

Understanding the textures and origin of shock melt pockets in Martian meteorites from petrographic studies, comparisons with terrestrial mantle xenoliths, and experimental studies

Erin L. WALTON^{1*} and Cliff S. J. SHAW²

¹Department of Earth and Atmospheric Sciences, University of Alberta, 1-26 Earth Sciences Building, Edmonton, AB T6G 2E3, Canada

²Department of Geology, University of New Brunswick, 2 Bailey Drive, Fredericton, NB E3B 5A3, Canada

*Corresponding author. E-mail: ewalton@ualberta.ca

(Received 10 February 2008; revision accepted 17 July 2008)

Abstract—We present a textural comparison of localized shock melt pockets in Martian meteorites and glass pockets in terrestrial, mantle-derived peridotites. Specific textures such as the development of sieve texture on spinel and pyroxene, and melt migration and reaction with the host rock are identical between these two apparently disparate sample sets. Based on petrographic and compositional observations it is concluded that void collapse/variable shock impedance is able to account for the occurrence of pre-terrestrial sulfate-bearing secondary minerals in the melts, high gas emplacement efficiencies, and S, Al, Ca, and Na enrichments and Fe and Mg depletion of shock melt compositions compared to the host rock; previously used as arguments against such a formation mechanism. Recent experimental studies of xenoliths are also reviewed to show how these data further our understanding of texture development and can be used to shed light on the petrogenesis of shock melts in Martian meteorites.

INTRODUCTION

In igneous petrology, textures are key to the interpretation of petrogenesis. Regardless of the planetary body on which they occur, we expect, on the basis of thermodynamics, that processes such as melting, crystallization and dissolution will be the same for systems of similar chemical composition. If this is the case, then the resulting textures should also be the same if the processes act over the same time scale. The ideas presented in this paper stem from the observation that textures in shock melt pockets found in Martian meteorites are very similar to those observed in terrestrial mantle-derived peridotites. In both meteorites and peridotites, the features of interest are the result of transient phenomena—the melt responsible for texture development is present in a liquid or reactive form for a short period (seconds to hours) before it crystallizes, or quenches to a glass. The textures of interest are developed because the melt is not in equilibrium with earlier-formed igneous minerals, in particular olivine, pyroxene and spinel. In this study, we demonstrate through petrographic observations, experimental studies and comparison with terrestrial mantle xenoliths, that the interaction between melts generated by shock metamorphism and meteorite minerals can be used to constrain the origin of shock melt

pockets, as well as aiding in the understanding of texture development in these features.

The recognition that rocks can be accelerated to escape velocity without vaporization or complete melting has had a profound effect on impact physics (Melosh 1985). Studies of the shock features recorded in Martian meteorites have placed constraints on impact ejection models (e.g., Melosh 1985; Artemieva and Ivanov 2004). The melt pockets are of particular interest to students of shock metamorphism because similar features, although documented in other meteorite types (e.g., L-group chondrites) (Dodd and Jarosewich 1982), have not yet been observed in terrestrial impact craters, nor have they been produced in shock-recovery experiments. The pockets also relate to studies of the Martian atmosphere and the Martian surface, because they have been found to contain a sample of the contemporary atmosphere of Mars and Martian soil/alteration products (Bogard and Johnson 1983; Gooding and Muenow 1986; Artemieva and Ivanov 2004). Some shock melts (veins and small pockets) contain ferromagnesian and Al-bearing silicate minerals stable at temperatures and pressures of the Earth's transition zone and lower mantle (Beck et al. 2004) and so are also of interest to mantle petrologists. Shock metamorphism has also been shown to disturb the radiogenic isotope systematics of a meteorite (Borg et al. 2005). An

understanding of the type and degree of shock damage, and how isotopic systems (U-Pb, Sm-Nd, K-Ar etc.) respond to this damage, can aid in choosing the most suitable materials for dating (see current arguments over the age of shergottites meteorites; Borg et al. 2005; Bouvier et al. 2008). Despite the importance of melt pockets in Martian meteorites, the formation mechanism remains unresolved, though there are two schools of thought—both mechanisms involve melt formation during shock metamorphism, but invoke a different origin for the melt. The first mechanism envisions pocket formation by injection of externally generated melts into the target rock along cracks and fractures, the second, in situ melting of the target rock by collapse of void space. Bogard et al. (1986, 1989), and Wiens and Pepin (1988) originally proposed these two mechanisms as possible scenarios to account for observed gas emplacement efficiencies measured in Elephant Moraine (EET) 79001 shock melts. Later proponents of an injection mechanism invoke a mixing model involving variable proportions of three end members: host rock mineral fines, plagioclase and Martian soil (Rao et al. 1999) to explain the bulk properties of shock melts in EET 79001 (compositional anomalies and noble gas content). As an alternative Beck et al. (2007), Heider and Kenkmann (2003), and Stewart et al. (2007) have all stressed the importance of the role of density discontinuities in rocks (fractures, grain boundaries, crystal defects, etc.), which in turn affects how the rock responds locally to shock. In this scenario, melt veins and pockets are the result of in situ localized melting due to shock impedance (shock wave velocity $\times$ phase density) contrasts, where, in the most extreme case, collapse of void space in the target rock gives rise to the highest dissipation of energy. The bulk properties of the melting products produced from collapsed voids will be dominated by the local mineralogy, and whatever materials were in the void prior to shock. Although only the work of Beck et al. (2007) refers specifically to the pockets in Martian meteorites, the results of Heider and Kenkmann (2003) and Stewart et al. (2007) apply to any polyminerale rock containing open (void) space. Moreover, the importance of shock impedance contrasts in generating local stress and temperature concentrations at grain/lithological boundaries, or at the free surface of a preexisting void space due to shock wave reverberations, has long been attributed to the formation of localized melts in chondrites (see reviews by Stöffler et al. 1991; Bischoff and Stöffler 1992 and references therein).

Terrestrial mantle-derived xenoliths have been the subject of considerable research because they are an essential source of information about the nature and evolution of the lithospheric mantle. Much of our understanding of the mantle is based on observations of depleted mantle xenoliths that have suffered variable degrees of melt extraction and enriched mantle xenoliths that have suffered enrichment by mantle metasomatism. However, several recent petrologic studies (Shaw and Klügel 2002; Shaw et al. 2006) have

shown that xenoliths may reside in crustal magma chambers for significant periods of time prior to eruption. During such residence periods, xenoliths are subject to significant textural modification through interaction between magma and xenolith minerals that may mimic the effects of mantle metasomatism. To avoid erroneous interpretations of the state of the mantle, its mineralogy and chemical composition, it is pertinent to identify those textures that are of mantle origin versus those imposed on the xenoliths during magma transport/residence. In addition to separating mantle and transport related processes, such studies have shown that it is possible to use textural and associated chemical data to understand the residence time and therefore the transport times of mantle xenoliths in a magmatic system (Klügel et al. 2000; Shaw and Klügel 2002; Shaw 2004).

Mantle xenoliths and Martian meteorites present similar problems for petrologic interpretation. In both cases, the samples are from much larger, otherwise inaccessible bodies, and they have been subjected to violent phenomena, which have left a record in the rocks in quenched reaction textures. In this paper, we present a review of textures from terrestrial and extraterrestrial samples focusing on the petrography of reaction zones formed by the interaction of melt with earlier-formed igneous minerals. Although the melts form by different means (mantle-derived melts versus shock-induced melting), the thermodynamics of the reactions will be the same. Thus, if the samples are quenched before equilibrium is attained we might expect to see similar textures. Here, we note that the two sample sets compared in this study (mantle xenolith reaction textures versus meteorite shock melts) may have formed over different time-scales, thereby limiting quantitative conclusions; however, the qualitative similarity of reaction textures is striking. Although results from mantle xenolith studies (reaction times etc.) may not be directly applied to the meteorites, it is demonstrated that a similar experimental and analytical approach, used to gain a better understanding of the origin and alteration of mantle-derived peridotites, can also be applied to shock melts in Martian meteorites. The goal of this study is to address two outstanding issues in Martian meteorite research that need to be resolved, namely: 1) shock melts in shergottites exhibit complex textures, some of which have not been well explained, and 2) the origin of the localized shock melt pockets is controversial.

Since this study is primarily concerned with texture, it is useful to define the terms used to describe features in peridotites and meteorites. Rounded cusped contacts between crystals and melt are depicted in this study that are very similar to cellular textures developed in alloys (Federov 1995). This term [cellular] has thus been adopted here to describe this texture. Sieve texture is a reaction texture in which an inclusion-free mineral core is completely or partially surrounded by a crystalline, vermiform rim with abundant melt inclusions (quenched to glass). Pockets of

crystallized/quenched glass in the meteorites are referred to as “melt pockets” to distinguish them from the “shock veins” described by (Langenhorst and Poirier 2000) in the same meteorites. The latter veins shear and displace adjacent minerals, exhibit sharp, discordant contacts with the host rock and are interpreted to form by frictional melting. The products of crystallization or quenching of the melt pockets are glass or glass + crystals. These melt products are labeled as “melt” in figures. Secondary textures in mantle peridotite xenoliths are located both at the xenolith—host melt contact and along veins of infiltrated host as well as secondary melts. Polyminerals rinds on orthopyroxene in xenoliths, where they are in contact with a silica-undersaturated melt, are referred to simply as reaction zones. It should be noted that Martian meteorites also contain “melt pockets or inclusions” enclosed by igneous minerals (typically olivine, pyroxene and spinel) (Goodrich 2003). These inclusions are more correctly referred to as “magmatic inclusions,” as they represent evolved liquids trapped in igneous minerals during their growth and crystallization from the parent melt, and are not related to shock metamorphism (some magmatic inclusions have been overprinted by later shock melting; Fig. 6, Walton et al. 2005).

NATURAL SAMPLES

In this study, we describe pockets of glass and crystals, and reaction zones associated with them that are restricted to strongly shocked Martian meteorites (20–55 GPa; Fritz et al. 2005). These pressure estimates are most reliably obtained by measuring the shock-induced reduction of the refractive index of plagioclase (Stöffler et al. 1986); however, shock pressure can also be estimated by comparing observed shock effects in meteorites with those accurately calibrated from a comprehensive set of shock recovery experiments on single crystals, terrestrial rocks and chondrites. The pockets are interpreted to be the quenched remains of local hot spots generated by shock melting (up to 2500 K), most likely associated with the impact events that also launched the rocks into Earth-crossing trajectories from the Martian near surface. The mineral assemblage of some pockets is indicative of crystallization at high temperature and pressure, confirming their impact origin (Beck et al. 2004). We report observations of melt pockets in Los Angeles, Dar al Gani (DaG) 476, Sayh al Uhaymir (SaU) 150, Elephant Moraine (EET) 79001, Allan Hills (ALH) 77005, Northwest Africa (NWA) 1950, NWA 4468, and NWA 1068.

In peridotites, reaction textures are common where melt films and pockets of crystals and glass are in contact and out of equilibrium with olivine, clinopyroxene, orthopyroxene and spinel. The origin of both the glass films and pockets, and the associated reaction textures, has been a matter of considerable debate. Glass is interpreted to have formed by partial melting (Laporte et al. 2006), infiltration of kimberlite-

like melts and other metasomatizing agents (Schiano and Clocchiatti 1994; Bonadiman et al. 2005), breakdown of amphibole (Shaw and Klügel 2002) and infiltration of silica-undersaturated host melts and reaction with xenolith minerals during excavation and transport from the mantle source (Shaw and Edgar 1997; Shaw et al. 2006).

The samples described in this study were collected from Tertiary to Quaternary alkaline basalts from Donegal, Ireland, Sal Island, Cape Verde and the West Eifel volcanic field, Germany.

PETROGRAPHY OF METEORITES AND PERIDOTITES

Overview of Petrography

The meteorites described in this study belong to the shergottite group of Martian meteorites and can be divided into two lithologies: basalts and “lherzolitic” basalts (McSween and Treiman 1995). The basalts range from fine- to coarse-grained and have a relatively simple mineralogy; clinopyroxene (pigeonite and augite) with lesser amounts of plagioclase (transformed by shock metamorphism to maskelynite), titanomagnetite, ilmenite, pyrrhotite, whitlockite and accessory apatite, quartz, baddeleyite, fayalite, and mesostasis. The pyroxenes in these basalts typically show preferred orientations. Some basalts are olivine-phyric or olivine-orthopyroxene-phyric although the origin of the megacrysts as xenocrysts or phenocrysts is debated. The basalts are interpreted as originating from lavas with entrained crystals erupted as flows on the Martian surface or as shallow intrusions, whereas the lherzolitic shergottites are interpreted as true cumulate (subsurface) rocks (McSween 1994).

Mantle xenolith samples overlap in mineralogy with that of the Martian meteorites investigated in this study, covering the entire range of peridotite types from dunite through harzburgite and lherzolite to wehrlite. They range from subangular to rounded and have diameters between one and ten centimeters. They are either completely enclosed by, or encased in, a 0.2 to 1 cm thick coating of the host lava.

Glass-Bearing Melt Pockets and Veins

In cross section, melt pockets in Martian meteorites typically have subrounded to irregular shapes, have variable measured diameters 100–5000 μm , and are typically associated with a halo surrounding the pocket that has been altered by infiltrated melt (see overview in Fig. 1). In basaltic and lherzolitic shergottites, the melt pockets show similar characteristics; both are heterogeneously distributed throughout the host rock and have variable bulk composition, they show an overlap in apparent width (micron to cm-size) and smaller pockets contain stishovite (Beck et al. 2003;

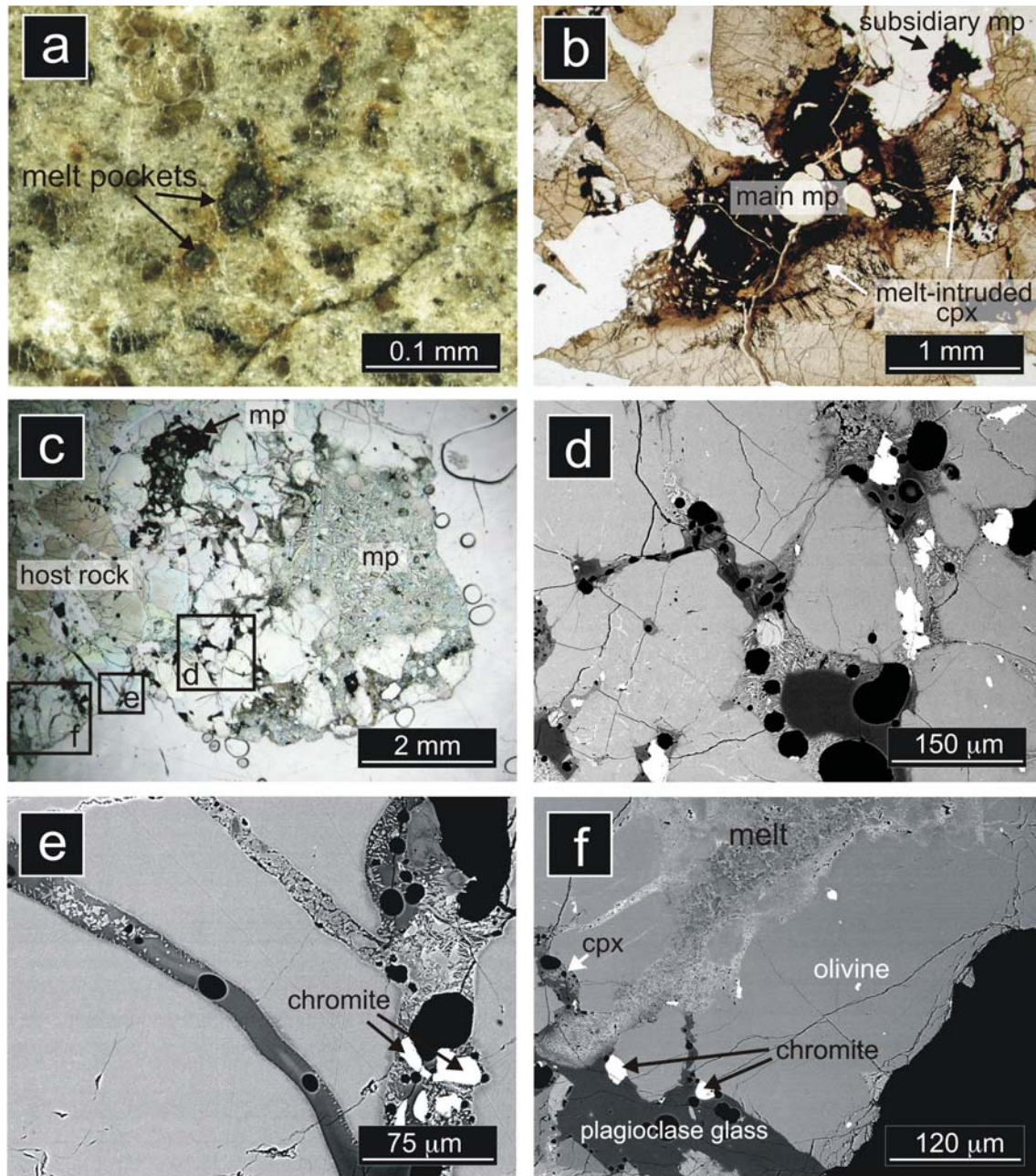


Fig. 1. Plane light (a), transmitted light (b, c) and BSE images (d–f) illustrating the often complex relationship between melt pocket (mp) and host rock. a) Small round pockets of melt show a simple relationship with the host basalt (NWA 1068). b) In LA, a melt vein, emanating from the main pocket is connected to a smaller, subsidiary pocket of melt. Clinopyroxene (cpx), in contact with the pocket, has been pervasively intruded by melt, giving it a blackened appearance. Another melt pocket (not shown) is located ~1 mm to the right of the melt pocket depicted in this image. c) Overview of a large melt pocket and associated melt network in ALH 77005. Textural relationships are shown as higher magnification BSE images in (d–f; see text for details).

Gillet et al. 2005). The main difference is the predominance of olivine-saturated melt pocket compositions in the lherzolitic shergottites, reflecting the higher modal abundance of this mineral in the igneous portion of these rocks.

The relationship between the host meteorite and melt pocket is complex and variable, even between different melt pockets in a single thin section. However, the following list

comprises the textural relationships that are most typical; (1) rounded pockets that show simple contact relationships (Fig. 1a), (2) abundant melt veins emanate from a main pocket some of which branch off into subsidiary pockets, some which are simply melt-filled cracks and fractures giving rise to a blackened appearance in transmitted light (Fig. 1b), and (3) a main melt pocket, surrounded by a

network of melt-filled veins that isolate unmelted host rock minerals, some of which connect to form additional pockets of melt (Fig. 1c). For the melt pocket shown in Fig. 1c, a series of BSE images are used to illustrate the complexity in texture and crystallization of the network surrounding the pocket. In Fig. 1d, a complex melt network surrounds unmelted olivine. The interior texture of the melt veins is shown in Fig. 1e, where the melt has partially crystallized, often nucleating from the margins inward, however, chromite grains (white) have not melted but exhibit partially rounded grain shapes. The crystallites formed in the veins range from clinopyroxene and olivine, to needles of oxides and silica. The compositions of these glass + crystals has not been reported. Figure 1f illustrates a smaller subsidiary pocket of crystallized melt ~4 mm from the main area of melting (mp in Fig. 1c). Arms and offshoots of melt emanate from the pocket into the adjacent rock.

Glass occurs in several distinct textural settings in peridotite xenoliths (Fig. 2). The most common is glass in brittle fractures that are connected to the outside of the xenoliths and the host lava. In this case the glass is texturally and chemically similar to the host lava. Where glass from the host lava is in contact with orthopyroxene a reaction zone is developed (see below) that contains variably silica- and alkali-rich glass. Where this melt has migrated along grain boundaries and has come in contact with spinel and clinopyroxene these minerals have sieve texture (see below) with abundant inclusions of glass. Glass derived from the breakdown of amphibole is silica-undersaturated, in common with the host lava, and is typically responsible for the formation of reaction zones on orthopyroxene.

SPECIFIC MINERAL/MELT REACTION TEXTURES

Low-Ca Pyroxene

In Martian meteorites, low-Ca pyroxene is restricted to the lherzolitic shergottites (ALH 77005, NWA 1950, and NWA 4468 in this study), forming large oikocrysts that enclose olivine and chromite, and as megacrysts in one basaltic meteorite investigated in this study (EET 79001). The texture observed in these grains is compared to those observed in orthopyroxene-bearing peridotites.

Low-Ca pyroxene in contact with melt exhibits two common textures. The first has been observed in two meteorites, NWA 1950 and ALH 77005 (Figs. 3a–c). At the margin, clinopyroxene dendrites comprise a zone ~30 μm wide that grades into a zone of dendritic spinel $\pm$ clinopyroxene of similar width and finally skeletal olivine. This contrasts with the granular texture that is most typical for olivine/melt contact (and to a lesser extent clinopyroxene, see following section). In EET 79001, a low-Ca megacryst has been observed with skeletal clinopyroxene nucleating from its margin into the melt pocket (Fig. 3c).

One of the most common features of mantle xenoliths is

the development of reaction zones on orthopyroxene where it is in contact with the host lava or with melts generated by breakdown of amphibole (Tracy 1980; Shaw and Edgar 1997; Shaw and Klügel 2002). These zones comprise olivine, clinopyroxene and variably silica- and alkali-rich glass (Figs. 3d–g) as rinds ranging from 10–300 μm , and as veins, around and through orthopyroxene, to complete replacement of 0.75 cm porphyroclasts. Within the reaction zones, discrete sub to euhedral olivine grains show optical continuity over the length and width of the reaction zone. Although the reaction products are distinct from those observed in the meteorites (Fig. 3b versus Figs. 3d–g), they are very similar in terms of development of sieve texture and cellular margins.

Olivine

Olivine occurs as (1) a vermicular intergrowth of fayalite + hedenbergite + silica formed from the breakdown of pyroxferroite in some Martian basalts, (2) megacrysts (xenocrysts?) in some Martian basalts, and (3) a cumulus phase in lherzolitic shergottites. Olivine is thus present in all of the thin sections investigated and displays a variety of textures when in contact with shock melt material. Due to the range in its occurrence, olivine grain size varies between samples, from ~30–3200 μm . Pyroxferroite breakdown material is not present in terrestrial peridotites and its response to shock melting has not been described here.

Compositionally zoned olivine has been noted (Figs. 4a, 4b, 4e). Although igneous olivine is zoned continuously from core to rim, interpreted to be a primary growth zonation, the zoning associated with the shock melt pockets is distinct in that it is only observed when the grain is in direct contact with melt. Many cases of scalloped or cellular rims on olivine have been observed, sometimes with cellular texture and compositional zoning developed in the same grain (Figs. 4b and e). In the larger mm- to cm-size pockets, olivine shows a gradational contact texture progressing from igneous olivine to a zone composed of anhedral to subhedral, strongly zoned olivine crystals (1–5 μm size) (Figs. 4c and 4d). The zoned crystals lack interstitial glass and has been referred to as ‘leopard’ or granular texture (Walton and Herd 2007a, 2007b) because of its spotted or mottled appearance in BSE images (Fig. 4d). Within the same pocket, when the melt has infiltrated between olivine and low-Ca pyroxene, the reaction texture developed at the olivine/melt contact is not granular, but is compositionally zoned with small dendrites nucleating from the olivine margin (see Fig. 4e).

Olivine also occurs as a common secondary mineral in Martian meteorites, formed during cooling and crystallization of shock melts (e.g., Walton and Herd 2007a, 2007b). In this occurrence, olivine has crystal shapes indicative of diffusion-limited growth from a melt (e.g., skeletal; Walton and Herd 2007a) and the compositions are/

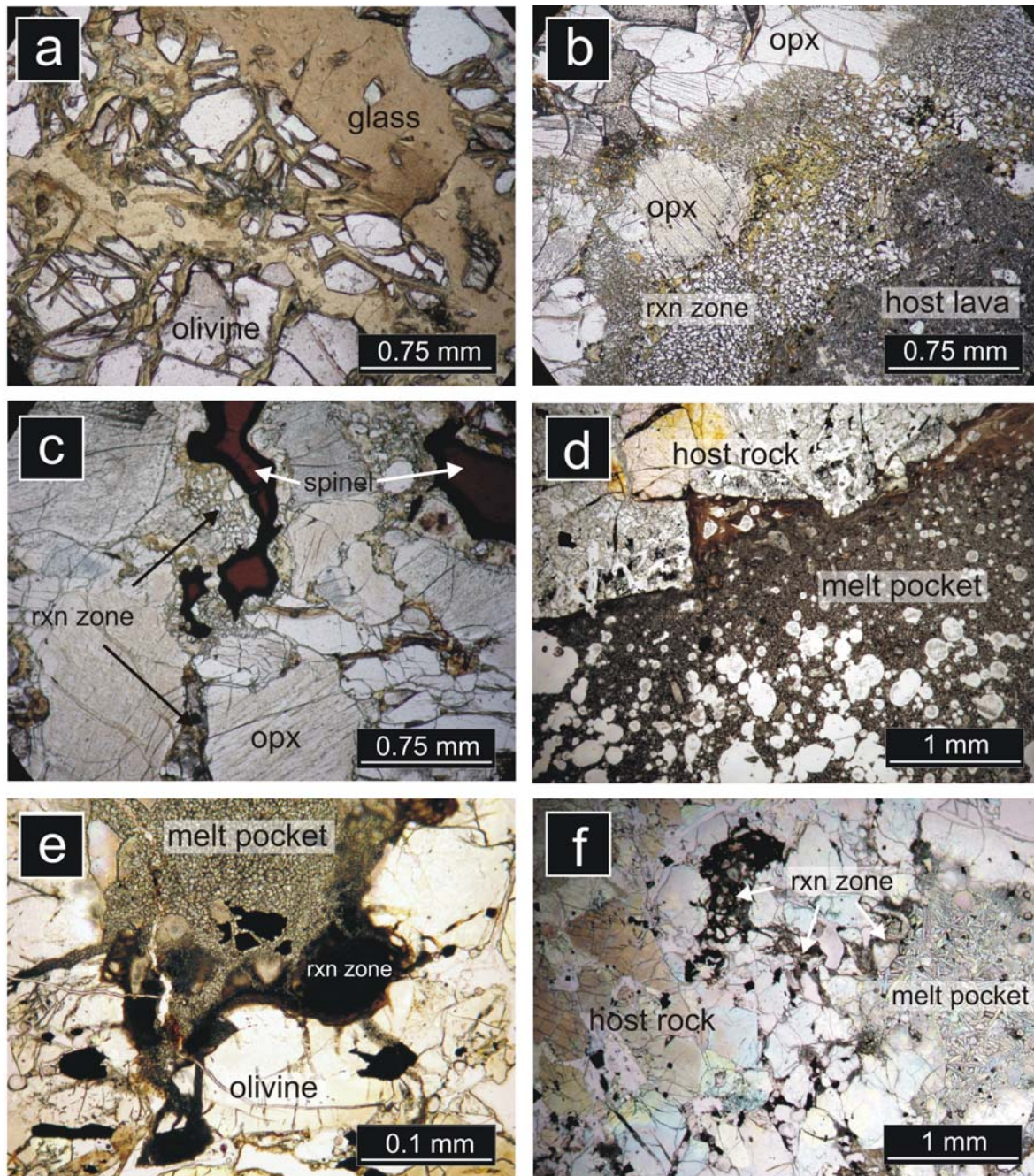


Fig. 2. Photomicrographs of melt products in (a–c) peridotites, and (d–f) meteorites. a) Glass that has brecciated an olivine rich xenolith (Donegal). b) Xenolith from the Eifel volcanic field (Germany). Reacted orthopyroxene (opx) with a glass rich reaction (rxn) zone, which also cuts across the grains in a thick vein. c) Opx and adjacent reaction zone. Spinel in the image has completely reacted (dark grain with black rim). d) Contact between the host rock (basalt) and melt pocket in EET 79001. e) Higher magnification image of a reaction zone developed at the contact between the melt pocket and olivine in ALH 77005. f) Overview of a larger melt pocket in ALH 77005 and the host rock.

more magnesian compared to igneous olivine in the host rock; however, detailed profiles across olivine in contact with the melt have not been reported.

Primary olivine in the peridotite xenoliths ranges in size from 0.5 to 4 mm and exhibits two common textures where it is in contact with infiltrated or internally generated melt. The margins of the grains are commonly scalloped

or cellular (Fig. 4f) and there is a variably thick zone of Fe-Mg enrichment/depletion together with Ca-enrichment (Figs. 4g and 4h), similar to the melt/olivine contact relationships observed in meteorites (Figs. 4a, 4b, 4c, 4e). Granular textures observed in the meteorites and shown in Fig. 4d have not been observed in peridotite olivine.

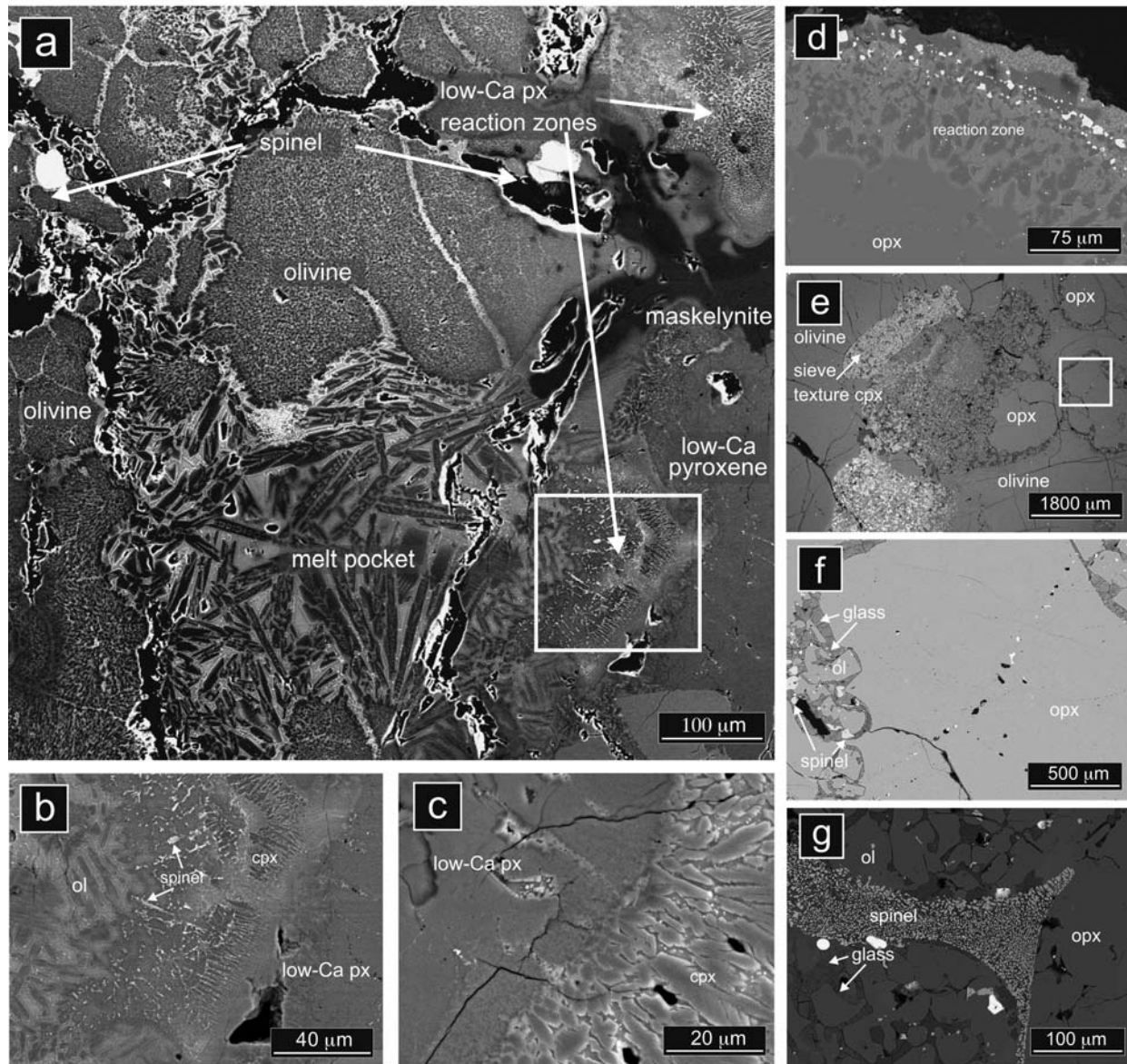


Fig. 3. SEM BSE images of low-Ca pyroxene reaction textures observed in NWA 1950 (a, b) and EET 79001 (c) in Martian meteorites, and orthopyroxene in peridotite xenoliths from Inver, Donegal, Ireland (d) and Sal Island, Cape Verde (e–g). a) Melt pocket overview illustrating the different contact textures between the two mineral phases; olivine displays a granular texture and low-Ca pyroxene exhibits a reaction zone, shown at higher magnification in (b). b) At the contact between low-Ca pyroxene and the melt, a zoned texture is observed of cpx dendrites, spinel dendrites and finally olivine (ol). c) Skeletal clinopyroxene grains nucleate from the margin of low-Ca pyroxene. d) A reaction zone on the edge of an orthopyroxene (opx) which comprises olivine with interstitial glass, light grey clinopyroxene and spinel (white phase). The Si, Na and K content of the glass vary from the inner part of the reaction zone to the outer (not shown). e) Reaction zone around a large opx crystal with a detail of the reaction zone in (f). The mineral assemblage of the reaction zone is olivine + glass + spinel. g) Orthopyroxene reaction zone containing olivine and silica- and alkali-rich glass. The reaction zone envelopes a primary spinel that has completely reacted to form sieve-texture spinel.

Clinopyroxene

Pigeonite and augite are the dominant minerals in Martian basalts, they are strongly zoned, sometimes with complex zoning profiles (e.g., EET 79001; Mikouchi et al. 1997). Exsolution lamellae within pigeonite are observed in some of the basalts. In the lherzolitic shergottites pigeonite

occurs as smaller interstitial grains and augite forms exsolution lamellae in low-Ca pyroxenes, as well as small interstitial grains.

No detailed compositional data are currently available for reaction zones observed in Martian meteorites, however, some typical textures for clinopyroxene/melt contacts are shown in Fig. 5a–e. In Los Angeles, a reaction rim

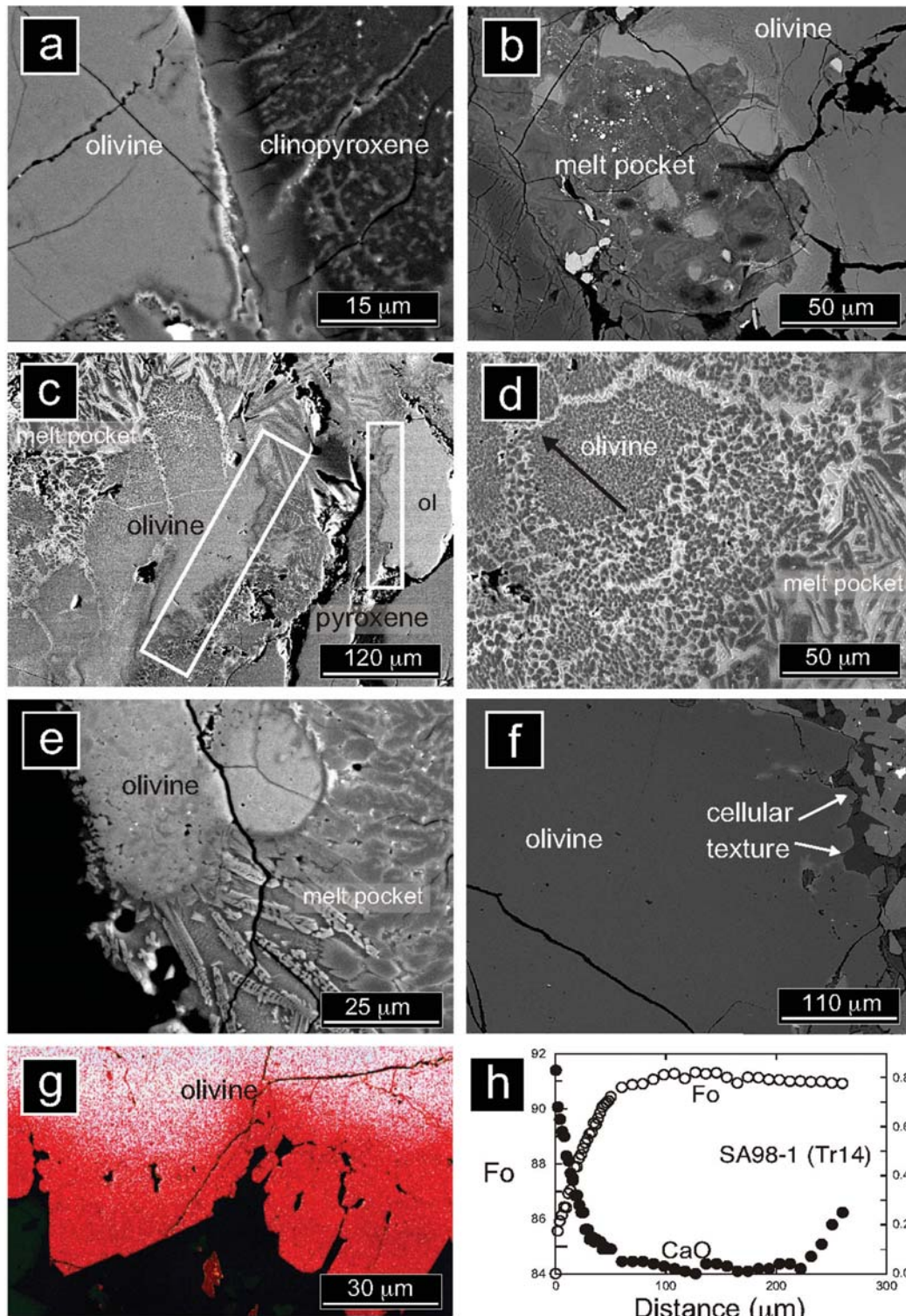


Fig. 4. BSE images of olivine-melt reaction textures from EET 79001 (a, e), NWA 1068 (b), NWA 1950 (c, d) and peridotites (f–h). a) Zoned olivine in contact with clinopyroxene near a melt pocket. b) Zoned and scalloped edges. c) In contact with melt, olivine shows a granular texture, however, when in contact infiltrated melt along the pyroxene/olivine grain boundary a cellular margin with a zoning profile is developed (area inside boxes). d) As the texture is traced toward igneous olivine (direction of arrow) the granular texture become increasingly fine-grained and the zoning less pronounced. e) Cellular texture at the olivine/melt margin. f) Cellular texture developed at the edge of an olivine xenocryst. g) Fe K α X-ray map of zoned olivine from Meerfelder. h) Compositional profiles through olivine in contact with silica-undersaturated host lava (xenolith from Sal Island, Cape Verde).

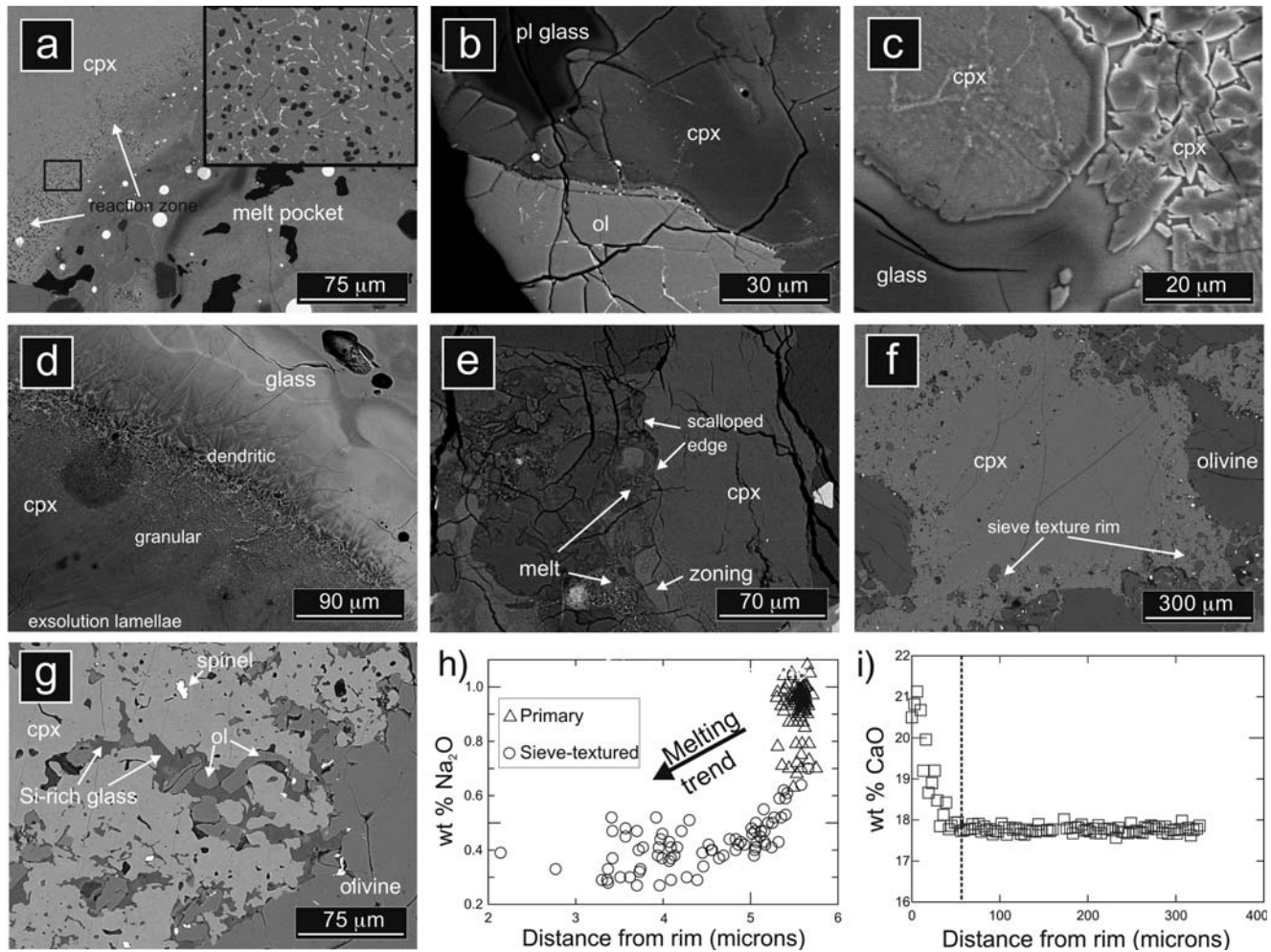


Fig. 5. SEM BSE images of clinopyroxene (cpx) reaction textures observed in Martian meteorites Los Angeles (a, d), EET 79001 (b, c), NWA 4468 (e) and natural peridotite samples from Cape Verde (f, g). a) Clinopyroxene in contact with shock melt, reacted to form clinopyroxene + FeTi oxides + silica (insert; light grey = cpx, dark = silica, white = oxides). b) Diffusion profiles in clinopyroxene and plagioclase glass. c) Cpx in direct contact with the melt exhibits a zoning profile and cusped rim superimposed on the original prismatic grain shape. A mottled texture in the grain interior is also noted. d) The texture of cpx progresses from igneous (exsolution lamellae) to granular to cpx dendrites nucleating into glass. e) A scalloped and compositionally zoned cpx rim. f) Clinopyroxene (in contact with olivine) has a clear (homogeneous) core surrounded by a sieve-textured rim. Small grey grains within the rim are olivine (ol), clinopyroxene and silica-rich glass (g). h) The composition of primary cores versus sieve rims for clinopyroxene analyzed from Shaw et al. (2006). The arrow shows the expected trend of clinopyroxene compositions during melting of peridotite (from Hirose and Kawamoto 1995). i) A compositional traverse from clinopyroxene rim to core (left to right) showing the difference in CaO between the sieve zone and the homogeneous core (from Shaw et al. 2006). The cross-hatched line marks the onset of sieve texture.

~100–200 μm wide is observed consisting of clinopyroxene + FeTi-oxides + silica $\pm$ FeS (Fig. 5a). Diffusion profiles are developed on clinopyroxene near the shock melt in contact with plagioclase glass and olivine (Fig. 5b) or in direct contact with the melt (Fig. 5c). Granular texture, identical to that described for olivine in the same meteorites—small (~5 μm) grains of strongly zoned clinopyroxene grains giving rise to a mottled or “leopard” texture—has been observed at clinopyroxene/melt contacts (Fig. 5d) with a gradation between igneous clinopyroxene with exsolution lamellae, to granular, to dendritic clinopyroxene crystals nucleating into glass (Fig. 5d). Scalloped or cellular margins with diffusion profiles have also been observed in the same grain (Fig. 5e).

Clinopyroxene also occurs as a secondary mineral that has crystallized as a liquidus phase in some melt pockets (Walton et al. 2006). These clinopyroxene grains have compositions that are distinct from that of the igneous pyroxene and have skeletal and dendritic shapes.

Clinopyroxene in the peridotite xenoliths ranges in size from 0.1 to 4 mm. It occurs in apparent textural equilibrium with olivine, as well as in veinlets and pockets that are interpreted to be the result of mantle metasomatism. Unreacted clinopyroxene is optically clear with local trains of melt and fluid inclusions.

Sieve-textured reaction rinds and cross cutting veins in sharp contact with regions of melting are common in

clinopyroxene that is in contact with infiltrated Si-rich melt in mantle xenoliths (Shaw and Edgar 1997; Shaw et al. 2006). Clinopyroxene in these sieve-textured regions has a cellular texture that is closely associated with 1–5 mm glass inclusions and secondary olivine with diffuse grain boundaries (Figs. 5f and 5g). Although cellular texture and compositional profiles similar to those in peridotites are also observed in meteorites (e.g., Figs. 5b, 5c, 5f versus Figs. 5f, 5g), the reaction products in the meteorites (silica + oxide + clinopyroxene; Fig. 5a) are distinct from those in peridotite. Granular textures have not been observed in peridotite clinopyroxene, nor is sieve texture developed at the contacts between clinopyroxene and silica-undersaturated melt veins and pockets.

In peridotites, the compositional variation from the unreacted core to the sieve-textured rim is the same regardless of locality. The sieve-textured regions are depleted in Al and Na and enriched in Fe, Cr and Ca (see Fig. 5h for Na_2O wt% and Fig. 5i for CaO wt% profiles). Detailed profiles across the transition from the core to the sieve-textured rim indicate that the compositional variations are abrupt and there is little evidence for extensive diffusion across the contact (Fig. 5i). The compositional variation between the core and rim is similar to that observed in partial melting experiments (Mysen and Boettcher 1975; Hirose and Kawamoto 1995) indicating that the rims are formed by selective removal of low melting temperature components (Fig. 4c; see further details in the discussion section).

Oxides

The oxide minerals in Martian basalts are typically ilmenite, titanomagnetite and ulvöspinel, whereas in lherzolitic basalts the primary oxides phases are chromite and ilmenite. Chromite, entrained within shock melts, exhibits a sieve texture rim in direct contact with the melt. The rim consists of vermicular spinel with abundant melt inclusions (Fig. 6a, b). Although sieve rims on spinel are a common feature of entrained grains, other grains within the same melt pocket may have completely reacted with the melt (Fig. 6c). Ulvöspinel and ilmenite in the host rock, with one side in direct contact with the melt pocket, exhibit a variety of textures including cellular, sieve and in some cases, has been partially melted to form stringers and blebs, or even completely melted to form spherules (Figs. 6d–f). Detailed compositional data from sieve-texture spinel and ilmenite in meteorites has not been reported.

The texture of spinel in the peridotites is very similar to that of clinopyroxene, in which sieve-textured rims are closely associated with infiltrated Si-rich alkaline glass (Figs. 6g and i) and are identical to reaction textures shown for Martian meteorites (Figs. 6a–d). However, in some samples, there are two distinct textures often on the same grain; sieve texture with a subhedral, isolated to

interconnected spinel rim and glass inclusions that are in sharp contact with the host spinel (Fig. 6g), and zoned grains with homogeneous inclusion free rims that have diffuse contacts with the host grain (Fig. 6h).

Compositional profiles across both types of xenolith spinel show similar patterns with depletion of Al, Cr, Mg and Ti and enrichment of Fe (see Fig. 6i for Cr_2O_3 wt% profiles for Sal xenoliths; additional compositional profiles can be found in Shaw et al. 2006). The contact between the sieve zone and host spinel is sharp where the sieve rim is separated from the host by a glass film, however, where the two are in contact (as well as in the homogeneous grains) there is a diffusion profile developed.

EXPERIMENTAL STUDIES

To date, experiments on Martian meteorite shock melts have been designed to estimate the conditions of melt crystallization such as the role of heterogeneous nucleation in crystallization kinetics and the cooling interval. These experiments focus on the internal crystal shapes of the pocket, rather than the contact relationships between the melt and host rock, and have been successful in reproducing observed crystal shapes and assemblages (Fig. 7).

Earlier melt pocket crystallization studies used homogeneous starting material having the composition of three pockets observed in several Martian basalts (Walton et al. 2006), using crystal shape and composition to determine fits between the experimental products and the natural samples. Internal textures of the natural pockets were microporphyritic with dendritic (disequilibrium) crystal shapes, set within a glassy groundmass. These particular pockets lacked abundant visible nucleation sites (e.g., schlieren, vesicles, entrained host rock clasts). The experiments were conducted assuming that the melt pockets formed by cooling of what was originally a superheated melt (i.e., liquid free of nuclei). The results showed the best fits for rapid cooling rates and the duration of the heating event was found to be on the order of minutes in a small, decimeter-size meteoroid, consistent with previous size estimates (e.g., Eugster et al. 2002; Artemieva and Ivanov 2004; see Walton et al. 2006 for details). Walton and Herd (2007b) investigated the use of such homogeneous starting materials to reproduce the texture of objects that frequently contain observable nucleation sites. In this study, a series of crystallization experiments were performed on synthetic glasses matching the composition of a melt pocket found in ALH 77005 in order to evaluate the heterogeneous nucleation potential of the melt and the effect of oxygen fugacity on crystallization. The starting temperature of the experiments varied from superliquidus to subliquidus and was cooled over a range of rates and $f\text{O}_2$ conditions. By comparing the experimental products with the natural samples, it was shown that strong thermal gradients existed within the melt at the onset of

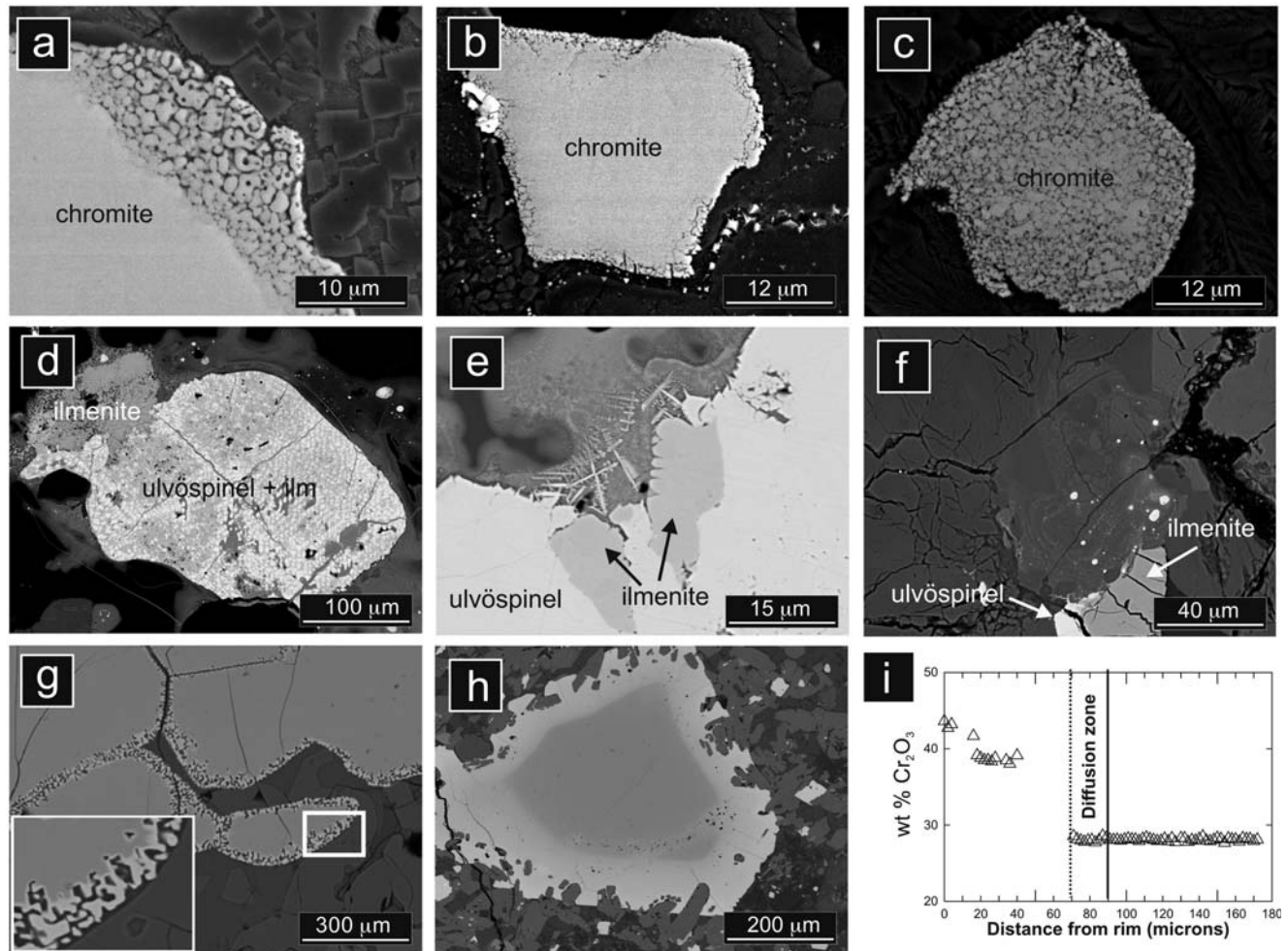


Fig. 6. SEM BSE images of spinel-melt reaction textures ALH 77005 (a), EET 79001 (b, c), Los Angeles (d, e), SaU 150 (f), and Sal, Cape Verde (g, h). a) The reaction zone on chromite has a vermicular textures and is developed only on portions of the rim. b) Reaction zone present as a rim completely surrounding the grain. c) Chromite reaction zone complete throughout the grain. d) Near the host rock/melt contact, a large clast of ulvöspinel + ilmenite has been entrained within the melt. The reaction texture is pervasive, yet the original clast perimeter has been retained. e) At the rock/melt contact, ulvöspinel displays a distinct scalloped or cellular texture. f) Ilmenite and ulvöspinel in the host rock, in direct contact with the shock melt showing partial melting (stringers) and complete melting (spheres). g) Sieve textured reaction zone around spinel in a xenolith. The spinel is in contact with Si-rich glass derived from an orthopyroxene reaction zone. At higher magnification (inset) the spinel shows cellular texture with abundant melt inclusions. h) Reaction texture of spinel in the host lava from the same sample depicted in (g). Here, the texture is distinctly different, with an inclusion-poor rim around the host grain and a diffusion zone between the two. The difference between (g) and (h) is interpreted to be due to the viscosity of the melts (see Shaw et al. [2006] for details). i) Variation in spinel composition from Sal xenoliths.

crystallization, giving rise to a heterogeneous distribution of nucleation sites resulting in gradational textures of olivine and chromite. The oxygen fugacity of the melt pocket was found to be set by the $\text{Fe}^{3+}/\text{Fe}^{2+}$ ratio imparted by melting of the host rock (FMQ -2.6), rather than external factors such as incorporation of CO_2 -rich Martian atmosphere, or melting and injection of oxidized surface (e.g., regolith) material.

Reaction textures similar to those described in the previous section for melt pockets (Figs. 4–6), have been reproduced experimentally by placing discs of natural peridotite in synthetic leucite melt (Shaw and Dingwell 2008) under conditions of 1 atmosphere for a duration of hours to days. Although the mineral assemblage, as well as

melt and mineral compositions of these experiments differ slightly from those observed in the meteorites, similar relationships between melt/host rocks are observed. The experimental data show that the final reaction texture is the result of a multistage reaction process, where passively infiltrated silica-undersaturated melt reacts with orthopyroxene, which undergoes incongruent dissolution to produce a silica and alkali-rich melt + olivine. This Si-rich alkaline melt migrates along grain boundaries, causing incongruent dissolution of clinopyroxene and spinel, followed by nucleation and growth of secondary clinopyroxene/spinel once the reacting melt is saturated in these phases. This process gives rise to the cellular textured

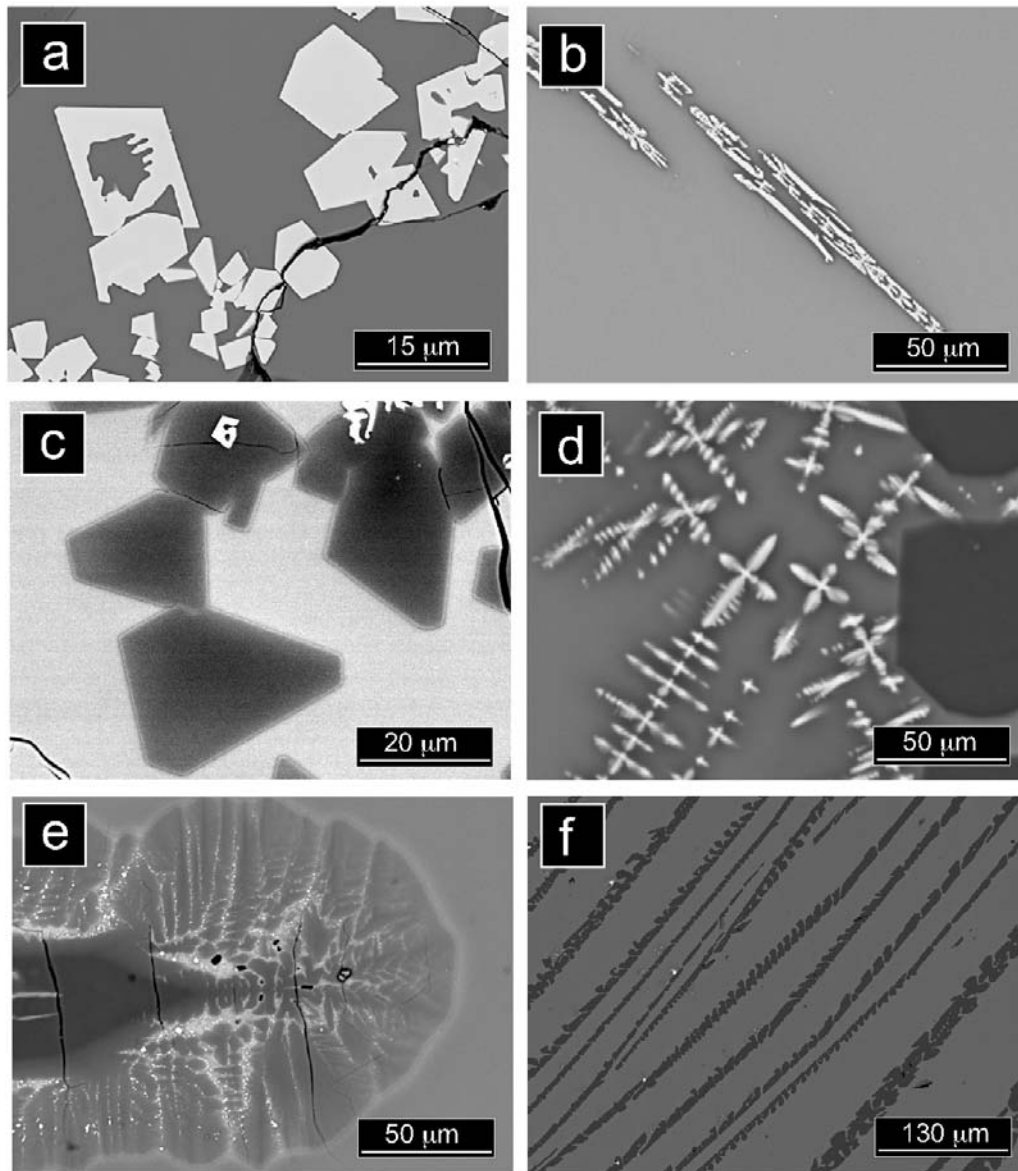


Fig. 7. SEM BSE images of crystal shapes produced experimentally that are identical to those observed Martian meteorite melt pockets. All images are from Walton et al. (2006) and Walton and Herd (2007).

rims (Shaw and Dingwell 2008). However, for spinel the occurrence of cellular texture on both types of sieve-textured rim is more consistent with growth than with dissolution (Roeder et al. 2001). This apparent paradox can be explained by examining the nature of the incongruent melting/dissolution process with reference to the analogue in situ experiments of Maiwa et al. (2003). These experiments show that during the breakdown of an incongruently melting/dissolving solid there are three steps to the process. First, the solid breaks down congruently, dissolving completely into the solvent. This is followed by nucleation of a new phase in the now oversaturated boundary layer melt. Finally, the new phase crystallizes from in the boundary layer, forming crystals whose morphology is a function of their growth rate.

These experiments shed light on the reason for the two different textures in spinel; sieve textured spinels (and clinopyroxene) are the result of a large oversaturation of the components that formed the rims. Homogeneous rims are the result of small degrees of oversaturation. This is analogous to the oversaturation caused by undercooling in the experiments of Lofgren (1980).

Olivine in the peridotite is fragmented and veined by the leucite solvent, and, in several cases, complete necking and isolation of olivine is observed. A series of images are shown in Fig. 8, illustrating the similarity in textures produced in the experiments of Shaw and Dingwell (2008) and “rock textures” observed in Martian meteorite, i.e., macrotextures between shock melts and host rock,

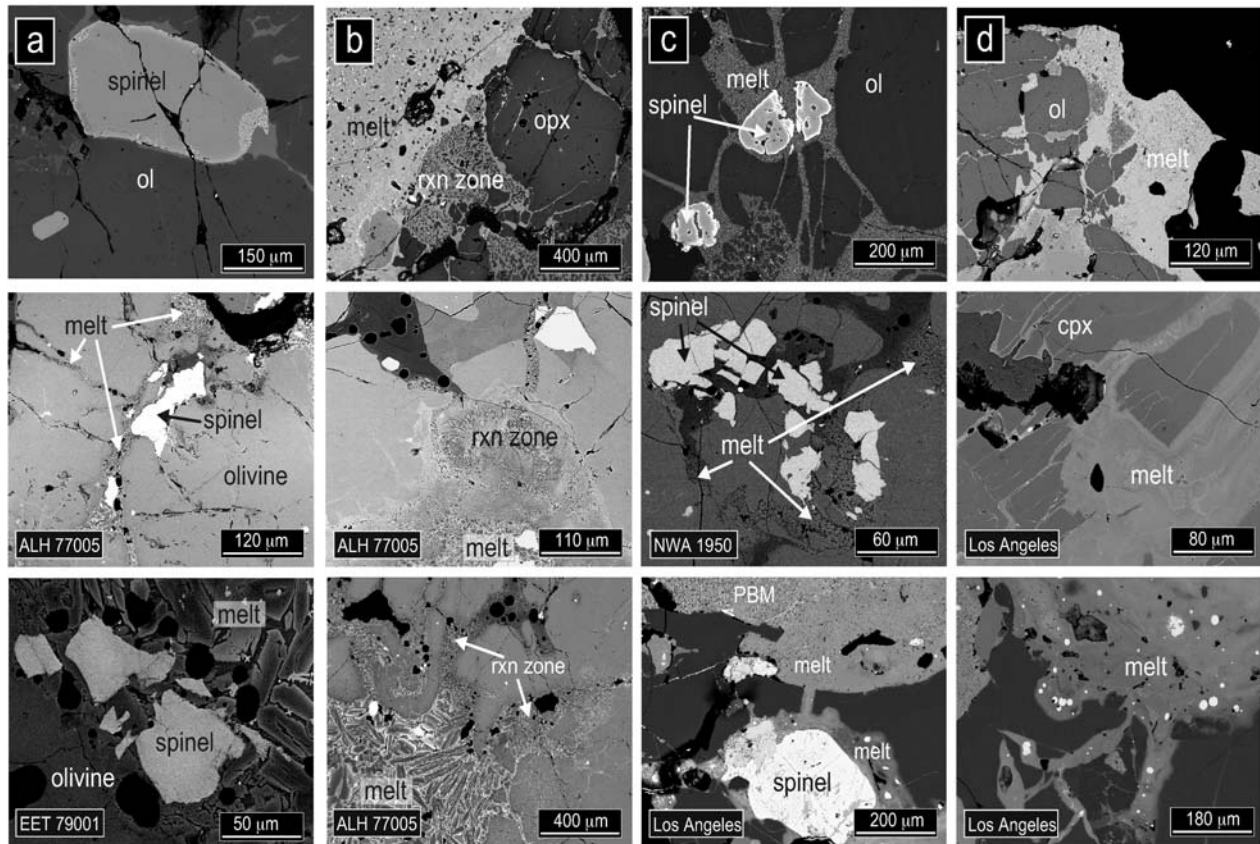


Fig. 8. a–d) SEM BSE images of peridotite–melt reaction textures produced by immersing a peridotite disc in leucite melt (Shaw and Dingwell 2007; a–d). Similar host rock/melt textures observed in Martian meteorites between host rock minerals and shock melt are shown in a series of BSE images below the experiment products. PBM = pyroxferroite breakdown material, rxn: reaction zone. The meteorites from which the textures have been observed are given in the bottom left hand corner of the individual BSE images.

compared to the microtextures examined for individual minerals in previous sections. In an experimental study of the reaction of peridotite and silica-undersaturated melt, Shaw and Dingwell (2008) showed that similar zonation developed, even over the short duration of their experiments. The shape of the compositional profiles can be interpreted in terms of Fe–Mg interdiffusion (Klügel 1998; Shaw and Klügel 2002; Shaw 2004), while the cellular texture is interpreted to be the result of dissolution/precipitation reactions (Shaw and Dingwell 2008). Experimental studies have also been used to define this interdiffusion coefficient as a function of temperature and oxygen fugacity (Chakraborty 1997). Application of these experimentally derived diffusion coefficients allows extraction of melt–olivine contact times for the compositional profiles in xenolith olivine. However, diffusion in olivine is anisotropic and unless the orientation of a compositional traverse relative to the principle crystallographic axes is known, contact durations can have errors of up to ± 6 times the calculated value. Orientations, however, can be determined using electron backscatter diffraction (Shaw et al. 2006).

For the Eifel samples, Shaw and Klügel (2002) and Shaw

(2004) determined xenolith residence times from these profiles after using mineral–melt thermobarometers (Putirka et al. 1996, 2003; Putirka 1997, 1999) and the MELTS thermodynamic model (Ghiorso and Sack 1995) to constrain the temperature of reaction. Profiles such as the one in Fig. 4g can be modeled with least squares fitting routines to extract contact times. For the profile shown in Fig. 4g a contact time of 220 days is calculated, assuming that the profile was measured along the fast diffusion direction of the crystal.

COMPOSITION

Glass in peridotites covers a wide range of compositions (c.f. Shaw et al. 1998; Fig. 9). Much of this range has been interpreted to be due to interaction between Si-poor magma and xenocrystic xenoliths during their transport to the Earth's surface, such as sieve-texture clinopyroxene and spinel or the breakdown of amphibole (Shaw et al. 2006; Shaw and Dingwell 2008), as presented in this study. Alternative interpretations include anhydrous partial melting in the mantle (Franz and Wirth 2000) or fluid-induced partial melting in the mantle (Carpenter et al. 2002) and mantle metasomatism (Bonadiman et al. 2005). However, these

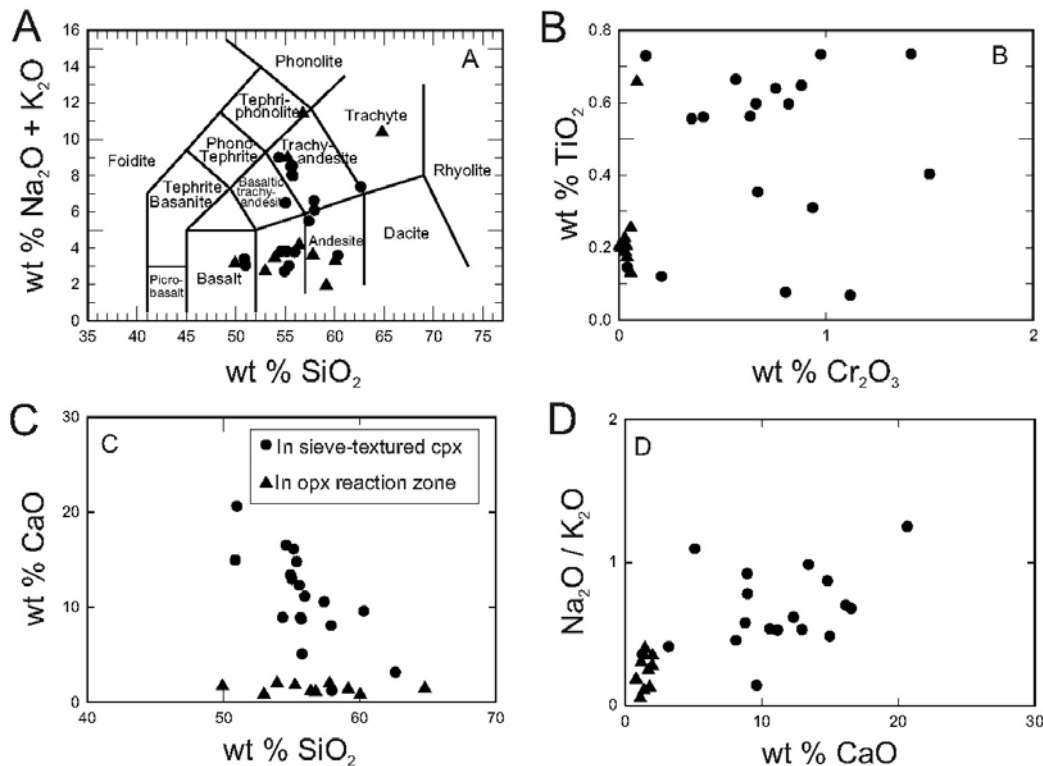


Fig. 9. Composition of glass in mantle xenoliths. a) Large range of compositions with glass developed in orthopyroxene reaction zones showing extreme silica and alkali enrichment. b–d) Differences in composition of glass associated with orthopyroxene and clinopyroxene. Note the enrichment of TiO_2 , Cr_2O_3 , CaO , and Na_2O in the glasses associated with sieve textured clinopyroxene indicating that these oxides were derived from breakdown of the pyroxene.

hypotheses have yet to be tested experimentally. In addition, melt can be formed by decompression melting of hydrous phases in the xenoliths, in particular amphibole (Shaw and Klügel 2002; Kaeser et al. 2007).

The bulk composition (glass + crystals or glass) of shock melts show a wide range in composition that is variable between pockets in a single meteorite (on the thin section scale) and between pockets found in different meteorite samples. This range has been attributed to melting of variable portions of host rock minerals by shock, where the resultant liquid composition will be a function of the composition and amount of minerals melted, and any trapped Martian alteration or soil products entrained within this liquid. This can give rise, at one end of the spectrum, to silica- and alkali-rich compositions formed by melting of late stage igneous mesostasis, to olivine-saturated melts formed by melting of cumulate minerals and compositions that reflect preferential melting of plagioclase (McSween and Jarosewich 1986; Beck et al. 2004). One melt pocket in Los Angeles was noted by Walton et al. (2006) in which whitlockite was the liquidus phase. This extreme range in bulk composition of the melts will most certainly be key, along with textures, to resolving the petrogenesis of the melts.

A very simple experimental setup can be used to demonstrate how extremely complex compositions and textures can be generated by quenching a reaction before

equilibrium is reached (Shaw 2000). Here, a quartz sphere is submerged within a basanite melt. An X-ray elemental map shows the variable melt composition and texture (Fig. 10a). This experiment conceptualizes the compositions and textures observed in meteorite melt pockets—for example, in Los Angeles a single melt pocket is observed with two immiscible silicate melts (Figs. 10 b–d). Although the melts were formed during shock, diffusion rates were too slow to homogenize the pocket before it cooled below the glass transition so there is variable melt composition even within a single melt pocket.

USE OF TEXTURES—DISCUSSION

Experiments that have reproduced textures and glass compositions observed in peridotites and have proved invaluable as tests of the ideas generated from petrologic studies for their origin (e.g., metasomatism versus residence and transport in magma). We regard the similarity in texture between meteorite/melt and peridotite/melt interaction, to be key to our interpretation of shock melts in Martian meteorites. The following textures in the meteorites are distinguished from growth textures in the melts, and are interpreted to form by reaction between the shock melt and igneous minerals: (1) diffusion profiles in olivine (Figs. 4a, 4b, 4e) and clinopyroxene (Figs. 5b, 5c), (2) cellular and sieve texture on

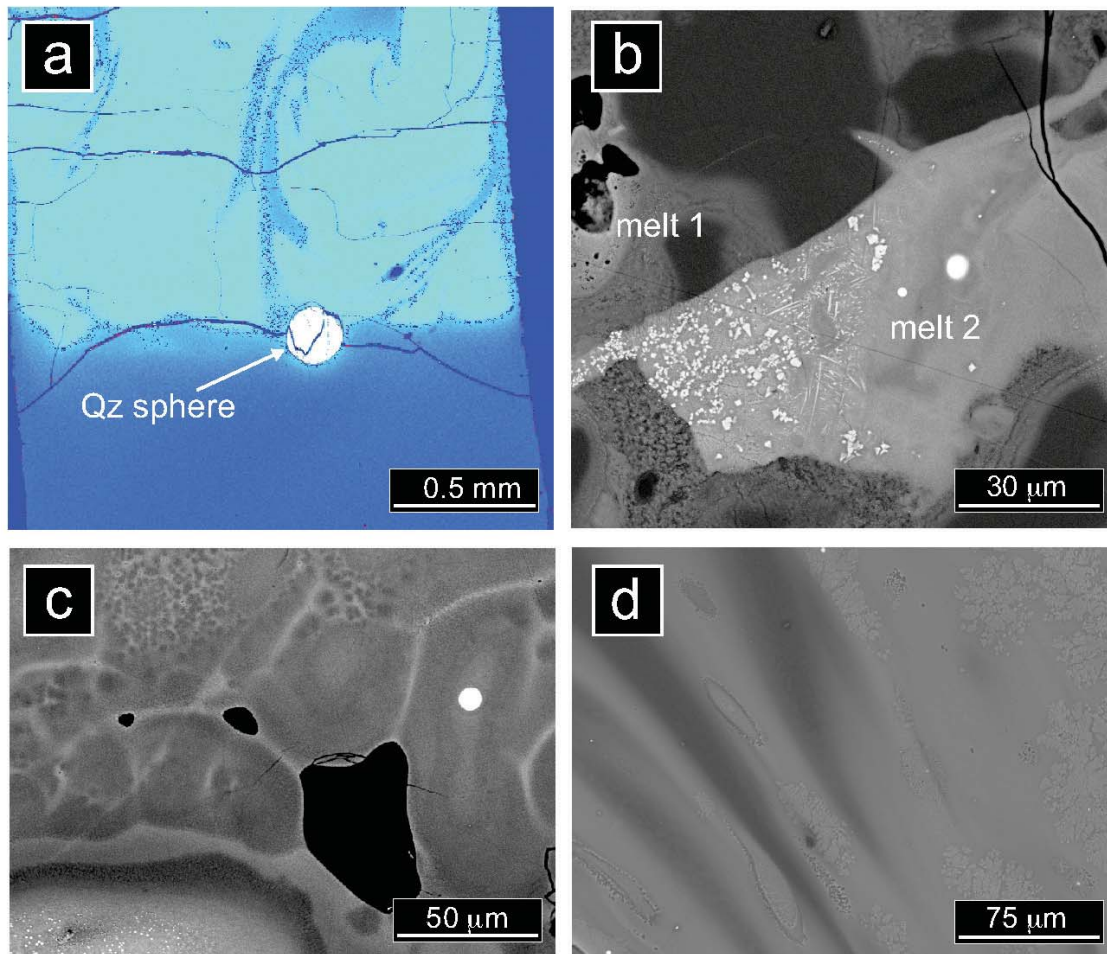


Fig 10. a) Dissolution experiment in which a quartz sphere is immersed within a basanite melt. The system is quenched before it can homogenize, demonstrating the range in melt compositions that can be generated from compositionally simple starting material. b) Two immiscible silicate melt compositions observed within a single melt pocket in the Los Angeles meteorite. c, d) Heterogeneous glass in Los Angeles melt pockets.

margins of low-Ca pyroxene (Fig. 3c), olivine (Figs. 4a, 4b), clinopyroxene (Figs. 5c, 5e), and oxides (Figs. 6a, 6b, 6e), and (3) migration of melt along cracks and fractures causing veining and necking of host rock minerals into the melt (i.e., passive brecciation; Fig. 8). Based on these textural similarities between meteorite melt pockets and glass-bearing peridotites, a number of inferences can be drawn which shed light on texture development in shock melts: a melt formed by shock is sited within pockets in the host rock. As the melt cools and crystallizes it reacts with host rock minerals that are adjacent to, and in disequilibrium with, the shock melt. As the melt and minerals react, calving or necking of host rock fragments occurs. These host rock fragments are entrained within the melt and develop sieve and cellular textures at their margins, and, in some cases, completely react with, but do not completely dissolve into, the melt. The melt also migrates along grain boundaries, preexisting cracks and fractures in the host rock. This melt is out of equilibrium with the invaded minerals and reacts with and partially dissolves them either congruently or incongruently. The mineral assemblages that

form in these veins are distinct from that observed in the main melt pocket (from which the veins emanate) because the bulk composition of the vein evolves as it reacts with the host rock minerals. Cooling times are rapid and the glass transition temperature is reached before equilibrium can be established, thereby freezing in the reaction textures. Although the texture comparison between peridotites and meteorites is at this point only qualitative, we find the textural similarities compelling, specifically in development of sieve and cellular texture on grain margins, and the migration of melt from the main melt pocket along cracks and fractures in the host rock. Thus, we regard melt reaction and migration as likely giving rise to some of the more complex melt pocket textures (Figs. 2 and 8) and variation in glass composition.

Injection versus in Situ Melting for the Origin of Shock Melts

In the previous section we have outlined various processes that we believe are key to deciphering the final

texture of melt pockets in Martian meteorites. Although this does not directly shed light on the mechanism of melt pocket formation, we are able to evaluate which mechanism would most likely give rise to the observed textures by asking the question: is the formation mechanism able to account for the petrography of the samples?

First, we consider the injection scenario, as originally proposed by Bogard et al. (1986, 1989) and Wiens and Pepin (1988), and later advocated by Rao et al. (1999). The work of Rao et al. (1999) invoke mixed rock and soil components to explain the observed disparity between shock melt composition and that of the bulk rock (EET 79001-lithology A); specifically, enrichment of S, Al, Ca, and Na, and depletion of Fe and Mg were noted for the melts (veins and pockets). The high abundance of sulfur in the melts indicates the addition of sulfur-rich material, attributed the S-enriched nature of Martian soil (Rao et al. 1999) as shown by in situ measurements by Pathfinder and Viking data (Rieder et al. 1998; Clark et al. 1982). Addition of a soil component to the shock melts is supported by the occurrence of grains of oxidized (S^{6+} -bearing) alteration products in EET 79001 (Gooding et al. 1988), inferred to be pre-terrestrial based largely on textural arguments.

Most, but not all, Martian meteorites are part of thick lava flows or shallow igneous intrusions (Greshake et al. 2004; see also reviews in McSween 1994 and McSween and Treiman 1995). One of the most well studied is Zagami, a Martian basalt thought to have crystallized in a lava flow >10 m thick (McCoy et al. 1992). The injection scenario envisions molten target rock mixed with soil, that is injected into the target rock along thin melt veins (e.g., Rao et al. 1999). The soil component within the rock constrains this melt to originate from near the Martian surface, but must travel along minute conduits in coherent target rock before it is sited within pockets in the rock. For Zagami, and most Martian meteorites, this distance would be >10 meters. The problem with this scenario is that it is hard to envision melt traveling along such minute conduits for distances over a few cm, since there would be a large temperature difference between the melt and host rock that would lead to rapid quenching. Experiments have shown that some melt pockets crystallized from temperatures of at least 1510 °C. For transport to occur over meters, the melt would have had to remain above its glass transition temperature at a low enough viscosity for flow. At thicknesses of a few hundred to 10s of microns (some veins are even thinner in apparent width; Fig. 15a) it is very hard to envisage a mechanism that would allow melt flow but not rapid quenching. For veins that are 10–300 μm wide to serve as conduits to fill a cm-size pocket there must have been sustained flow of material from the vein into the rock, yet textures show that the vein cooled very quickly (schlieren rich). Although one might remark that melt injection could occur at a few km/s velocities, taking ~10 ms for the melt to travel 10 m and that over such a time scale, quenching would

not occur; there are strong textural arguments that suggest the vein material was not transported distances greater than a few hundred microns. Namely, that clasts within the shock veins can be traced to nearby, typically adjacent, host rock minerals (Figs. 11a, 11c, 11d), suggesting that the veins did not experience significant transport but were formed in slip events with offsets on the order of a few 10s of microns (as measured by mineral offsets; Fig. 11b). In addition, in direct contact with the larger (mm-size) melt pockets, grains in the host rock show evidence of reaction for distances of a few hundred microns from the contact with the melt; therefore, the melt must be injected into the pocket and retain temperatures high enough to generate large temperature gradients (Figs. 11e, 11f). A further complication lies in the range of bulk composition of melt pockets in the meteorites. The injection mechanism cannot account for variations in composition on a mm-scale within the meteorites. The target rock + soil material would be melted, mixed and then injected. It is inconceivable to have an alkali-, silica-rich melt pocket and a P_2O_5 -rich melt pocket, as observed in Los Angeles (Walton and Spray 2003), within a few centimeters of each other in the host rock—the melt pockets would presumably all have the composition of the initial target rock + soil (i.e., basaltic). This argument is strengthened by the observation that not all melt pockets contain detectable sulfur (Gnos et al. 2002; Walton and Spray 2003). Assuming melt pockets form by the same mechanism, if they arise by injection, then they should all contain a vestigial record of Martian soil (measured by excess S; Rao et al. 1999).

Secondly, we consider the void collapse/variable shock impedance model. In this model, open spaces in the target rock (cracks, fractures, vesicles, etc.) collapse upon passage of the shock wave during impact. This collapse causes a second shock wave to superpose on the primary wave, which acts to amplify local pressure and temperature (Heider and Kenkmann 2003). The pressure spike is transient, however the temperature profile (up to 2500 K realized by the melt; Beck et al. 2007) will be controlled by thermal diffusivity of the host rock, the latter of which would have been heated above its ambient (pre-shock) temperature (ΔT 70–800 K; Fritz et al. 2005). This mechanism is able to account for the temperatures of crystallization, determined from experiments, and the large thermal gradients from the melt pocket into the host rock (Fig. 11f). The collapsed structure and temperature spike would give rise to a melt that could vary in composition on a cm- or mm- scale in the host rock (depending on the original grain size) that would reflect the composition and portions of minerals melted. The heterogeneous distribution of shock melts in the host rock is easily explained by this mechanism, because a melt will form where the void and shock wave interact. The composition of the melt will then reflect the modal mineralogy in the vicinity of the open space. It is easy to conceptualize then how the range in bulk composition relates to the grain size of the meteorite and the size of the

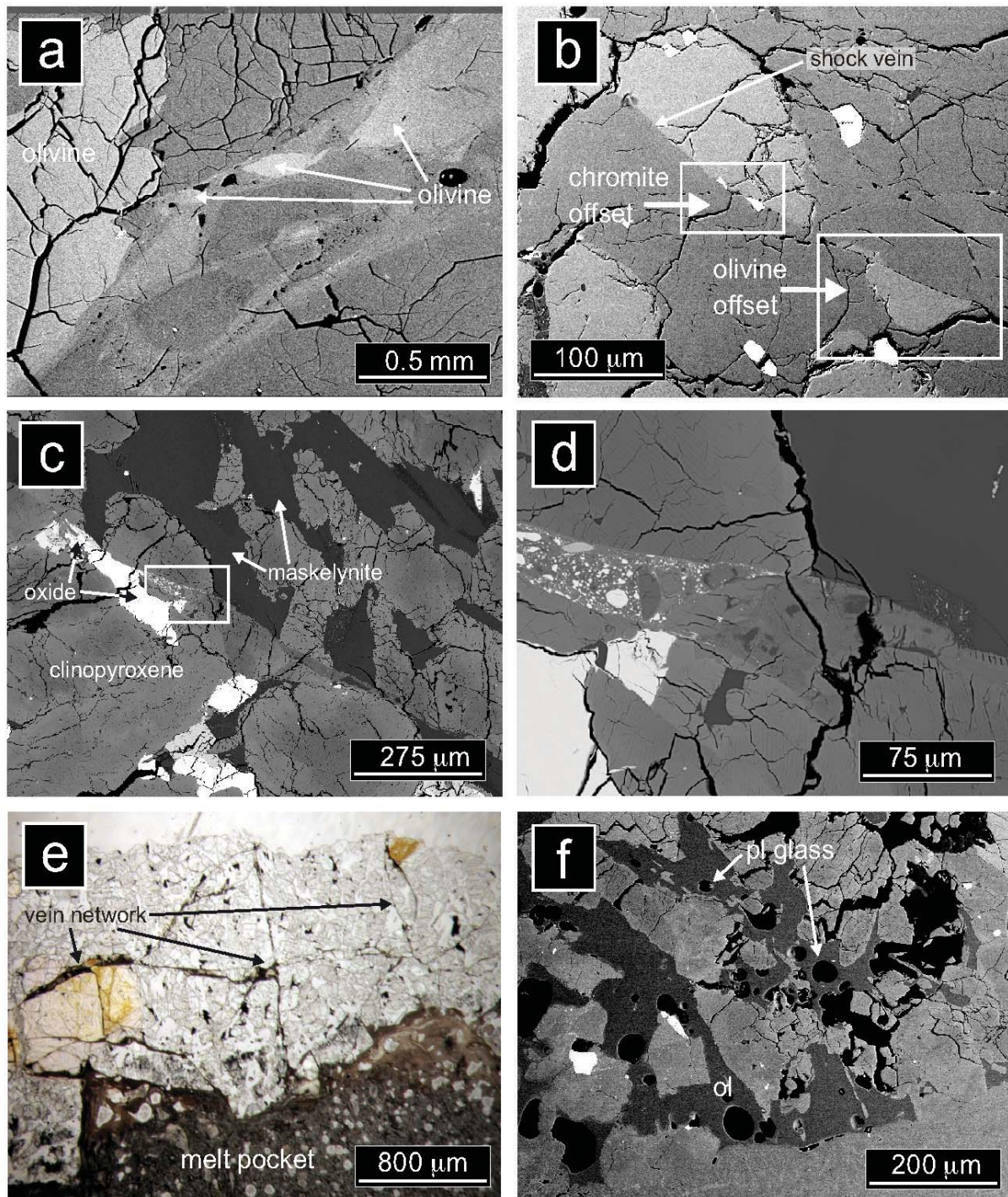


Fig. 11. BSE images of shock veins in EET 79001 (a, e, f), ALH 770056 (b) and Zagami (c, d). a) Clasts of olivine are only observed in the shock vein where the vein cuts across igneous olivine, suggesting they are locally derived and that the vein has not transported material for large distances. b) Shearing of chromite and olivine along a shock vein yield a displacement of ~ 30 mm, as measured in the two-dimensions of the thin section. c) The shock vein is laden with clasts and spheres of ilmenite, titanomagnetite and FeS, where it truncates a similar assemblage in the host rock. These phases are absent from the vein within a few hundred microns of the host rock assemblage, again, as in (a), suggesting that the material in the veins was not transported large distances. e) Transmitted light photomicrograph showing the relationship between the veins and melt pocket. The veins are, at maximum, a few hundred microns thickness, and more typically a few tens of microns thickness down to a melt-coated microfault. For the injection model, these veins must serve as conduits for the melt material in the pocket (bottom). f) Thermal gradients observed in the host rock minerals in direct contact with the melt pocket. At distances a few hundred microns from the contact, plagioclase is present as a normal glass and not maskelynite.

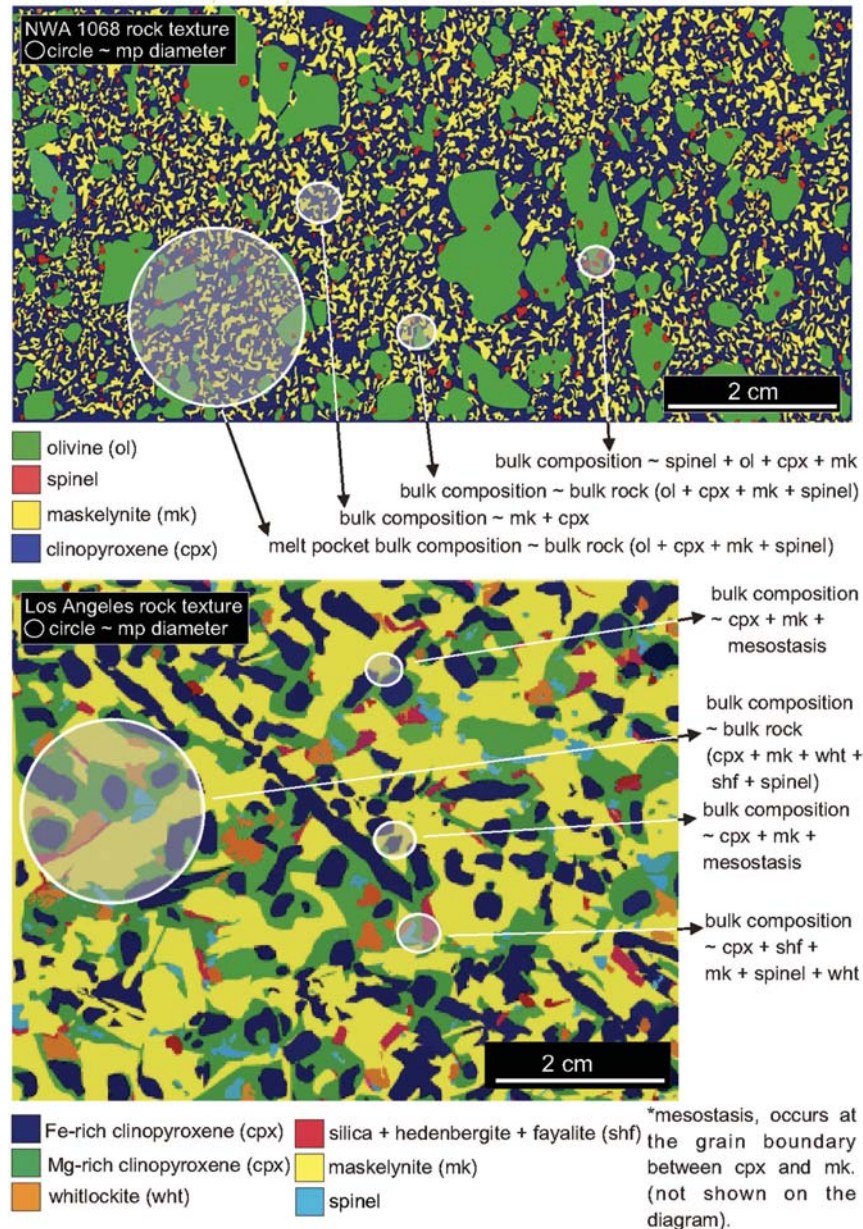


Fig. 12. Maps of NWA 1068 (olivine-phyric basalt) and Los Angeles (coarse-grained basalt) constructed from X-ray elemental maps. The general, simplified relationship between melt pocket diameter, grain size and bulk composition is demonstrated. Igneous phases were mapped by Si, Fe, Mg, Ca, Al, Cr, and P distribution. The location of shock melt pockets, observed using optical microscopy and SEM, have been removed from the map and the original extent of the shock melt is shown by the white circular outlines with a transparent interior. Grain boundaries from the X-ray maps have been projected into the melted area within the transparent overlay using an illustration program. Bulk compositions indicated are those of the melt pocket (see text for additional details).

pocket (Fig. 12). In a fine-grained rock (e.g., NWA 1068; Fig. 12a), melt pockets will approach that of the bulk composition as their size increases because a greater number of minerals are melted. In contrast, for a coarse-grained rock such as Los Angeles (Fig. 12b) even large melt pockets will only incorporate a few minerals and the bulk composition of the pockets will be highly variable. In the void collapse model, presumably this texture at the contact between the host rock and shock melt would have brecciated margins or some

features that reflect the violent formation process, however, once the melt is formed it can interact with and partially digest neighboring igneous minerals. The contact between host rock and the melt will reflect this interaction, which will have erased the original, brecciated texture. With continued cooling, the process of reaction, diffusion, and veining and necking of host rock minerals into the melt become frozen in. In this study, we find that the void collapse scenario is consistent with the petrography of the samples.

At present, our favored model of void collapse faces a number of criticisms based on observations of shock melts, including: 1) occurrence of pre-terrestrial sulfate-bearing secondary minerals in the melt products, 2) Sr isotopic anomalies compared to the bulk rock, 3) high emplacement efficiencies of noble gases, N and CO₂, and 4) S, Al, Ca, and Na enrichments, and Fe and Mg depletion compared to bulk rock composition. We contend that not only is the void collapse model consistent with long known characteristic features of shocked rocks containing void space, but it is also able to explain all “anomalous” compositional and isotopic observations as listed previously. Shock recovery experiments, calibrated with minerals (plagioclase and olivine) and rocks (basalt, gabbro, dunite pyroxenite and chondrites), have unequivocally demonstrated that “disequilibrium” shock effects (local mineral melts) require local peak shock pressures of 70–80 GPa (e.g., Stöffler et al. 1986; Schmitt 2000); at these conditions, the emplacement efficiency of gases would be 100% (Bogard et al. 1986, 1989; Wiens and Pepin 1988). This is consistent with the Martian atmospheric gases being specifically sited within shock melts, and absent in the host rock (Bogard and Garrison 1999; Walton et al. 2007). Secondary alteration minerals of Martian origin could be present in the void space prior to shock as grains trapped in cracks, fractures, microfaults or vesicles. During impact the void spaces become preferential sites for melting and the products of this melting carry a signature that is dominated by that of the host rock, with a small signature of Martian alteration or soil component, the magnitude of which is dependent on the original size and abundance of these grains prior to shock. In addition, the compositional variations (plagioclase enrichment/mafic mineral depletion) observed in EET 79001 shock glasses (Rao et al. 1999) have been noted from shock glasses from the Lonar Impact Crater (Schaal et al. 1979), as well as from ordinary chondrites (Dodd and Jarosewich 1982). Plagioclase enrichment in shock melts can be attributed to the low impedance of this mineral by shock compression (Stöffler 1984). The highly compressible nature of sulfides has also been noted for shock melts in ordinary chondrites, resulting in preferential mobilization of troilite during shock (Stöffler et al. 1991; Bischoff and Stöffler 1992). Thus, the abundances of S, Al, Ca, and Na in shock melts can be attributed to the local igneous mineralogy and the mineral-specific nature of shock melting, rather than the nature of the Martian soil.

Insights into Shock Melt Formation by Comparison with Xenolith Studies

Although limited quantitative conclusions can be drawn from mantle xenolith research and applied to the shock melts described in this study, similarities in textures between these disparate features are compelling. We contend that the methods used to decipher the origin of reaction textures in

mantle xenoliths could also be implemented in Martian meteorite research to understand complex textures and variations in glass composition, and in developing a complete model for melt pocket formation. For example, disequilibrium conditions between shock melts and meteorite minerals result in reaction textures. From these textures, certain parameters such as contact times may be resolved by examining Fe-Mg exchange between minerals and melts. However, in peridotite systems, the melts are in contact with the minerals for days to years. In meteorites, it has been determined from experiments and physical models that the melts quench on very short time scales, seconds for melts preserving high-pressure high-temperature minerals, and minutes to hours for larger pockets that remain molten after shock pressure release to form a low-pressure high-temperature mineral assemblage. For contact times of this short duration, diffusion profiles would be only a few microns wide, which would be difficult to resolve by microprobe and may require EDX analysis with a TEM. Analyses of this type have been used by Shaw and Klügel (2002) and Shaw (2004) to determine xenolith residence times, and a similar approach applied to Martian meteorites could be used to verify cooling duration results from crystallization experiments (e.g., Walton et al. 2006). The results would have implications for the noble gases contained in the melts, since cooling durations will affect diffusion of these gases. In addition, detailed glass analyses coupled with reaction experiments, as in Shaw and Dingwell (2008), could be used to constrain reaction processes between shock melts and earlier formed minerals, to determine the origin of glasses in the samples (e.g., incongruent versus congruent dissolution). In Shaw and Dingwell (2008) the experiments were also key to deciphering chemical data; without knowledge of the link between incongruent breakdown of orthopyroxene and the formation of sieve texture, one is hard pressed to correctly interpret the textures.

The granular texture, commonly observed in meteorite pyroxene and olivine, has not been observed in mantle xenoliths. We interpret this as a growth texture (see textural depiction in Figs. 7a, 8c, 8d, 8f and 9c, 9e), possibly an extension of the results of Walton and Herd (2007b), who showed that a heterogeneous distribution of nucleation sites in the melt gave rise to a gradation in crystal shapes of olivine and chromite within the pocket (i.e., elongate skeletal shapes in the melt pocket center grading to hopper and equant euhedral shape near the margin). The granular or leopard texture could result when the nuclei are so densely packed that growth of crystals occurs with no interstitial glass. We do note that the absence of similar textures in peridotitic olivine could suggest quenching from highly deformed or amorphous solids in the meteorites, or by back reaction from a high-pressure precursor; experiments and detailed compositional data will be key to unraveling the origin of this texture.

CONCLUSIONS

The availability of experimental studies and the fact that processes like shock metamorphism and xenolith transport are transient makes them very amenable to experimental investigation. Features of shock melt pockets in Martian meteorites and glass pockets in terrestrial peridotites are similar because the minerals/melts are in such extreme disequilibrium. As we have shown in this study, the textures found in some meteorites are very similar to those in peridotites, and the experimental data generated from study of xenolith/melt reactions can be applied to the meteorite systems shown (shock melts). Although experiments have been crucial in constraining the crystal growth kinetics and redox states involved in melt pocket formation and evolution, they have only considered those pockets in which the melt pocket texture is dominantly that of growth. However, the similarity between reaction/infiltration textures in terrestrial peridotites and some textures observed in melt pockets indicates that the melt generated from shock would have been able to interact with host rock minerals for a brief time. Solvent textures are most likely able to explain some of the more complex features observed at a finer scale. Unraveling these complex textures can lead to a more detailed understanding of the processes and conditions that give rise to melt pocket textures. Based on texture alone it is concluded that void collapse is the most viable formation mechanism for development of localized pockets of melt products in Martian meteorites.

Acknowledgments—This work has been funded by the Canadian Space Agency through the space science fellowships program, awarded to E. L. W. and an NSERC Discovery Grant awarded to C. S. J. S. Thanks to Gary Lofgren and an anonymous reviewer, whose comments, along with those of the associate editor (Ed Scott), improved an earlier draft of this work.

Editorial Handling—Dr. Edward Scott

REFERENCES

- Artemieva N. and Ivanov B. 2004. Launch of Martian meteorites in oblique impacts. *Icarus* 171:84–101.
- Beck P., Ferroir T., and Gillet P. 2007. Shock-induced compaction, melting, and entrapment of atmospheric gases in Martian meteorites. *Geophysical Research Letters* 34:L01203.
- Beck P., Gillet P., Gautron L., Daniel I., and El Goresy A. 2004. A new natural high-pressure (Na,Ca)-hexaluminosilicate $[(\text{Ca}_x\text{Na}_{1-x})\text{Al}_{3+x}\text{Si}_{3-x}\text{O}_{11}]$ in shocked Martian meteorites. *Earth and Planetary Science Letters* 219:1–12.
- Bischoff A. and Stöfner D. 1992. Shock metamorphism as a fundamental process in the evolution of planetary bodies: Information from meteorites. *European Journal of Mineralogy* 4:707–755.
- Bogard D. D. and Garrison D. H. 1999. Argon-39–argon-40 “ages” and trapped argon in Martian shergottites; Chassigny and Allan Hills 84001. *Meteoritics & Planetary Science* 34:451–473.
- Bogard D. D., Hörz F., and Johnson P. H. 1986. Shock-implanted noble gases: An experimental study with implications for the origin of Martian gases in shergottite meteorites. Proceedings, 17th Lunar and Planetary Science Conference. *Journal of Geophysical Research* 91:E99–E114.
- Bogard D. D., Hörz F., and Johnson P. H. 1989. Shock-implanted noble gases II: Additional experimental studies and recognition in naturally shocked terrestrial materials. *Meteoritics* 24:113–123.
- Bonadiman C., Beccaluva L., Coltorti M., and Siena F. 2005. Kimberlite-like metasomatism and ‘garnet signature’ in spinel-peridotite xenoliths from Sal, Cape Verde Archipelago: Relics of a subcontinental mantle domain within the Atlantic Ocean lithosphere? *Journal of Petrology* 46:2465–2493.
- Borg L. E., Edmunson J., and Asmerom Y. 2005. Constraints on the U-Pb isotopic systematics of Mars inferred from a combined U-Pb, Rb-Sr and Sm-Nd isotopic study of the Martian meteorite Zagami. *Geochimica et Cosmochimica Acta* 69:5819–5830.
- Bouvier A., Blichert-Toft J., Vervoort J. D., Gillet P., and Albarède F. 2008. The case for old basaltic shergottites. *Earth and Planetary Science Letters* 266:105–124.
- Carpenter R. L., Edgar A. D., and Thibault Y. 2002. Origin of spongy textures in clinopyroxene and spinel from mantle xenoliths, Hessian Depression, Germany. *Mineralogy and Petrology* 74: 149–162.
- Chakraborty S. 1997. Rates and mechanisms of Fe-Mg interdiffusion in olivine at 980° to 1300 °C. *Journal of Geophysical Research* 102:12,317–12,331.
- Clark B. C., Baird A. K., Weldon R. J., Tsusaki S. M., Schanbel L., and Candelaria M. P. 1982. Chemical composition of Martian fines. *Journal of Geophysical Research* 87:10,059–10,067.
- Dodd R. T. and Jarosewich E. 1982. The compositions of incipient shock melts in L6 chondrites. *Earth and Planetary Science Letters* 59:355–363.
- Eugster O., Busemann H., Lorenzetti S., and Terribilini D. 2002. Ejection ages from ^{81}Kr - ^{83}Kr dating and pre-atmospheric sizes of Martian meteorites. *Meteoritics & Planetary Science* 37:1345–1360.
- Federov O. 1995. Cellular pattern during directional growth of a single crystal in a quasi-two-dimensional system. *Journal of Crystal Growth* 156:473–479.
- Franz L. and Wirth R. 2000. Spinel inclusions in olivine of peridotite xenoliths from TUBAF seamount (Bismarck Archipelago/Papua New Guinea): Evidence for the thermal and tectonic evolution of the oceanic lithosphere. *Contributions to Mineralogy and Petrology* 140:283–295.
- Fritz J., Artemieva N., and Greshake A. 2005. Ejection of Martian meteorites. *Meteoritics & Planetary Science* 40:1393–1411.
- Ghiorso M. S. and Sack R. O. 1995. Chemical mass transfer in magmatic processes. IV. A revised and internally consistent thermodynamic model for the interpolation and extrapolation of liquid–solid equilibria in magmatic systems at elevated temperatures and pressures. *Contributions to Mineralogy and Petrology* 119:197–212.
- Gillet Ph., Barrat J. A., Beck P., Marty B., Greenwood R. C., Franchi I. A., Bohn M., and Cotton J. 2005. Petrology, geochemistry and cosmic-ray exposure age of Iherzolitic shergottite Northwest Africa 1950. *Meteoritics & Planetary Science* 40:1175–1184.
- Gnos E., Hofmann B., Franchi I. A., Al-Kathiri A., Hauser M., and Moser L. 2002. Sayh al Uhaymir 094: A new Martian meteorite from the Oman desert. *Meteoritics & Planetary Science* 37:835–854.
- Gooding J. and Muenow D. 1986. Martian volatiles in shergottite EETA79001: New evidence from oxidized sulfur and sulfur-rich aluminosilicates. *Geochimica et Cosmochimica Acta* 50:1049–1059.
- Goodrich C. A. 2003. Petrogenesis of olivine-phyric shergottites

- Sayh al Uhaymir 005 and Elephant Moraine A79001 lithology A. *Geochimica et Cosmochimica Acta* 67: 3735–3771.
- Greshake A., Fritz J., and Stöffler D. 2004. Petrology and shock metamorphism of the olivine-phyrlic shergottite Yamato-980459: Evidence for a two-stage cooling and a single-stage ejection history. *Geochimica et Cosmochimica Acta* 68:2359–2377.
- Heider N. and Kenkmann T. 2003. Numerical simulation of temperature effects at fissures due to shock loading. *Meteoritics & Planetary Science* 38:1451–1460.
- Hirose K. and Kawamoto T. 1995. Hydrous partial melting of lherzolite at 1 GPa: The effect of H₂O on the genesis of basaltic magmas. *Earth and Planetary Science Letters* 133:463–473.
- Kaesler B., Kalt A., and Pettke T. 2007. Crystallization and breakdown of metasomatic phases in graphite-bearing peridotite xenoliths from Marsabit (Kenya). *Journal of Petrology* 47:2149–2184.
- Klügel A. 1998. Reactions between mantle xenoliths and host magma beneath La Palma (Canary Islands): Constraints on magma ascent rates and crustal reservoirs. *Contributions to Mineralogy and Petrology* 131:237–257.
- Klügel A., Hoernle K. A., Schmincke H.-U., and White J. D. L. 2000. The chemically zoned 1949 eruption on La Palma (Canary Islands): Petrologic evolution and magma supply dynamics of a rift zone eruption. *Journal of Geophysical Research* 105:5997–6016.
- Langenhorst F. and Poirier J.-P. 2000. Anatomy of black veins in Zagami: Clues to the formation of high-pressure phases. *Earth and Planetary Science Letters* 184:37–55.
- Laporte D., Schiano P., and Boivin P. 2006. The composition of low degree melts of fertile peridotites at 1 and 1.3 GPa. *EMPG 11*.
- Lofgren G. 1980. Experimental studies on the dynamic crystallization of silicate melts. In *Physics of magmatic processes*, edited by Hargraves E.B. Princeton: Princeton University Press. pp. 487–551.
- Maiwa K., Nakamura H., Kimura H., and Miyazaki A. 2003. In situ observation of the crystallization via incongruent melting and peritectic reaction in Sr(NO₃)-H₂O system. *Journal of Crystal Growth* 255:379–385.
- McCoy T. J., Taylor G. J., and Keil K. 1992. Zagami—Product of a two-stage magmatic history. *Geochimica et Cosmochimica Acta* 56:3571–3582.
- McSween H. Y. Jr. 1994. What have we learned about Mars from SNC meteorites. *Meteoritics & Planetary Science* 29:757–779.
- McSween H. Y. Jr. and Jarosewich E. 1986. Petrogenesis of the Elephant Moraine A79001 meteorite: Multiple magma pulses on the shergottite parent body. *Geochimica et Cosmochimica Acta* 47:1501–1513.
- McSween H. Y. Jr. and Treiman A. H. 1995. Martian meteorites. In *Planetary materials*, edited by Papike J. Reston, Virginia: Mineralogical Society of America. pp. 6–9–6–53.
- Melosh H. 1985. Ejection of rock fragments from planetary bodies. *Geology* 13:144–148.
- Mikouchi T., Miyamoto M., and McKay F. A. 1997. Similarities in zoning of pyroxene from QUE 94201 and EET 79001 martian meteorites (abstract). 28th Lunar and Planetary Science Conference. pp. 955–956.
- Mysen B. O. and Boettcher A. L. 1975. Melting of a hydrous mantle: I. Phase relations of natural peridotite at high pressures and temperatures with controlled activities of water, carbon dioxide, and hydrogen. *Journal of Petrology* 16:520–548.
- Putirka K. 1997. Magma transport at Hawaii: Inferences based on igneous thermobarometry. *Geology* 25:69–72.
- Putirka K. 1999. Clinopyroxene + liquid equilibrium to 100 kbar and 2450 K. *Contributions to Mineralogy and Petrology* 135:151–163.
- Putirka K., Johnson M., Kinzler R., Longhi J., and Walker D. 1996. Thermobarometry of mafic igneous rocks based on clinopyroxene-liquid equilibria, 0–30 kbar. *Contributions to Mineralogy and Petrology* 123:92–108.
- Putirka K., Ryerson F. J., and Mikaelian H. 2003. New igneous thermobarometers for mafic and evolved lava compositions, based on clinopyroxene + liquid equilibria. *American Mineralogist* 88:1542–1554.
- Rao M., Borg L. E., McKay D. S., and Wentworth S. J. 1999. Martian soil component in impact glasses in a Martian meteorite. *Geophysical Research Letters* 26:3265–3268.
- Rieder R. T., Economou T., Wänke H., Turkevich A., Crisp J., Brückner, Dreibus G., and McSween H. Y. Jr. 1998. The chemical composition of Martian soil and rocks returned by the mobile alpha proton X-ray spectrometer: Preliminary results from the X-ray mode. *Science* 278:1771–1774.
- Roeder P. L., Poustovetov A. A., and Oskarsson N. 2001. Growth forms and composition of chromian spinel in MORB magma: Diffusion-controlled crystallization of chromian spinel. *Canadian Mineralogist* 39:397–416.
- Schiano P. and Clocchiatti R. 1994. Worldwide occurrence of silica-rich melts in sub-continental and sub-oceanic mantle minerals. *Nature* 368:621–624.
- Shaw C. S. J. 1999. Dissolution of orthopyroxene in basanitic magma between 0.4 and 2 GPa: Further implications for the origin of Si-rich alkaline glass inclusions in mantle xenoliths. *Contributions to Mineralogy and Petrology* 135:114–132.
- Shaw C. S. J. 2000. The effect of experiment geometry on the mechanism and rate of dissolution of quartz in basanite at 0.5 GPa and 1350 °C. *Contributions to Mineralogy and Petrology* 139:509–525.
- Shaw C. S. J. 2004. The temporal evolution of three magmatic systems in the West Eifel volcanic field, Germany. *Journal of Volcanology and Geothermal Research* 131:213–240.
- Shaw C. S. J. and Dingwell D. B. 2007. Experimental peridotite-melt reaction at one atmosphere: A textural and chemical study. *Contributions to Mineralogy and Petrology*, doi 10.1007/s00410-007-0237-1.
- Shaw C. S. J. and Dingwell D. B. 2008. Experimental peridotite-melt reaction at one atmosphere: A textural and chemical study. *Contributions to Mineralogy and Petrology* 155:199–214.
- Shaw C. S. J. and Edgar A. D. 1997. Post-entrapment mineral-melt reactions in spinel peridotite xenoliths from Inver, Donegal, Ireland. *Geological Magazine* 134:771–779.
- Shaw C. S. J., Heidelberg F., and Dingwell D. B. 2006. The origin of reaction textures in mantle peridotite xenoliths from Sal Island, Cape Verde: the case for “metasomatism” by the host lava. *Contributions to Mineralogy and Petrology* 151:681–697.
- Shaw C. S. J. and Klügel A. 2002. The pressure and temperature conditions and timing of glass formation in mantle-derived xenoliths from Baarley, West Eifel, Germany: The case for amphibole breakdown, lava infiltration and mineral—melt reaction. *Mineralogy and Petrology* 74:163–187.
- Shaw C. S. J., Thibault Y., Edgar A. D., and Lloyd F. E. 1998. Mechanisms of orthopyroxene dissolution in silica-undersaturated melts at 1 atmosphere and implications for the origin of silica-rich glass in mantle xenoliths. *Contributions to Mineralogy and Petrology* 132:354–370.
- Stewart S., Seifert A., Kennedy G., Furlanetto M., and Obst A. 2007. Measurements of emission temperatures from shocked basalt: Hot spots in meteorites (abstract #2413). Lunar and Planetary Science Conference, CD-ROM.
- Stöffler D. 1984. Glasses formed by hypervelocity impact. *Journal of Non-Crystalline Solids* 67:465–502.
- Stöffler D., Keil K., and Scott E. R. D. 1991. Shock metamorphism

- of ordinary chondrites. *Geochimica et Cosmochimica Acta* 55: 3845–3867.
- Stöffler D., Ostertag R., Jammes C., Pfannschmidt G., Sen Gupta P. R., and Beauchamp R. H. 1986. Shock metamorphism and petrography of the Shergotty achondrite. *Geochimica et Cosmochimica Acta* 50:889–903.
- Tracy R. J. 1980. Petrology and significance of an ultramafic xenolith suite from Tahiti. *Earth and Planetary Science Letters* 48:80–96.
- Walton E. L. and Herd C. D. K. 2007a. Dynamic crystallization of shock melts in Allan Hills 77005: Implications for melt pocket formation in Martian meteorites. *Geochimica et Cosmochimica Acta* 71:5267–5285.
- Walton E. L. and Herd C. D. K. 2007b. Localized shock melting in lherzolitic shergottite Northwest Africa 1950: Comparison with Allan Hills 77005. *Meteoritics & Planetary Science* 42: 63–80.
- Walton E. L. and Spray J. G. 2003. Mineralogy, microtexture, and composition of shock-induced melt pockets in the Los Angeles basaltic shergottite. *Meteoritics & Planetary Science* 38:1865–1875.
- Walton E. L., Spray J. G., and Bartoschewitz R. 2005. A new Martian meteorite from Oman: Mineralogy, petrology and shock metamorphism of olivine-phyric basaltic shergottite Sayh al Uhaymir 150. *Meteoritics & Planetary Science* 40:1195–1214.
- Walton E. L., Kelley S. P., and Spray J. G. 2007. Shock implantation of Martian atmospheric argon in four basaltic shergottite: A laser probe $^{40}\text{Ar}/^{39}\text{Ar}$ investigation. *Geochimica et Cosmochimica Acta* 71:497–520.
- Walton E. L., Shaw C. S. J., and Spray J. G. 2006. Crystallization rates of shock melts in three martian basalts: Experimental simulation with implications for meteoroid dimensions. *Geochimica et Cosmochimica Acta* 70:1059–1075.
- Wiens R. C. and Pepin R. O. 1988. Laboratory shock emplacement of noble gases, nitrogen, and carbon dioxide into basalt, and implications for trapped gases in shergottites EET 79001. *Geochimica et Cosmochimica Acta* 52:295–307.
-