Chinese Journal of Polar Science, Vol. 20, No. 2, 172 - 186, December 2009

Shock metamorphism in meteorites

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Received October 20, 2009

Abstract Shock metamorphism resulting from hypervelocity collisions between planetary bodies, is a fundamental processes in the solar system. The term "shock metamorphism" is used to describe all changes in rocks and minerals resulting from the passage of shock waves. Most meteorites have experienced collisions and have a record of shock metamorphism, which includes brecciation, deformation, phase transformation, local melting and crystallization. The key to reading this record is to use the shock features to estimate the pressure and duration of shock event. In this paper, the history of the study of shock metamorphism is reviewed, basic knowledge of shock physics is discussed, recent 10 years' studies of shock-induced melt veins are summarized, and finally a short note to the shock metamorphism in general is given.
Keywords shock metamorphism, meteorite, shock-induced melt vein, impact

1 Background of shock metamorphism

Studies of impact cratering, laboratory shock experiments and space exploration have contributed to our knowledge of shock metamorphism[1]. Lunar and terrestrial cratering studies suggest that the circular craters on the lunar surface and some terrestrial craters originated from impact[2,3]. The natural occurrence of coesite and stishovite confirmed the impact origin of Meter Crater, Arizona[4-6]. Laboratory shock wave experiments produced distinctive deformational effects in samples which are directly comparable to effects produced in rocks affected by natural impacts[7-13]. It became possible to calibrate shock pressures by comparison of natural and experimental shock effects. Planetary geology and space exploration, particularly after the Apollo program of 1970's, have shown that the impact cratering and the resulting shock metamorphism is one of the most fundamental processes in the solar system[14]. Meteorites, which have experienced collisions and have a record of shock metamorphism, provide the only record of the shock history of the early solar system.

Shock metamorphism was studied in terrestrial minerals such as coesite and stishovite found in Meteor Crater[4-6,8]. Shock deformation features in meteorites were studied intensively by optical microscopy and X-ray diffraction techniques in late 1960's[10,15-22]. Since the late 1970's, shock-induced deformation microstructures, such as planar deformation features (PDF), twinning and plastic deformation, have been studied intensively using transmission electron microscopy (TEM)[23-30]. The main goal of these studies has been to constrain shock conditions (pressure, temperature, and duration).

The widely used shock-classification scheme and pressure calibration for chondrites[31] is based on the comparison of deformation and transformation effects between natural and experimentally shocked samples. Based on shock effects in olivine and plagioclase as recognized in thin section, six stages of shock (S1 to S6) are defined by Stöffler *et al.* (1991). The shock pressure calibration defines the S1/S2, S2/S3, S3/S4, S4/S5, S5/S6 and whole rock melting transitions at < 5, 5—10, 15—20, 30—35, 45—55, 75—90 GPa, respectively[31]. Shock stage S6 (~50 to ~85 GPa) the highest shock stage below whole rock melting and commonly shows evidence of high-pressure minerals, such as ringwoodite. This shock classification system is easy to apply and correctly represents the progressive shock-pressure sequence from weak to strong. However, pressure calibration based on shock recovery experiments is problematic for some features, such as phase transformations that depend on reaction kinetics. For example, ringwoodite, the high-pressure polymorph of olivine and common indicator of S6 is stable between 18 to 22 GPa in static high-pressure experiments[32-35]. It has never been recovered from a shock experiment. Based on this lack of shock-produced ringwoodite, Stöffler *et al.* (1991) inferred that P> 50 GPa is required to form ringwoodite by shock. The transformation of olivine to ringwoodite during shock is both time and temperature dependent, therefore the duration of a shock experiment may not be sufficient for observable transformation.

Phase transformations that occur during shock are generally reconstructive and are strongly dependant on duration and temperature. The durations of shock experiments are short (nano- to microseconds) compared to large natural impacts (up to several seconds and even longer). Shock duration is critical for kinetic processes such as phase transformations, making direct comparison of natural and experimental shock transformation pressures inappropriate. Shock temperature and internal energy input in shock experiments using multiple-reflection loading path are considerably lower than those in single shock events (natural cases)[36-39]. Shock effects on porous, multiphase materials are extremely heterogeneous and the amount of heat input is much larger than in nonporous materials[12, 40]. Pre-shock temperature also affects onset pressure for different shock features[41]. In high-temperature shock experiments, the transformation of oligoclase to glass, the onset of mosaicism in orthopyroxene, the recrystallization or melting of olivine, and the onset of shock-induced localized melting start at lower shock pressures in comparison to low-temperature shock experiments. All these kinetic factors limit the use of pressure calibrations based on shock experiments.

An alternative means of investigating shock pressure in natural samples is to use the mineralogy of melt veins to estimate crystallization pressure based on phase equilibrium data from static high-pressure experiments[38, 42-50]. The assemblage of majorite + magnesiowüstite in the melt vein of Sixiangkou L6 chondrite was recognized as the stable liquidus assemblage at approximately 23 to 27 GPa and 2000 ℃[38, 42-50]. This assemblage, commonly observed in heavily shocked meteorites, is the same liquidus assemblage synthesized between 23 and 27 GPa and 2000 ℃ in high-pressure experiments on the carbonaceous chondrite Allende[51], and the same assemblages synthesized in high-pressure melting experiments on KLB-1 peridotite[52]. Close similarities in mineralogy, grain size, composition and microstructures between the experimental samples and melt-vein material in meteorites suggest that the results of static high-pressure experiments can be used to estimate the conditions of melt-vein crystallization.

The high-pressure minerals in the shock-induced melt veins of meteorites not only provide pressure indicators for shock events, but also provide natural samples of deep Earth materials. Shock-induced melt veins commonly contain two lithologies. One consists of polycrystalline grains that transformed from host rock fragments by solid-state mechanisms. The other consists of quenched silicate grains and metal-sulfide grains that crystallized from the melts. Both lithologies in highly shocked (S6) meteorites commonly contain high-pressure minerals, which are the same as those expected in Earth's transition zone (410 to 660 km depth) and lower mantle. These minerals, including wadsleyite, ringwoodite, majorite, akimotoite, silicate perovskite, hollandite-structured plagioclase, post-stishovite SiO_2 polymorphs, and Fe_2SiO_4-spinel, are not found in ordinary crustal rocks, but occur naturally in close association with melt veins in shocked meteorites[45, 46, 53-61]. Hollandite-structured plagioclase[62, 63] has recently been discovered in association with a large impact crater. Other occurrences of these mantle minerals in unusual settings are construed as evidence either of impact or mantle origin. The melt veins of meteorites provide a unique natural high-pressure laboratory for Earth-mantle minerals.

2 Shock physics

An understanding of basic shock physics is essential for interpreting shock effects in meteorites. The fundamental physics of shock wave propagation have been described in a number of articles and textbooks[13, 18, 44, 64-67]. Here I introduce the basic concepts and principles of shock waves in solids.

A shock wave is defined as the propagation of a discontinuity of the thermodynamic and mechanical properties of the medium: pressure, density, energy, temperature, and material velocity[67]. A stress wave propagating with supersonic velocity is formed when the surface of a solid body is rapidly accelerated, for example, by the impact of a projectile or by a chemical or nuclear explosion[18]. Since the compressibility of a solid generally decreases with increasing pressure, the stress wave will immediately steepen to a shock wave, which represents a discontinuity in pressure (P), density (ρ) or specific volume (V) where $V = 1/\rho$, and internal energy (E). The material behind the shock front is compressed to a higher density (ρ) and its constituent particles are accelerated to high velocity (U_p) (Fig. 1).

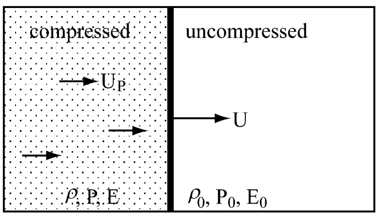

Fig. 1 Schematic of quantities describing the shocked state of the medium, such as particle velocity U_P, density ρ, pressure P, and internal energy E (per unit mass), jumping discontinuously across a shock front (modified after Melosh, 1989). The shock front propagates at velocity (U).

Shock pressures are relieved by release waves (also called rarefaction waves). Rarefaction waves are acoustic and propagate at a velocity inversely proportional to material compressibility. Since the compressibility increases as the pressure decrease, one generally refers to a "rarefaction fan" which cause pressure release to be more gradual with increasing propagation distance. The release wave originates from a free surface such as the rear side of a projectile, and gradually overtakes the shock wave causing a decrease of peak pressure and shock wave velocity. The site near the impact interface has a longer high-shock-pressure duration and a steeper release, while the site far away from the impact interface has a shorter high-pressure-shock duration (Fig. 2), and a broader release.

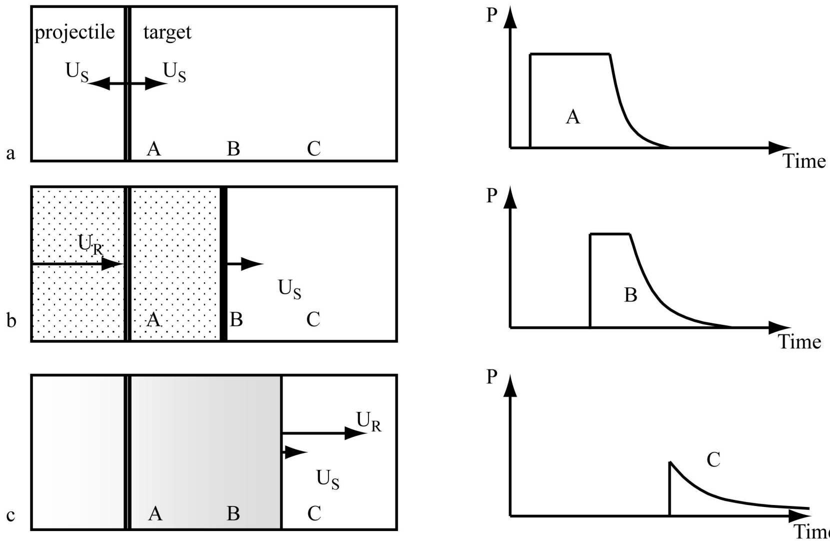

Fig. 2 Simplified schematics illustrating the production of shock and rarefaction waves in a planar projectile and target. a) At the time of impact, shock waves are generated simultaneously in the projectile and target, and travel at the velocity of the shock wave (Us), in opposite directions. b) When the shock wave in the projectile reaches the back free surface, it is reflected back into the projectile as a rarefaction wave. The dotted area represents compressed material. c) The release wave (UR) travels faster than shock wave through compressed material. When the release wave (UR) catches the shock front, the peak pressure starts to decrease. Location (A) is close to the shock interface and has a flat-topped and long shock pulse. B) Further away from the interface, the flat-topped portion is shorter. C) Far from the interface, the release wave has caught up to the shock wave and there is no flat-topped portion.

P. H. Hugoniot (1887, 1889)[68, 69] derived the fundamental equations of the conservation of mass, momentum, and energy to relate the quantities on either side of the shock front. Assuming a plane wave geometry and hydrodynamic conditions (absence of material strength), the pressure P, specific volume V, particle velocity U_P, shock wave velocity U and internal energy E in the shock state are related to the corresponding parameters P_0, V_0, E_0 in the un-shocked material by the Hugoniot equations.

$\rho(U - U_P) = \rho_0 U$ (1. Conservation of Mass)

$P - P_0 = \rho_0 U_P U$ (2. Conservation of Momentum)

$E - E_0 = (P + P_0)(V_0 - V)/2$ (3. Conservation of Energy)

The third equation is called Rankine-Hugoniot relation. This relation describes the locus of all shock states (P_1, V_1, E_1) achievable by shock waves of various intensities in a particular solid from the initial state (P_0, V_0, E_0). The graphical representation of this locus in the pressure - volume profile is termed the Hugoniot curve (Fig. 3).

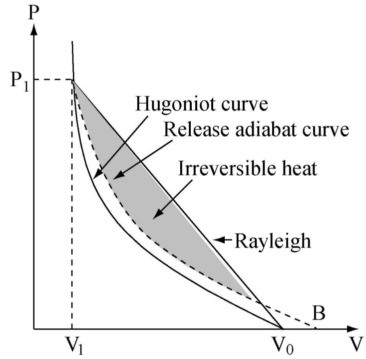

Fig 3 Hugoniot curve and expansion adiabat line for shock state A (P_1, V_1) for ideal hydrodynamic conditions, modified after Stöffler (1972)[18]. The gray area between the release adiabatic curve and the Rayleigh line represents irreversible heat (also called waste heat) produced by a shock wave with peak pressure P_1.

The Rankine-Hugoniot equation determines the internal energy change in the compressed material, which can be shown graphically in the pressure-volume plane (Fig 3). The material shocked to state A (P_1, V_1) is compressed along the Raleigh line (line V_0A), and decompresses along a release adiabat (curve AB). The work done by the shock compression equals the area of the triangle $V_1 - V_0 - A$ in Fig 3. The energy converted by decompression equals the area between the release adiabatic curve and lines AV_1, V_1B in Fig 3. The difference between these two areas is equal to the amount of irreversible work done by the shock. The irreversible work results in post-shock heat (waste heat), which increases with increasing peak shock pressure. Consequently, material can be melted or even vaporized after shock release if the shock pressure is high enough[64]. The waste heat is easy to estimate if the release adiabat is known or can be approximated by the Hugoniot curve. In general, the release adiabatic curve closely approximates the Hugoniot curve, except that it decompresses to a slightly larger specific volume than the initial volume. This difference is due to thermal expansion of the hot decompressed material[13]. The shock and post-shock temperatures can be calculated with the aid of equation-of state data for common minerals and rocks.

In order to understand naturally produced shock metamorphic effects, it is necessary to reproduce them under known conditions (P, T, t) in laboratory shock experiments. Among the 3 thermodynamic variables (P, V, E) and the 2 kinematic variables (U_P, U) which are in the conservation law equations, only the kinematic variables U and U_p can be obtained from reliable measurements. The pressure (P) and specific volume (V) can be calculated via the Hugoniot equations from measured velocity (U_P and U_S). This procedure leads to the determination of the Hugoniot curve and finally to the equation of state (e. g., entropy, shock temperature). The data of Hugoniot equation of state for rocks and minerals have been collected since World War II[70]. Shock loading experiments have been well developed after World War II[1]. The most common shock loading techniques include laser and electron-beam loading, exploding foil techniques, gun techniques and high explosive systems. Gun techniques are most popular in recent literature. They produce a plane shock and easily are applied in shock recovery experiments for the study of residual shock effects. The relation between shock effects and peak shock pressure has been explored experimentally resulting in a pressure calibration for natural shock effects For more detail and deep discussion, please check reference of Sharp and Decarli 2007.

3 Shock-induced melt vein

Melt veins in meteorites have been known for many years, but the shock-induced origin was unknown until Fredricksson *et al.* (1963)[11] first synthesized black veins in shock experiments Melt veins have been described as "polymict breccias"[71], "shock-induced veins"[11], "shock blackening vein", "polymineralic mixed melts", "shock-induced localized melting" and "S-type pesudotachylite". These veins are actually sheets, ranging in thickness from ~ 1 μm to several mm, which appear vein-like in thin section These veins represent material that was locally melted and quenched by conduction to surrounding cooler material.

Studies of the mineralogy of shock-induced melt veins provide evidence relevant to the shock history of a meteorite. The crystallization pressure and temperature of the melt veins are constrained by the observed mineral assemblages together with phase relations obtained from static high-pressure experiments The shock pulse duration can be constrained by calculations of quench time of a melt vein. The key to using crystallization pressure of the melt vein minerals to infer the shock pressure is to understand when the melt vein crystallized The key question is does crystallization pressure correspond to peak shock pressure or to some pressures along the adiabatic release path

3.1 *The formation and crystallization model of melt vein*

The exact mechanism of melt vein formation is unknown, but several mechanisms are possible Shock-wave interactions between different shock impedance materials may cause localized melting which is most pronounced at the interface of metal-troilite and silicates, and the interface of minerals and pore space[12, 31, 41]. Friction by shock shearing along the contacts of materials of vastly contrasting shock impedance and along fractures may produce local melting[40, 72-75]. Adiabatic shear can produce high temperatures that are thousands of degrees hotter in shear regions than in immediately adjacent material[73]. In the high-temperature regions, transformation to high-pressure phases or to a liquid form of the high-density phase is possible, even on a sub-microsecond time scale, by a conventional thermally activated nucleation and growth process[73].

We treat the meteorite as if it were made up of two components matrix and veins The peak temperature of the melting vein must have been above the liquidus at a certain pressure The quenched high-pressure phases present in the narrow veins can provide information on the peak pressure. With knowledge of the peak pressure, the Hugoniot of the meteorite, and the heat capacity, we can calculate the shock temperature for the whole rock Since the veins are only a small fraction of the total volume, we assume that the calculated whole rock temperature is equal to the matrix shock temperature The hot veins must cool by conduction to the surrounding cooler matrix Typical time-temperature-pressure profiles are shown in Fig 4

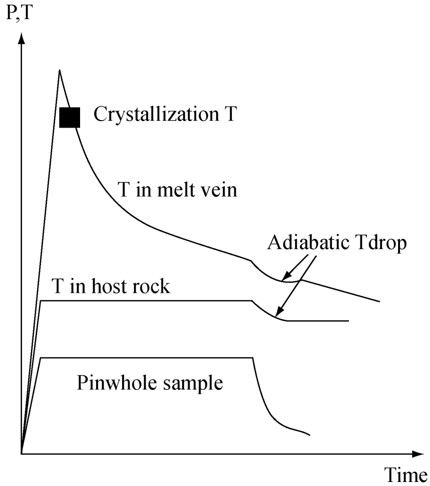

Fig. 4 Schematic pressure and temperature vs. time profile illustrates a heterogeneous temperature and a homogenous pressure scenario. The adiabatic temperature drop due to pressure release is relatively small.

Although the pressure distribution in a multi-phase sample is heterogeneous on the scale of the grain size as the shock wave passes, pressure differences quickly dampen out to an equilibrium shock pressure. This heterogeneity is caused by differences in shock impedance among different mineral grains. However, these pressure variations are dampened by shock-wave reverberations between the grains of varying shock impedance. Assuming a shock wave velocity of ~ 10 km/s, reverberations between 100 μm grains or veins occur on a (10^{-4} m/10^{4} m/s) 10^{-8} s (10 ns) time scale, resulting in pressure equilibration after several reverberations. However, the duration of shock pulse is proportional to the size of the projectile[13], t≈ d/v, t is shock duration, v is impact velocity, d is diameter of shock projectile. For a one-km projectile body and a impact velocity of 10 km/s, shock pulse duration is ~ 100 millisecond (10^{-1} s) near the impact site. The initial pressure fluctuations on the scale of the grain size are generally neglected.

The temperature difference between the melt vein and host rock determines the rate at which the melt vein quenches[43, 50, 76]. The large temperature differences between the melt-vein and host-rock results in conduction of heat from the melt vein to the host rock. The melt vein quenches by thermal conduction with surrounding relatively cooler host rock, rather than by pressure release. Quenching begins as soon as the vein forms, independent of the pressure pulse duration. The temperature drop due to adiabatic pressure release is almost negligible compared to the temperature drop due to thermal conduction. For a modest pressure of 25 GPa, the adiabatic cooling of a chondrite such as Tenham is less than 100℃, and does not drive the quench of the melt. Crystallization of the melt vein starts after the melt-vein temperature drops below the liquidus, which for a chondrite is around 2000℃ at 25 GPa[51]. The solidification of the melt vein starts from the vein edge and ends in the vein center. The time required to quench a whole melt-vein can be estimated with thermal modeling. We use a finite element heat transfer (FEHT) program to calculate temperature – time profiles from vein center to vein edge[77].

We can think of crystallization in terms of three scenarios (Fig. 5): (1) If the shock pressure pulse duration is longer than quench time, the melt vein may crystallize at constant equilibrium shock pressure; (2) If the pressure pulse duration is similar to or somewhat shorter than quench time, the melt vein crystallization may occur during pressure release; (3) If the shock pressure pulse is much shorter than quench time, the melt vein crystallization can occur after pressure release. These scenarios should be reflected in the melt-vein mineralogy. If crystallization occurs at constant pressure, we should see the same assemblage throughout a given melt-vein. If crystallization occurs during pressure release, we should see an assemblage change from the vein edge to vein center, with a lower pressure assemblage in the vein center. Therefore, if we see the same high-pressure assemblage throughout the vein, we can conclude that the crystallization pressure represents the equilibrium shock pressure. If we see the same low-pressure assemblage throughout the vein, we can conclude that crystallization occurred after pressure release. If we see a lower-pressure assemblage in vein center and higher-pressure assemblage in the vein edge, we can conclude that crystallization occurred during pressure release.

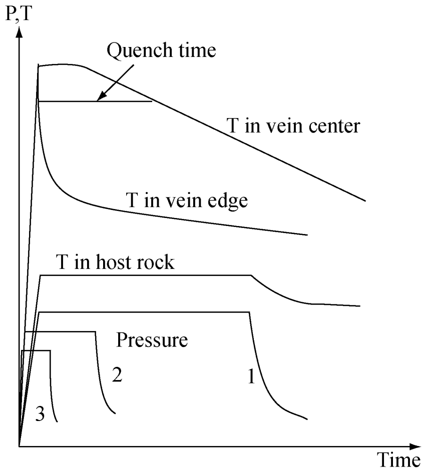

Fig 5 Pressure and temperature vs time profiles are shown in three quench scenarios (1) Quench at high shock pressure; (2) Quench involving pressure release; (3) Quench after pressure release.

3.2 Host rock fragments in melt veins

The melt veins are analogous to pseudotachylites in that they comprise sheets of locally shock-induced melted material in which unmelted host-rock fragments are commonly entrained (Fig 6). The melt veins generally contain two distinct parageneses. One consists of a matrix of silicates, metals and sulfides that crystallized from immiscible silicate and sulfide melts. The other high-pressure paragenesis in the melt vein consists of polycrystalline grains, produced by solid-state transformations of host-rock fragments that were entrained in the melt veins. Figure 8 and Figure 9 show the ringwoodite polycrystalline and hollandite polycrystalline which transformed by solid-state transformation from host-rock fragments in the shock-induced melt vein.

High-pressure minerals are common in and around melt veins in highly shocked meteorites. They can form either by crystallization of silicate melt or by solid-state transformation

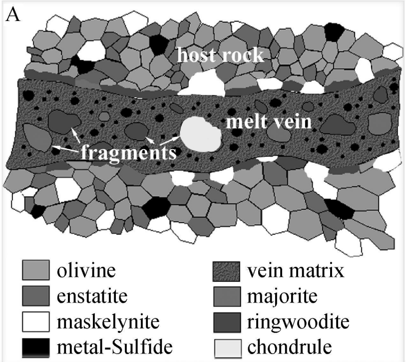

Fig 6 Schematic cartoon of a shock-induced melt vein with host-rock fragments embedded in the melt vein matrix. Widths of melt veins range from several μm to one mm. The vein matrix (dark) commonly contains sub-micrometer silicates and solidified metal-sulfide blebs.

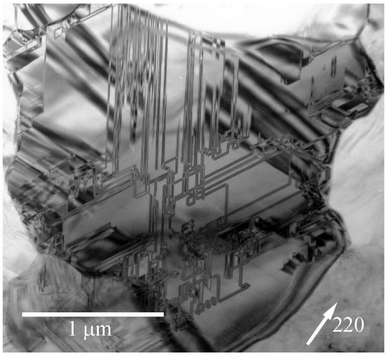

Fig 7 Bright-field TEM image of a ringwoodite crystal in a large polycrystalline aggregate from Sixiangkou. The image was obtained using g= 220, which highlights the distinctive stacking faults on (110). Referred from (Chen *et al* 1996)[42].

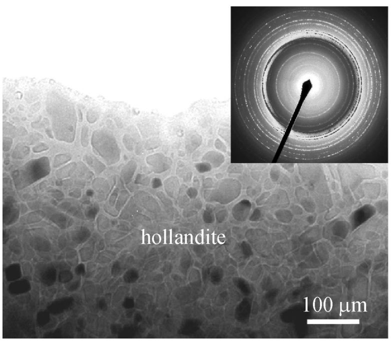

Fig 8 Bright-field TEM image of nano-crystalline hollandite from Tenham. The individual crystallites range in size from 20 to 100 nm. The selected-area electron diffraction pattern contains diffraction rings that confirm the hollandite structure and indicate random orientations of the grains.

of host-rock fragments entrained in the melt vein during shock. However, most early studied high-pressure minerals are host-rock fragments because they are bigger and easy to be observed. More recent work concerning high-pressure minerals in meteorites[28, 38, 45, 48, 78-80] have focused on the host rock fragments too. The focuses are back to interpreting shock conditions and durations using the growth kinetics of the high-pressure polycrystalline.

3.3 *Constraints on shock pressure from melt-vein crystallization*

Using melt-vein crystallization to estimate shock pressure is controversial in the field of shock metamorphism because it uses phase equilibrium data obtained in static high-pressure experiments (Fig. 9). However, there are several reasons why high-pressure melting relations can be applied to the interpretation of melt-vein crystallization. First, the most common melt-vein assemblage seen in S6 chondrites, majorite plus magnesiowüstite (Fig. 10), is also produced in static high pressure melting experiments on both Allende[51] and on Kilburn-hole-1 peridotite[52]. The textures and crystal sizes in the centers of large chondritic melt veins, such as those in Tenham, Sixiangkou and Acfer 040[42, 81, 82], are very similar to the textures and crystal sizes produced in the static experiments[51]. Similarly, the chemical compositions of the crystallized majoritic garnets are very similar to the compositions of garnets in the experiments. Compared to solid-state reconstructive phase transitions, melt-vein crystallization involves much smaller kinetic barriers.

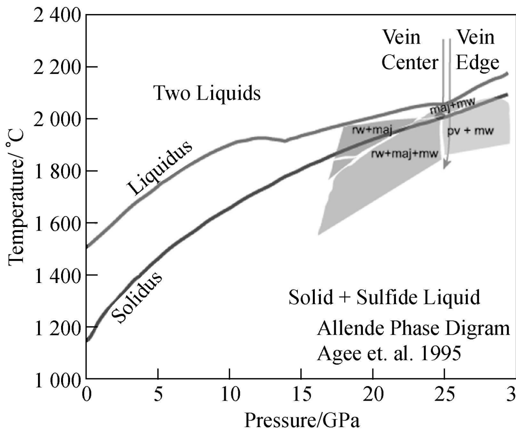

Fig. 9 The crystallization-pressure regions illustrated on a simplified version of the Allende phase diagram[51]. rw = ringwoodite; maj = majorite; mw = magnesiowüstite; pv = perovskite.

Melt-vein crystallization has a great advantage over solid-state transformation for constraining shock pressure histories. Because the cooling produced by adiabatic pressure release is relatively small, shock melt cools predominantly by conduction to the surrounding, relatively cool, host meteorite. This results in crystallization that starts at the vein margins and moves inward to the melt-vein core as crystallization proceeds. The resulting crystallization sequence provides a record of shock pressure through time. This record can be several hundred milliseconds long. If the recorded pressure-temperature-time history exceeds the period of elevated pressure, crystallization assemblages should record the pressure release.

The crystallization assemblages in a given melt vein will depend on the time required for melt vein quench versus the duration of high shock pressures (Fig. 5). If the shock duration were longer than the crystallization time of the melt vein, then we would expect crystallization to have occurred during the period of high shock pressure and therefore record the continuum shock pressure. This appears to be the case for S6 samples such as Tenham (Fig. 10)[50], RC106[81] and Sixiangkou[42]. If the shock duration was the same duration as melt-vein crystallization, then it is likely that melt-vein crystallization would record both the continuum shock pressure and a lower pressure of partial release such that the core of the vein might contain an assemblage that crystallized at a lower pressure than that of the rest of the vein. This is not common. If the shock pulse were shorter than the melt-vein quench time, one would expect crystallization of low-pressure assemblages in the core of the melt vein. This appears to be the case for S4 samples Kunashak and La Lande, which contain plagioclase-bearing crystallization assemblages[83]. Finally, we note that some veins may form at low pressure during pressure release. The mineralogy of these veins would be unrelated to either the magnitude or duration of peak shock pressure.

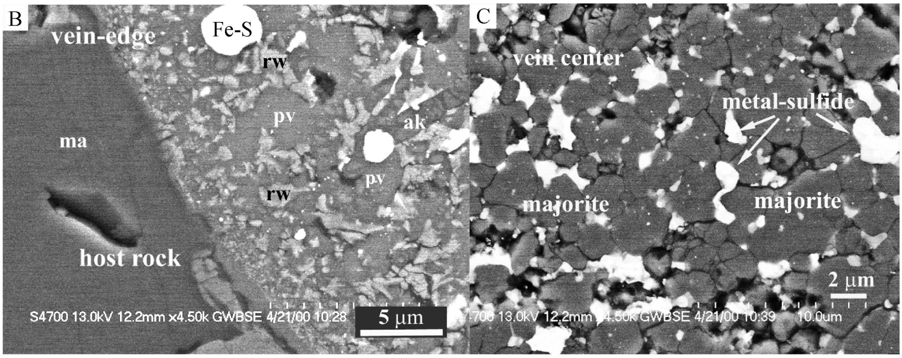

Fig. 10 Field-emission SEM images of a melt vein in Tenham. The melt-vein margin on the left contains ringwoodite (rw), akimotoite (ak) and vitrified silicate perovskite (pv) along with solidified droplets of Fe-sulfide melt. The vein core on the right contains the common assemblage of majoritic garnet and magnesiowüstite along with blebs of solidified metal-sulfide melt.

4 Shock metamorphism in general

The review of the study of shock metamorphism and the basic shock physics is one general topic for all shock features. The shock-induced melt vein of meteorites only represents the shock features at relatively higher shock degree and one special topic of shock metamorphism in recent years. Actually there are many interested shock features worth to study in detail, such as planar deformation feature, and shock-induced breccia at lower shock degree, or shock-induced whole rock molten, shock-induced vaporization and condensation at much higher degree.

In addition, there are many interesting topics relating to shock metamorphism were not covered here, such as the impact model, impact crating mechanism, shock effects on different minor minerals (like troilite, iron-nickel metal, etc.), shock effects on different meteorites, shock effects on planetary bodies (such as Mars, the Moon, etc), shock effects on terrestrial rocks, terrestrial impact effects on the evolution of the life and geology history, evaporation and condensation by shock wave in the beginning nebular stage, and so on. In general, shock wave and shock metamorphism play an important role in our solar system.

Acknowledgements Thanks to my Ph D. supervisor Prof Thomas Sharp at Arizona State University and shock-physics expert Paul Decarli at SRI International for their wonderful teaching and guidance to the wonderful shock field Also thanks to Prof Wang Daode and Wang Henian for their support and encouragement to finish this paper

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