to as RA1), and 844 cm⁻¹ (referred to as RA2) correspond to ringwoodite, γ-(Fe,Mg)SiO₃. These vibrations have also been observed in natural ringwoodite lamellae in the Sixiangkou meteorite (Chen et al. 2007), NWA 1662 (Taran et al. 2009), and in synthetic ringwoodite (Akaogi et al. 1984; Chopelas et al. 1994; Kleppe et al. 2002a; McMillan and Akaogi 1987). The Raman mode at 844 cm⁻¹ corresponds to pure Si-O bond stretching, \( \nu_{1g} \) symmetry. The signals found at 301, 589, and 799 cm⁻¹ are \( T_{2g} \) modes, opposite oscillations of \( \nu \) modes, symmetric. The signals found at 514.5, and 830 nm confirm that the peak observed at 877 cm⁻¹ corresponds to the ringwoodite structure. The same peak was observed elsewhere and researchers have proposed that it is caused by the presence of a defect-induced vibrational mode, to a inverse spinel structure in which the Mg and Fe ions are both in fourfold and sixfold coordination, or to the presence of glassy material (Kleppe et al. 2002b; Nagy et al. 2010a, 2010b).

As shown in Figures 5 and 6, the intensity of the 877 cm⁻¹ signal presents a remarkably accurate spatial correlation with the color observed by optical microscopy and polarized light. The intensity is higher when the ringwoodite is deep blue, intermediate in clear blue areas, and absent when the ringwoodite is colorless. This

![Figure 3](image)

**Figure 3.** Olivine inside the pocket in the melt vein (A) and in the host rock (B) a few millimeters away from the melt vein in the same microscope slide. The olivine inside the pocket is intermingled with the ringwoodite formed after shock. The olivine in the host rock is fine grained and found among pyroxenes. Raman lines are shifted up inside the melt vein indicating a higher Fo content than the host olivine.

**Table 1.** Peak positions of DB1 and DB2 Raman spectra of olivine recorded in different lithologies of the Taiban meteorite

<table>
<thead>
<tr>
<th>Olivine location</th>
<th>DB1 (cm⁻¹)</th>
<th>DB2 (cm⁻¹)</th>
<th>Fo(%) (Raman)</th>
<th>Fo(%) (EPMA)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ol in crack</td>
<td>823.8</td>
<td>856.4</td>
<td>93.8</td>
<td>93.6</td>
</tr>
<tr>
<td>Ol within ringwoodite</td>
<td>823.4</td>
<td>868.6</td>
<td>87.6</td>
<td>92.1</td>
</tr>
<tr>
<td>Ol host rock</td>
<td>820.9</td>
<td>852.4</td>
<td>77.8</td>
<td>75.5</td>
</tr>
</tbody>
</table>

Note: Mg/(Mg+Fe) ratios are calculated both using the Raman calibration developed by Kuebler et al. (2006) and by EPMA. Accuracies of the Raman calculated values are within Fo±10%, and those of the EPMA are within Fo±5%.

**Table 2.** EPMA analyses of minerals in the Taiban chondrite

<table>
<thead>
<tr>
<th></th>
<th>Mjt</th>
<th>Cpx</th>
<th>Rgt l-bl</th>
<th>Rgt cl</th>
<th>Rgt cl2</th>
<th>Ol crk</th>
<th>Ol 1</th>
<th>Ol 2</th>
<th>Ol 3</th>
<th>Ol host</th>
</tr>
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<tbody>
<tr>
<td>SiO₂</td>
<td>52.99</td>
<td>53.87</td>
<td>37.73</td>
<td>37.95</td>
<td>37.38</td>
<td>36.80</td>
<td>40.47</td>
<td>40.63</td>
<td>40.38</td>
<td>40.80</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.17</td>
<td>0.46</td>
<td>–</td>
<td>0.02</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>0.18</td>
<td>0.53</td>
<td>–</td>
<td>0.01</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>0.10</td>
<td>0.77</td>
<td>–</td>
<td>0.18</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>FeO</td>
<td>14.53</td>
<td>4.89</td>
<td>24.82</td>
<td>24.69</td>
<td>25.96</td>
<td>25.28</td>
<td>25.82</td>
<td>25.28</td>
<td>25.96</td>
<td>25.28</td>
</tr>
<tr>
<td>MnO</td>
<td>0.04</td>
<td>0.08</td>
<td>0.08</td>
<td>0.04</td>
<td>0.04</td>
<td>0.07</td>
<td>0.25</td>
<td>0.25</td>
<td>0.25</td>
<td>0.25</td>
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<tr>
<td>MgO</td>
<td>28.33</td>
<td>16.87</td>
<td>37.17</td>
<td>37.28</td>
<td>36.28</td>
<td>37.12</td>
<td>49.19</td>
<td>48.11</td>
<td>48.37</td>
<td>48.37</td>
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<tr>
<td>CaO</td>
<td>0.70</td>
<td>21.85</td>
<td>0.05</td>
<td>0.07</td>
<td>0.04</td>
<td>0.02</td>
<td>0.04</td>
<td>0.06</td>
<td>0.07</td>
<td>0.07</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.01</td>
<td>0.48</td>
<td>0.01</td>
<td>–</td>
<td>0.01</td>
<td>0.01</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>0.01</td>
</tr>
<tr>
<td>Total</td>
<td>97.53</td>
<td>100.02</td>
<td>99.96</td>
<td>100.32</td>
<td>99.85</td>
<td>99.42</td>
<td>99.87</td>
<td>98.97</td>
<td>99.42</td>
<td>99.44</td>
</tr>
<tr>
<td>Fo₆₈</td>
<td>72.75</td>
<td>72.91</td>
<td>71.36</td>
<td>72.35</td>
<td>93.17</td>
<td>91.02</td>
<td>92.02</td>
<td>93.44</td>
<td>75.46</td>
<td>75.46</td>
</tr>
</tbody>
</table>

Calculated cations per formula unit

<table>
<thead>
<tr>
<th></th>
<th>Si</th>
<th>Ti</th>
<th>Al</th>
<th>Cr</th>
<th>Fe</th>
<th>Mn</th>
<th>Mg</th>
<th>Ca</th>
<th>Ni</th>
<th>P</th>
<th>Na</th>
<th>O</th>
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<tbody>
<tr>
<td>wt%</td>
<td>3.9180</td>
<td>0.0097</td>
<td>0.0154</td>
<td>0.0059</td>
<td>0.8964</td>
<td>0.0303</td>
<td>3.1231</td>
<td>0.0552</td>
<td>0.0016</td>
<td>0.0019</td>
<td>0.0019</td>
<td>0.0012</td>
</tr>
</tbody>
</table>

Notes: Mjt = majorite; Cpx = high-Ca pyroxene; Rgt l-bl = light-blue ringwoodite; Rgt cl = colorless ringwoodite; Rgt bl = blue ringwoodite; Rgt cl2 = clear ringwoodite; Ol crk = olivine in the crack of the studied ringwoodite; Ol 1, Ol 2, Ol 3 = three measurements of olivine within the studied ringwoodite crystal; Ol host = olivine in the host rock; – = below detection limit. Fo₆₈ = [Mg/(Fe+Mg)] × 100.
applies to at least five different ringwoodite grains of different and variable colors in different instruments, ruling out crystal orientation being the cause of the 877 cm$^{-1}$ peak. The ringwoodite spectra also present twofold broadening of line widths in blue ringwoodite with a comparable 877 cm$^{-1}$ peak contribution with respect to the lines RA1 and RA2. White ringwoodite, with a negligible 877 cm$^{-1}$ contribution, shows a half width of 22 cm$^{-1}$ for the RA1 peak, while blue ringwoodite shows an increased half width to 54 cm$^{-1}$.

The ringwoodite spectrum shows additional bands at 495 cm$^{-1}$ and a strong band at 225 cm$^{-1}$ with a shoulder at 175 cm$^{-1}$. These bands (also observed by Sharp et al. 2009 and Taran et al. 2009 for blue ringwoodite) are not due to fluorescence because they are observed at the same Raman shift with 532, 785, and 830 nm laser excitation wavelengths. As in the observations of Sharp et al. (2009) and Taran et al. (2009), these bands appear only for blue ringwoodite when excited with a 532 nm laser; however, they are observed for both clear and blue ringwoodite when excited with near-IR wavelengths (785 and 830 nm), although less intense in the clear samples. These bands probably correspond to translational lattice modes in the ringwoodite structure because the positions are close to those measured for lattice modes of forsterite (Iishi 1978). Raman measurements also show that it is anhydrous ringwoodite, because no O-H modes have been detected under 244 or 514.5 nm excitation wavelengths. Minor amounts of wadsleyite are inferred from weak Raman peaks at 723 and 919 cm$^{-1}$, observable under 514.5 nm excitation (Chopelas 1991b; McMillan and Akaogi 1987).

Previous observations showed two different mechanisms for the transformation from olivine to its high-pressure polymorphs. Two mechanisms have been proposed for the crystallization of ringwoodite based on a solid-state transformation: incoherent nucleation on olivine grain boundaries; or coherent intracrystalline nucleation of ringwoodite (Chen et al. 2006; Kerschhofer et al. 1996, 1998, 2000; Kubo et al. 2004; Ohtani et al. 2004; Putnis and Price 1979). More recent observations (Feng et al. 2011; Miyahara et al. 2008) suggest fractional crystallization of wadsleyite and ringwoodite from a melt of olivine composition. Ringwoodite in the Taiban meteorite contains minor amounts of wadsleyite and olivine depleted in Fe to different extents in the vicinity of the ringwoodite grains with respect to the host olivine (Fo$_{78}$–Fo$_{94}$, Tables 1 and 2). Thus, our results suggest that some chemical fractionation took place. The size, shape, and assemblages of the ringwoodite clasts in the pyroxene white matrix entrained in the melt vein and the absence of wadsleyite crystals suggest a solid-state transformation. Reflected light microphotography of the relict Fe-depleted olivine entrained in the bigger ringwoodite crystal shows darker olivine surrounded by ringwoodite due to the difference in light reflection coefficients.

![Figure 4](image_url)  
**Figure 4.** Olivine showing dendritic textures intermingled with the ringwoodite grain. Inset shows the same area in backscattered electrons image; scale bar is 10 μm.

![Figure 5](image_url)  
**Figure 5.** Raman spectra of (A) colorless and (B) deep-blue ringwoodite. The intensity of the 877 cm$^{-1}$ peak is enhanced in the blue ringwoodite due to the resonance Raman Effect with excitation at 785 nm. Several new bands also appear in the low-frequency region for deep-blue ringwoodite. 785 nm laser excitation, 10 mW power output, 150 s acquisition time. (C) Curve-fitting result for the spectrum of deep-blue ringwoodite with Lorentz-Gauss curve shapes. 830 nm laser excitation, 5 mW power output, 300 s acquisition time.
between the two fractionated polymorphs (Fig. 4). The dendritic textures of the olivinated cores are similar to those reported by Xie et al. (2012) in olivine cores inside ringwoodite rims in GRV022321 and support the idea of a solid-state transformation with diffusion controlled growth between olivine and ringwoodite.

Ideally, the approximate composition of ringwoodite can be inferred from the Raman peaks (Feng et al. 2011). The main peaks measured in the bulk of the ringwoodite grain are two major vibrations located at 798.8 and 843.2 cm⁻¹. Near the boundaries of the grain, the peaks are shifted to 797.5 and 843.5 cm⁻¹ near the majorite, and to 798.7 and 844.3 cm⁻¹ near the crack filed with olivine. This suggests slight differences in Fe and Mg ions content. Pure anhydrous γ-MgSiO₃ forms clear colored crystals with vibrations at 796 (RA1) and 835 cm⁻¹ (RA2) (Chopelas et al. 1994) while hydrous γ-(Mg₀.₈₆Fe₀.₁₄)SiO₃ 1.6 wt% H₂O is blue in color and the RA2 peak is shifted toward higher frequencies, 841 cm⁻¹. However, the position of the RA2 peak is not dependent on ringwoodite Fe-Mg contents despite ranging from 835 to 849 cm⁻¹ (Feng et al. 2011). The ringwoodite studied in the Taiban meteorite would be equivalent to 72–78% Fo according to the Raman calibration developed by Feng et al. (2011), however that calibration is not reliable near pure Fo values and is not consistent with earlier measurements (Chopelas et al. 1994; McMillan and Akaogi 1987). Nevertheless, the Raman calculated values in our study fall within ±6% Fo when compared to the EPMA results shown in Table 2.

The Raman pattern observed in Taiban is close to those of the ringwoodite γ-(Mg₀.₈₆Fe₀.₁₄)SiO₃ in the Tenham meteorite, 799 and 845 cm⁻¹ (Sato and Nakamura 2010), and of some of the ringwoodite in the lamellae of the Sixiangkou meteorite (22–24 wt% FeO), 799 and 844 cm⁻¹ (Chen et al. 2006). A possible explanation for this small shift found in the natural ringwoodite in these meteorites could be the presence of residual stress from the shock that triggered the transformation or the following relaxation. Sazonova et al. (2006) measured shifts of the RA1 peak up to 814 cm⁻¹ (814, 843; and 812, 846, in two different spectra) in ringwoodite containing Al up to 9 wt%, Ti up to 0.2 wt%, and K up to 1 wt% synthesized in a shock experiment from biotite. Thus, the presence of some ions dissolved in the ringwoodite, namely Al, Ti, or K, could also affect the position of the RA1 Raman line of ringwoodite. EPMA results show that the position changes can also be due to slight variations on Ti, Cr, Mn, and P in different points in the ringwoodite.

The cause of the blue ringwoodite color has been discussed in the past but it still remains mostly unknown. Lingemann and Stöffler (1998) and Price et al. (1979) observed glassy material around ringwoodite crystals in blue colored aggregates. However, this glassy material was not present in the colorless ringwoodite in the Sixiangkou meteorite (Chen et al. 1996). Iron has been ruled out as the cause of the blue coloration because of the absence of correlation between the ringwoodite color and the iron content (Coleman 1977; Lingemann and Stöffler 1998; Taran et al. 2009). Lingemann and Stöffler (1998) proposed that the blue color is due to the light-scattering effect produced by the glassy material, and Nagy et al. (2011) suggested that the color is due to scattering in grain boundaries in submicrometer grains. However, Taran et al. (2009) performed transmission electron microscopy on two foils from a colorless to dark blue ringwoodite grain and found no glassy phase in the colored areas.

The coloration of the ringwoodite in the Taiban meteorite varies gradually within the mineral grains. Colorless areas coexist with blue to dark blue zones forming bubbly or filament patterns inside otherwise completely homogeneous mineral grains. In our study we found a relation between the Raman spectra and the coloration of the ringwoodite. We assume the 877 cm⁻¹ peak has remained mostly unnoticed in previous Raman measurements (with the exception of Kleppe et al. 2002b and Nagy et al. 2010a, 2010b) because of the strong dependence of the Raman signal on the excitation wavelength. As shown in Figure 7, the 877 cm⁻¹ peak and the blue ringwoodite color. The arbitrary color scale shows the intensity of the peak between 860 and 900 cm⁻¹.
peak is not observable with 244 nm laser excitation. The peak is weak with respect to RA1 and RA2 with the widely used Ar 514.5 nm laser line and comparable in intensity when excited with 785 or 830 nm laser wavelengths. We attribute this intensity dependence to near-resonance Raman enhancement in the blue ringwoodite. Absorption spectra from dark and blue natural ringwoodite from NWA 1662 and NWA 463 meteorites present a strong absorption line centered at 793 nm (Taran et al. 2009). Synthetic blue ringwoodite presents similar absorption bands due to its iron content (Kepler and Smith 2005). In natural samples, the absorption is stronger in darker colored grains and weaker in colorless ringwoodite. This absorption band remarkably follows the same color correlation as observed in the 877 cm\(^{-1}\) Raman peak and explains why the enhancement is only observed in colored ringwoodite. The color correlation of the absorption spectra and the Raman spectra suggest that the cause of the color lies in the crystal structure of the ringwoodite. The absorption line was assigned by Kepler and Smith (2005) and Taran et al. (2009) to the electronic spin allowed \(^3\)T_g \rightarrow \(^3\)E_g transition of \(^3\)Fe^{2+} split by the trigonal distortion or Jahn-Teller effect. Similar Raman enhancements have been observed for iron containing oxides and oxyhydroxides, for which near resonance Raman occurs with 636 nm excitation due to an absorption band located near 640 nm; this band is a result of the \(\Lambda_5(S) \rightarrow \Lambda_1(G)\) ligand field transitions of octahedrally coordinated Fe\(^{2+}\) (Nieuwoudt et al. 2011; Sherman and Waite 1985). Resonance Raman has also been confirmed in several spinel structured oxides (Kashyap et al. 2009; Koshizuka et al. 1975; Lutz et al. 1991). Furthermore, the presence of unassigned peaks in Raman spectra from spinels due to disorder is not uncommon. The appearance of a new band in the non silicate spinel MgAl\(_2\)O\(_4\) after heating induced disorder has been described as a contribution to the Raman spectrum from cation disorder in the form of Al ion redistribution from octahedral to tetrahedral sites inside the spinel structure (Cynn et al. 1992, 1993). Thus, the peak we observed at 877 cm\(^{-1}\) is probably due to cation redistribution inside the spinel structure.

Taran et al. (2009) suggested an inverse spinel structure for the clear varieties of ringwoodite, based on the apparent different natures of the absorption spectra of blue and colorless varieties. In the inverse spinel, all Fe\(^{2+}\) ions and probably part of the Mg are found in tetrahedral sites of the cubic close-packed structure, and Si cations would be displaced to the octahedral sites. However, the observed Raman shifts of RA1 and RA2 do not significantly differ between colorless and blue ringwoodite, suggesting that most of the Si is in fourfold coordination. This is consistent with nuclear magnetic resonance (Stebbins et al. 2009) and X-ray diffraction observations (Hazen et al. 1993). On the other hand, the observed increase in the linewidths of RA1 and RA2 indicate some degree of structural disorder in blue ringwoodite. Therefore, our data are consistent with the explanation that the gradation in color of the ringwoodite is due to a spinel structure ranging from normal to marginally inverse. Assuming such cation disorder and that the band at 844 cm\(^{-1}\) comes from Si-O bonds stretching in tetrahedral sites (Yu and Wentzcovitch 2006), we tentatively assign the 877 cm\(^{-1}\) Raman band to symmetric Fe-O stretching of a small amount of iron in fourfold coordination. Because of the resonance Raman enhancement effect, we believe that we are able to detect low trace amounts of Fe\(^{2+}\) in tetrahedral sites. In the same structure, no Raman signal would be expected from Mg ions in four- or sixfold coordination due to the highly ionic nature of the Mg-O bonding. This position is relatively close to the Raman shifts of 830–832 cm\(^{-1}\) reported for FeO\(_4\) tetrahedra in K\(_2\)[FeO\(_4\)] by Gonzalez-Vilchez and Griffith (1972). The difference of approximately 45 cm\(^{-1}\) with respect to the \(\upsilon_5(A_1)\) band at 830 cm\(^{-1}\) could be due to contributions of the following: the coupling between SiO\(_4\) and FeO\(_4\) tetrahedra; induced distortion by the Si-O octahedra; or a shorter Fe-O bonding length, consistent with the high-pressure nature of the phase. Small bond distance differences can yield to large Raman shifts, e.g., in the transition between olivine and ringwoodite the increase of the Si-O distances of 0.02 ± 0.0005 Å (Hazen et al. 1993) induces a Raman shift of 10–25 cm\(^{-1}\) between the main peaks in olivine and ringwoodite. If the 877 cm\(^{-1}\) Raman band is produced indeed by Fe-O stretching, one would expect to also observe Fe-O bending modes and Si-O\(_4\) octahedral symmetrical stretching. The \(\upsilon_6\) and \(\upsilon_4\) modes of the FeO\(_4\) tetrahedral are reported to be in the region of 300–400 cm\(^{-1}\) (Gonzalez-Vilchez and Griffith 1972) and could possibly be the unassigned, enhanced bands we observed in that region (Fig. 5). The spectra of the SiO\(_4\) octahedral should resemble that of stishovite with two intense modes, \(\Lambda_{1g}\) at 753 cm\(^{-1}\) and \(\Lambda_{1g}\) at 231 cm\(^{-1}\) (Hemley et al. 1986). The band corresponding to the \(\Lambda_{1g}\) mode may be a shoulder of the RA1 ringwoodite band at 799 cm\(^{-1}\) (e.g., 765 cm\(^{-1}\), Fig. 5). Similarly, the band assigned to the \(\Lambda_{1g}\) mode could possibly contribute to the broad set of bands centered at 225 cm\(^{-1}\). The near-IR Raman enhancements of these bands (~231, 300–400, ~753, and 877 cm\(^{-1}\)) support these tentative assignments.

**Pyroxenes and majorite**

White grains show pyroxene Raman fingerprints (Fig. 8). The high-Ca clinopyroxene (augite/diopside, C\(_2\)/c monoclinic pyroxene) shows two intense peaks, the first one at 1013 cm\(^{-1}\) corresponding to the Si-O\(_6\) bond stretching in the [SiO\(_4\)]\(^2-\) unit of pyroxene chain, and the second peak at 667 cm\(^{-1}\) corresponding to the bridging oxygen stretch in the Si-O-Si bond of the [SiO\(_4\)]\(^2-\) unit. Three smaller peaks are present in the range 300–400 cm\(^{-1}\); 327, 369, and 395 cm\(^{-1}\) created by the displacements of cations from their equilibrium lattice positions or by or O-Si-O bending modes (Huang et al. 2000; Wang et al. 2001; Yang et al. 2009). A small amount of jadeite is also present in the white matrix. Jadeite has the ideal formula NaAl\(_2\)Si\(_2\)O\(_6\) and shows three main characteristic signals at 698, 377, and 204 cm\(^{-1}\). Similar to the high-Ca pyroxene, jadeite is a C\(_2\)/c monoclinic pyroxene and the 698 cm\(^{-1}\) band originates from Si-O-Si vibrations within the silicate chains, and 377 and 304 cm\(^{-1}\) are likely lattice modes involving cation-oxygen interactions or O-Si-O bending modes. Additional jadeite bands appear at 988, 1035 (Si-O stretching vibrations), and 756 cm\(^{-1}\) as well as other smaller bands in the range 200–600 cm\(^{-1}\) showing, respectively, cation vibrations and O-Si-O bending modes of SiO\(_4\) tetrahedra (Huang et al. 2000; Wang et al. 2001; Yang et al. 2009).

Along with the pyroxene crystals, majorite assemblages are present in the white matrix. The most intense modes of majorite are found at 597 and 929 cm\(^{-1}\) (Rauch et al. 1996). Majorite is the cubic garnet polymorph of pyroxene. It has previously been...
found in the Coorara, Catherwood, Pampa del Infierno, Tenham, Sixiangkou, and Peace River chondritic meteorites (Smith and Mason 1970; Coleman 1977). Majorite forms in these meteorites from low-Ca pyroxenes (Chen and Xie 2008; Coleman 1977; Mao et al. 1982; Smith and Mason 1970). Ca-rich majorite has been observed in the Shergotty and Yamato 75100 meteorites transformed from the host high-Ca pyroxene (Malavergne et al. 2001; Tomioka and Kimura 2003). Its low intensity is consistent with the observations for highly shocked plagioclase (Fritz et al. 2005). The spectra of both silicate glasses are shown in Figure 9. The spectrum is dominated by a broad set of bands characteristic of plagioclase feldspar glass, centered around 480 cm\(^{-1}\). This is in good agreement with the spectra reported by Sharma et al. (1983) for glass of anorthite composition and with earlier measurements of maskelynite from natural samples (Chen and El Goresy 2000; Fritz et al. 2005). This band has been attributed to the motion of the oxygen atom along a line bisecting the T-O-T angle (where T = Si or Al) and is characteristic of the feldspar structure, which contains four-membered rings of TO\(_2\) tetrahedra (Sharma et al. 1983). The weak shoulder located around 570 cm\(^{-1}\) has been assigned to three member rings of TO\(_2\) tetrahedra in the framework glasses (Galeener and Geissberger 1983; Sharma et al. 1997). Its low intensity is consistent with the observations for highly shocked plagioclase (Fritz et al. 2005). The spectra of plagioclase glass in the high-frequency range shows slightly different patterns when measured with different wavelengths. The Si-O stretch bands at 900–1100 cm\(^{-1}\) appear in the spectra recorded with 532 nm laser excitation. With 830 nm laser excitation a strong broad band appears at 990 cm\(^{-1}\) (904 nm). This band is probably due to rare earth elements fluorescence, likely Nd, whose strongest fluorescence band is also found around 907 nm. Lingunite, (Na,Ca)AlSi\(_2\)O\(_5\), is one of the high-pressure polymorphs of plagioclase. It possesses hollandite structure and was first observed by Gillet et al. (2000) in the Sixiangkou chondrite. Lingunite Raman peaks appear along with the glass spectrum in the maskelynite grains inside the melt vein, at 210, 274, and 763 cm\(^{-1}\). These peaks are characteristic of the hollandite structure: the signal at 763 cm\(^{-1}\) corresponds to SiO\(_2\) octahedral symmetrical stretching; and the peaks at 210 and 274 cm\(^{-1}\) correspond to the splitting of the B\(_{1g}\) mode. A band resulting from the splitting of the B\(_{2g}\) mode appears at 973 cm\(^{-1}\) (Gillet et al. 2000; Liu et al. 2004). In the Taiban meteorite pocket, the micrometer-sized jadeite crystals suggest that plagioclase melted and dissociated into jadeite and SiO\(_2\). The maskelynite in the host rock shows increasing deformation and flow textures toward the melt vein, consistent with total melting inside, where jadeite formed. The absence of the Raman signature peaks of SiO\(_2\) polymorphs can be explained if jadeite crystallizes from amorphous plagioclase while the crystallization of stishovite is significantly delayed (Kubo et al. 2010), leaving the SiO\(_2\) in form of glass.

**Maskelinite and lingunite**

Large amounts of plagioclase glass are found in the host rock areas compared to the relict plagioclase glass located in the margins of the silicate pocket, near the host rock. Plagioclase glass forms clear isotropic grains (Stöffler et al. 1991, their Fig. 12). The spectra of both silicate glasses are shown in Figure 9. The spectrum is dominated by a broad set of bands characteristic of plagioclase feldspar glass, centered around 480 cm\(^{-1}\). This is in good agreement with the spectra reported by Sharma et al. (1983) for glass of anorthite composition and with earlier measurements of maskelynite from natural samples (Chen and El Goresy 2000; Fritz et al. 2005). This band has been attributed to the motion of the oxygen atom along a line bisecting the T-O-T angle (where T = Si or Al) and is characteristic of the feldspar structure, which contains four-membered rings of TO\(_2\) tetrahedra (Sharma et al. 1983). The weak shoulder located around 570 cm\(^{-1}\) has been assigned to three member rings of TO\(_2\) tetrahedra in the framework glasses (Galeener and Geissberger 1983; Sharma et al. 1997). Its low intensity is consistent with the observations for highly shocked plagioclase (Fritz et al. 2005). The spectra of plagioclase glass in the high-frequency range shows slightly different patterns when measured with different wavelengths. The Si-O stretch bands at 900–1100 cm\(^{-1}\) appear in the spectra recorded with 532 nm laser excitation. With 830 nm laser excitation a strong broad band appears at 990 cm\(^{-1}\) (904 nm). This band is probably due to rare earth elements fluorescence, likely Nd, whose strongest fluorescence band is also found around 907 nm. Lingunite, (Na,Ca)AlSi\(_2\)O\(_5\), is one of the high-pressure polymorphs of plagioclase. It possesses hollandite structure and was first observed by Gillet et al. (2000) in the Sixiangkou chondrite. Lingunite Raman peaks appear along with the glass spectrum in the maskelynite grains inside the melt vein, at 210, 274, and 763 cm\(^{-1}\). These peaks are characteristic of the hollandite structure: the signal at 763 cm\(^{-1}\) corresponds to SiO\(_2\) octahedral symmetrical stretching; and the peaks at 210 and 274 cm\(^{-1}\) correspond to the splitting of the B\(_{1g}\) mode. A band resulting from the splitting of the B\(_{2g}\) mode appears at 973 cm\(^{-1}\) (Gillet et al. 2000; Liu et al.
FIGURE 9. Raman of plagioclase glass in the studied lithic pocket (a) and in the host rock (b) of the Taiban meteorite. Inside the melt vein (a), lingunite peaks appear along with the plagioclase glass Raman spectrum. Plots (A) and (B) show the spectrum of maskelynite inside the pocket (a); and (C) and (D) show the spectrum of maskelynite in the host rock (b). (A) 830 nm laser, 3 mW power output, 400 s acquisition time. (B) 532 nm laser, 3 mW power output, 300 s acquisition time. (C) 830 nm laser, 5 mW power output, 150 s acquisition time. (D) 532 nm laser, 3 mW power output, 300 s acquisition time. With 785 nm laser excitation spectrum was obscured by strong luminescence background.

According to the results of static high-pressure melting experiments of the Allende meteorite (Agee et al. 1995), the presence of crystals of ringwoodite, majorite, high-Ca pyroxene, and trues of wadsleyite can constrain the P-T formation conditions for the studied silicate pocket to 17–20 GPa with a T_{max} in the range 1850–1900 K. The presence of Fe-Ni droplets and various sulfides in droplets and filling cracks is also consistent with that P-T range. Our Raman and SEM observations in the studied area did not yield signals corresponding to higher pressure phases such as magnesiowüstite or silicate perovskite, thus the static pressure equivalent is probably below 20 GPa.

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The meteorite used for this study is the Taiban (b), L6-S6 ordinary chondrite. It was found in De Baca County, New Mexico in 1975. [Taiban (b) of Lange et al. 1980, catalog of the meteorite collection of the Institute of Meteoritics at the University of New Mexico. Special Publication 21, UNM Institute of Meteoritics.] Thin slices of the same meteorite were used in the work of Stöffler et al. (1991). This is not the Taiban (b) LL6 ordinary chondrite found in the same area in 1984 (Jeffrey N. Grossman, Meteoritical Bulletin 84, 2000), nor the Taiban meteorite found in 1934 (Grady, M.M. 2001, Catalog of Meteorites, 5th edition, Natural History Museum, Cambridge University Press). A new name must be assigned to the Taiban (b) meteorite used in this study to avoid further confusion.