EIGHT

Shock metamorphism

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8.1 Introduction

A requirement for the recognition and confirmation of meteorite impact structures is the presence of shock metamorphic indicators, either megascopic (e.g. shatter cones) or microscopic (e.g. planar deformation features in minerals), or high-pressure polymorphs (e.g. coesite and stishovite) and/or siderophile element (e.g. iridium) or isotopic (osmium) anomalies in specific geological settings. Crater morphology is not a sufficient argument, because a variety of circular features can be formed by completely different geological processes (e.g. volcanism or salt diapirism). The terms 'shock effects' or 'shock-metamorphic effects' cover all types of shock-induced changes, such as the formation of planar microstructures and phase transformations. Impact metamorphism is essentially the same as shock metamorphism, except that it also encompasses the melting, decomposition and vaporization of target rocks (Stöffler and Grieve, 2007). These irreversible changes are produced when rocks are subjected to shock pressures above their Hugoniot elastic limit (HEL). This limit is defined as 'the critical shock pressure at which a solid yields under the uniaxial strain of a plane shock wave' (Stöffler, 1972). The HEL of quartz is in the range of 5–8 GPa and ranges from approximately 1 to 10 GPa for most geological materials (e.g. Stöffler, 1972; Stöffler and Langenhorst, 1994). In nature, at the surface of the Earth, only hypervelocity impacts can generate such high shock pressures.

A great diversity of shock effects in minerals are known and have been abundantly described in the literature during the last 40 years mostly for quartz (e.g. French and Short, 1968; von Engelhardt and Bertsch, 1969; Stöffler, 1972; Stöffler and Langenhorst, 1994; Grieve *et al.*, 1996; French, 1998; Montanari and Koeberl, 2000; Langenhorst, 2002; and references therein) and to some extent for feldspar (e.g. Stöffler, 1967; Robertson, 1975; Ostertag, 1983; Dressler, 1990; Bischoff and Stöffler, 1992; and references therein), olivine (e.g. Reimold and Stöffler, 1978; Bauer, 1979; Stöffler *et al.*, 1991; Bischoff and Stöffler, 1992; Schmitt, 2000; and references therein), and pyroxene (e.g. Rubin *et al.*, 1997), mostly within meteorites. Less is known about the effects of shock in other minerals, such as the nesosilicates, double-chain inosilicates, phyllosilicates and the carbonate and sulfate minerals in general. Much of our knowledge on shock metamorphism comes from studies of minerals from impact craters formed in dense, non-porous crystalline rocks and from more or less porous meteorites. However, the effects of target lithology on the response of minerals to shock compression remain to be investigated in detail. To date, in the case of sedimentary rocks, only the response of quartz has been investigated in any detail (Table 8.1) – see Kieffer (1971), Kieffer *et al.* (1976) and Osinski (2007). These observational studies, together with recent numerical simulations (e.g. Wünnemann *et al.*, 2008), have revealed the complicating effects of porosity and volatiles on the response of quartz to impact in sedimentary targets. This will be discussed below, but the main observation is that large differences in shock impedance between solid grains and pore space results in a very heterogeneous distribution of shock wave energy at the microscopic scale. This has the effect of collapsing pore spaces and compressing grain boundaries and producing shock metamorphic effects of very different shock levels within an individual sample. In addition, it is apparent that energy is preferentially transferred into heat and melting of the target lithologies, as opposed to forming shock metamorphic effects in minerals, particularly in poorly or unconsolidated sediments and sedimentary rocks.

This chapter provides an overview of shock metamorphic effects in dense non-porous crystalline rocks and minerals and a comparison with the effects in sedimentary rocks. In addition, two examples of post-shock thermal effects/features, namely toasted quartz and ballen silica, for which occurrence is restricted to impact-derived rocks, are also discussed.

Impact Cratering: Processes and Products, First Edition. Edited by Gordon R. Osinski and Elisabetta Pierazzo.

Table 8.1 Classification of impact metamorphic effects in sandstones. Compiled with data from Barringer Crater (Kieffer, 1971; Kieffer *et al.*, 1976) and the Haughton impact structure (Osinski, 2007)

Class	Pressure range (GPa)[a]	Temperature range (°C)[a]	Hand specimen observations	Proportion of various SiO_2 phases and effects[b]					Microscopic observations
				Qtz	Dg	Lech	Coesite	Toast	
1a	<3	<350	Crude shatter cones	****	—	—	—	—	Recognizable porosity; no indications of shock
1b	3–5.5	<350	Well-developed shatter cones	****	—	—	—	—	No recognizable porosity using optical microscopy; fracturing of quartz grains; minor development of PFs
2	5.5–10	350–950	Well-developed shatter cones	***	*	—	*	*	Presence of 'jigsaw' texture; fracturing of individual grains and generation of micro-breccias; symplectic regions along some grain boundaries; PFs; PDFs
3a	10–20	>1000	Difficult to discern individual grains	***	**	—	**	**	Reduced grain size; multiple sets of PDFs; symplectic regions surrounding majority of quartz grains
3b	15–20	>1000	Difficult to discern individual grains	**	***	*	*	**	Widespread development of diaplectic glass; vesicular SiO_2 glass in original pore spaces
4	20–30	>1000	Few recognizable grains; faint layering white	**	***	**	*	**	Original texture of sandstone lost; major development of vesicular SiO_2 glass; symplectic regions surround all remaining quartz grains
5a	>30	>1000	Highly vesicular; white	*	**	***	*	*	Isolated remnant quartz grains; almost complete transformation to diaplectic or vesicular SiO_2 glass (lechatelierite)
5b	>30	>1000	Highly vesicular; white	—	*	***	*	—	Complete transformation to vesicular SiO_2 glass
6	>30	>1000	Dense; grey; translucent	****	—	—	—	—	Recrystallized SiO_2 glass

[a]Pressures and post-shock temperatures from Kieffer *et al.* (1976).
[b]Qtz: quartz; Dg: diaplectic quartz glass; Lech: lechatelierite or vesicular SiO_2 glass; Toast: toasted quartz; ****: only phase present (>99%); ***: abundant; **: present; *: rare (<10%); —: absent.

8.2 Shock metamorphic features

Minerals subjected to shock metamorphism occur in different petrographic assemblages and in different rock types. The full spectrum of diagnostic features may not necessarily be present in all impact structures and is strongly dependent on the lithology and other properties of the target rock(s), and is a function of the magnitude of the hypervelocity impact and of the level of erosion of the crater. The different rocks affected (or produced) by one or more hypervelocity impact(s) are called 'impactites' (see Chapter 7 and reviews by French (1998) and Stöffler and Grieve (2007)). The classification and definition of the various impactites is complex and ongoing; nevertheless, the name of the different rock types described in this chapter follows the classification by Stöffler and Grieve (2007) and is discussed further in Chapter 7.

8.2.1 Shatter cones

Shatter cones are the only distinctive shock-deformation feature (i.e. diagnostic evidence of hypervelocity impact) that can be seen with the naked eye (e.g. Dietz, 1960, 1968; French, 1998; French and Koeberl, 2010). These meso- to macro-scale features, consisting of conical striated fracture surfaces, are best developed in fine-grained lithologies (such as limestone; see Fig. 8.1), but can also be observed in coarser grained lithologies, such as granites and gneisses, although they are typically more poorly developed. By definition, shatter cones are 'distinctive curved, striated fractures that typically form partial to complete cones' (French, 1998). The striated surface of shatter cones is either a positive or a negative feature (Fig. 8.1e), with the striations radiating along the surface of the cone. The occurrence of shatter cones has been reported to date for more than half of the currently confirmed impact structures; they usually occur in the central uplifts of complex impact structures, and in some cases isolated fragments/clasts of shatter cones have been found in impact breccias, within or outside the crater (e.g. at the Haughton impact structure, Canada; Osinski and Spray, 2006; see Fig. 8.1b,c,e). The distribution of *in situ* shatter cones at an impact site has also been used as a parameter for estimating the original size of a structure, particularly for old and eroded impact sites, as they occur generally below the crater floor or in the central uplifts of large structures. It is generally accepted that, when restored to their original position prior to the impact, shatter cones' apexes indicate the point of impact. However, shatter cones are generally found as composite groups (rarely as single specimens) of commonly partial to complete cones, with very frequently opposite orientations at the centimetre to decimetre scale. Thus, the use of shatter cone apex orientation to determine the centre of a crater and then its size is likely to yield incorrect results.

The formation of shatter cones, widely accepted as unequivocal proof of a meteorite impact crater, is still not completely resolved (e.g. Dietz, 1960; Johnson and Talbot, 1964; Gash, 1971; Milton, 1977; Baratoux and Melosh, 2003; Sagy *et al.*, 2004; Wieland *et al.*, 2006), but current formation hypotheses suggest that shatter cones originate very early during the impact process. It is generally accepted that shatter cones form at relatively low shock pressures, typically between approximately 2 and 10 GPa (e.g. French, 1998). At the microscopic scale, planar fractures (PFs) and/or planar deformation features (PDFs – see below) occur in minerals in rocks containing shatter cones (e.g. Wieland *et al.*, 2006; Fackelman *et al.*, 2008; Ferrière and Osinski, 2010).

In some cases, shatter cones are not well formed, and it may be hard for non-experts to distinguish them from non-impact features such as cone-in-cone structures, ventifacts or even slickenslides – see discussion in French and Koeberl (2010). Concerning the cone-in-cone structures (Fig. 8.2a), they form uniquely in sedimentary rocks, and, typically, the cone axes are normal to the bedding (see Lugli *et al.* (2005)), whereas shatter cones form in all rock types and at various angles to pre-existing rock. Ventifacts are features commonly found in hot and cold deserts and are the result of preferential abrasion under the effect of the dominant winds (e.g. Higgins, 1956). Wind-abrasion features can resemble shatter cones (Fig. 8.2b); however, they develop only on outcrop surfaces (i.e. they are not penetrative) and, thus, can easily be discriminated from shatter cones. Finally, because the striations in slickensides tend to be parallel, they can also be easily discriminated from shatter cones, in which striations are divergent.

8.2.2 Deformation in quartz

8.2.2.1 Planar microstructures

Upon shock compression, quartz develops irregular fractures (which are not diagnostic shock effects) and planar microstructures. Planar microstructures in quartz are divided into PFs and PDFs (e.g. French and Short, 1968; von Engelhardt and Bertsch, 1969; Stöffler and Langenhorst, 1994; Grieve *et al.*, 1996; French, 1998). Both types of planar microstructures are crystallographically controlled; therefore, PFs and PDFs are oriented parallel to rational crystallographic planes. The four-digit notation (*hkil*), the so-called Miller–Bravais indices, is used for the indexing of the planes in the crystal. The indices (*hkil*) represent the inverse plane intercepts along the a_1, a_2, a_3 and c axes respectively (e.g. Bloss, 1971).

8.2.2.2 Planar fractures (PFs)

PFs are by definition planar, parallel, thin open fissures, generally greater than 3 μm wide and spaced more than 15–20 μm apart (see Fig. 8.3). The spacing between PFs can vary on the order of a few micrometres, but is wider than that of PDFs (Stöffler and Langenhorst, 1994; Grieve *et al.*, 1996; French, 1998; Montanari and Koeberl, 2000; Langenhorst, 2002; Morrow, 2007). They have been documented from both crystalline and sedimentary rocks from a variety of impact structures. The PFs are oriented parallel to rational crystallographic planes, such as (0001) and $\{10\bar{1}1\}$, and occasionally to $\{10\bar{1}3\}$.

It is notable that PFs commonly control and/or limit the distribution of adjacent PDF sets, which has been used to suggest that PF formation pre-dates PDF formation in a given quartz crystal grain (e.g. von Engelhardt and Bertsch, 1969; Stöffler and Langenhorst, 1994). Formed at pressures of approximately 5–8 GPa, PFs are not regarded as unambiguous evidence of shock

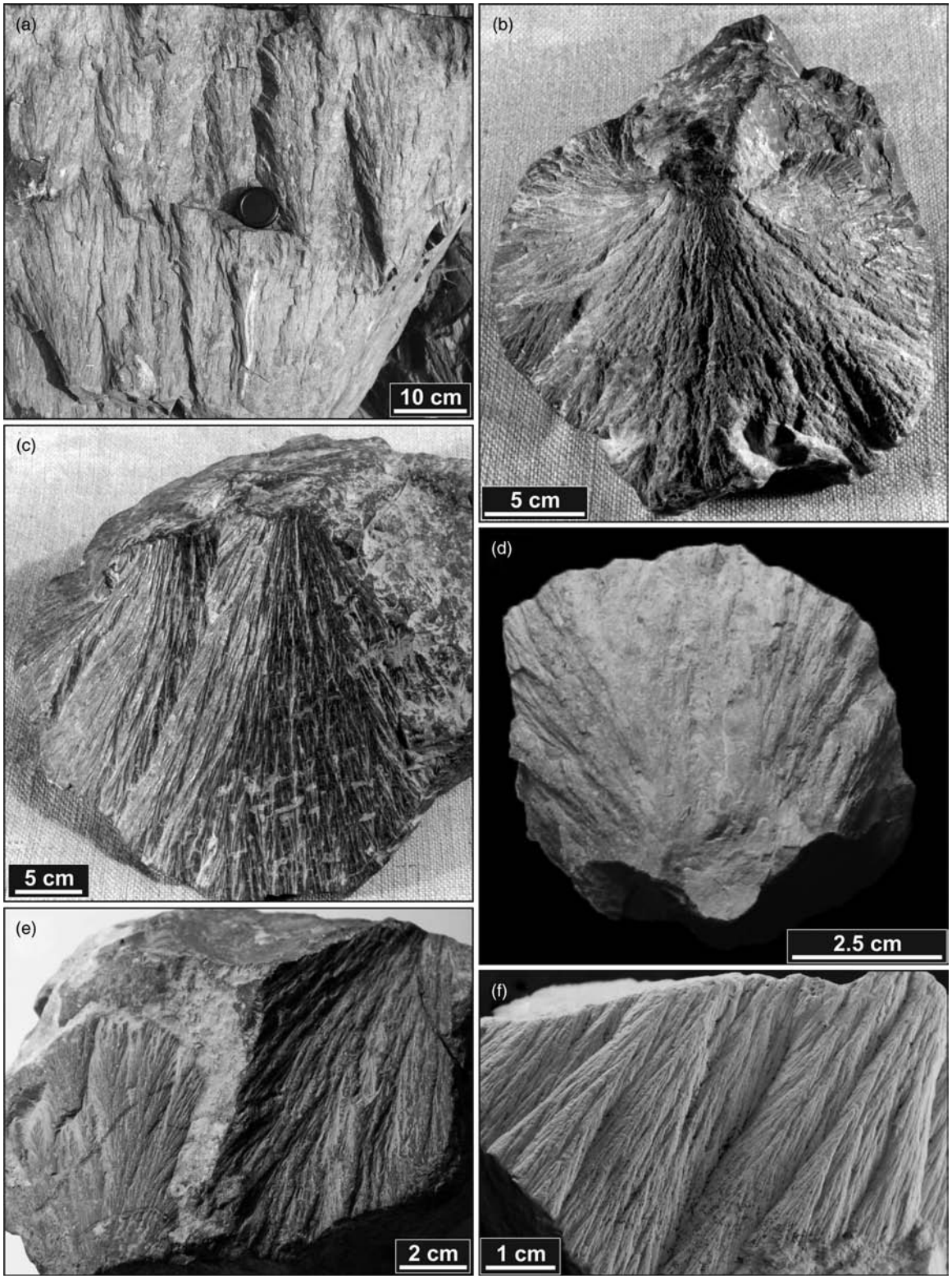

Figure 8.1 Photographs of typical shatter cones from different impact structures. (a) Exposure view of shatter cones in quartzite from the Sudbury structure (Canada). Note that shatter cone apices are oriented in one direction on this small outcrop-sized area. (b) A complete cone developed in limestone with apex pointing up and showing the typical divergence of striae away from the cone apex. (c) Two nicely developed partial shatter cones. (d) Hand specimen of shatter cone from the Eagle Butte structure (Canada). (e) Positive (convex) and negative (concave; 'cast') of shatter cone surfaces; note that apices point in opposite directions. (f) Typical horsetailing shatter cone surfaces developed in fine-grained limestone from the Steinheim structure (Germany). Samples illustrated in (b, c, e) are shatter cone clasts from crater-fill impact melt rocks from the Haughton structure (Canada).

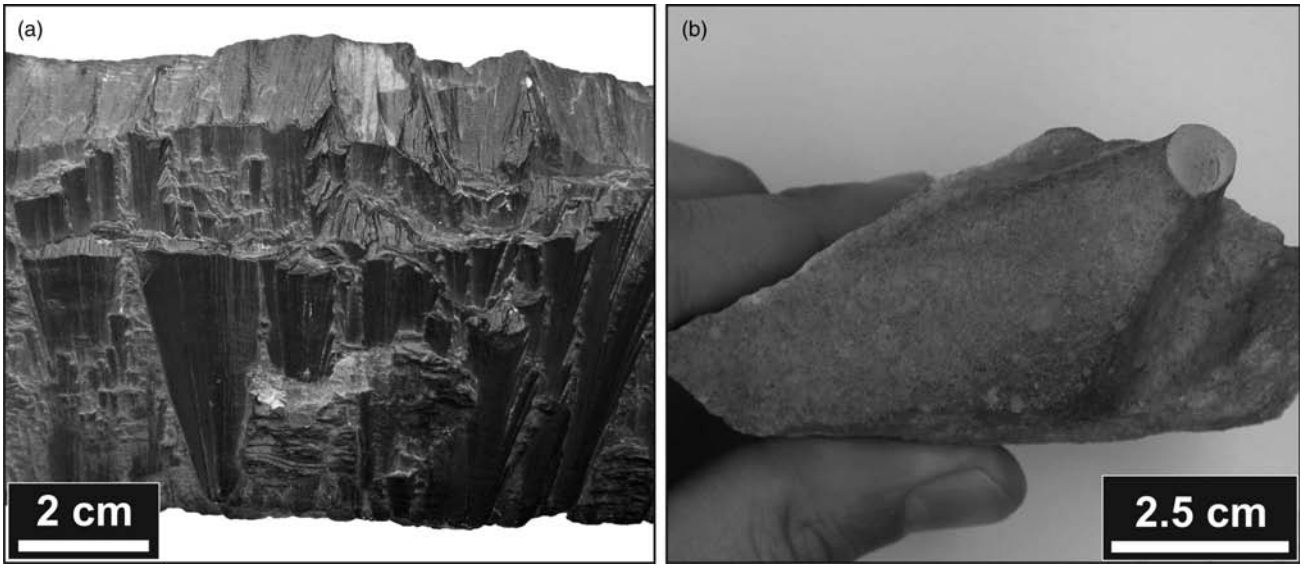

Figure 8.2 Photographs of non-impact-related conical features frequently misidentified as shatter cone. (a) Typical silicified cone-in-cone structures (sedimentary in origin) from the Tafilalt region of Morocco (see Lugli *et al.*, 2005). Photograph courtesy of S. Lugli. (b) Ventifacts (wind abrasion features) developed in quartzitic breccia (sample from the GKCF02 structure, Egypt). Note the presence of a clast at the apex of the cone.

metamorphism, as they also occur, rarely, in quartz grains from non-impact settings (e.g. French, 1998).

8.2.2.3 Feather features

In the case of some impact structures (in at least 26 according to Poelchau and Kenkmann (2011)) developed within both sedimentary and crystalline target rocks, the occurrence of thinly spaced, short, parallel to subparallel lamellae (Poelchau and Kenkmann, 2011) or incipient PDFs (French *et al.*, 2004) that branch off of PFs (Fig. 8.3c,d) have been observed. This somewhat unusual type of planar microstructure, called 'feather features', appears to be shock related, but its formation mechanism is poorly understood. In a recent publication, Poelchau and Kenkmann (2011) show that these features are crystallographically controlled to a certain degree and they suggest that these microstructures are caused by shearing of planar fractures during shock deformation; however, further investigations are necessary to validate this hypothesis.

8.2.2.4 Planar deformation features (PDFs)

PDFs in quartz grains, which develop over the pressure range of 5–10 GPa to approximately 35 GPa (see Stöffler and Langenhorst (1994) and French (1998) and references therein), are one of the best criteria for the identification of new impact structures. In contrast to PFs, PDFs are not open fractures. PDFs are typically composed of narrow, individual planes of amorphous material that are less than 2 μm thick, comprising straight, parallel sets spaced 2–10 μm apart (e.g. von Engelhardt and Bertsch, 1969; Stöffler and Langenhorst, 1994). Generally occurring as multiple sets per grain, and typically in more than one crystallographic orientation, PDFs can be either decorated or non-decorated (Fig. 8.4). The PDFs are generally decorated with tiny fluid inclusions or bubbles, usually less than 2 μm in diameter, which greatly facilitate their detection at the scale of the optical microscope. Decorated PDFs are considered secondary features, in which the decorations form by post-shock annealing and aqueous alteration of non-decorated amorphous PDFs (e.g. Stöffler and Langenhorst, 1994; Grieve *et al.*, 1996; Leroux, 2005).

PDFs are preferentially oriented parallel to rational crystallographic planes, such as $\{10\bar{1}3\}$, $\{10\bar{1}2\}$, (0001), $\{10\bar{1}1\}$, $\{11\bar{2}2\}$, $\{11\bar{2}1\}$, $\{21\bar{3}1\}$, $\{51\bar{6}1\}$, $\{10\bar{1}0\}$ and $\{11\bar{2}0\}$, and more rarely to other planes; the measurement of PDF orientations is possible using transmission electron microscopy (TEM; e.g. Goltrant *et al.*, 1991), as well as with the spindle stage (e.g. Bohor *et al.*, 1987), or using the universal stage technique (e.g. Ferrière *et al.*, 2009a). As specific orientations of PDFs in quartz are formed at different shock pressures (e.g. Hörz, 1968; Müller and Défourneaux, 1968; Huffman and Reimold, 1996), several workers, including Robertson and Grieve (1977), Grieve *et al.* (1990) and Dressler *et al.* (1998), have derived average shock pressure values for a given sample, based on laboratory shock experiments that bracket the pressure ranges associated with the development of individual PDF orientations or assemblages of orientations in quartz grains. For summaries of this method and references, see reviews by Stöffler and Langenhorst (1994) and Grieve *et al.* (1996); however, De Carli *et al.* (2002) and Ferrière *et al.* (2009a) provide a cautionary discussion regarding the potential limitations of this technique.

Because PDFs cannot be clearly resolved under the optical microscope, TEM techniques are required for the characterization of their microstructure (e.g. Kieffer *et al.*, 1976; Goltrant *et al.*, 1991; Trepmann and Spray, 2006). Based on TEM investigations, Kieffer *et al.* (1976) and later Goltrant *et al.* (1991) showed that

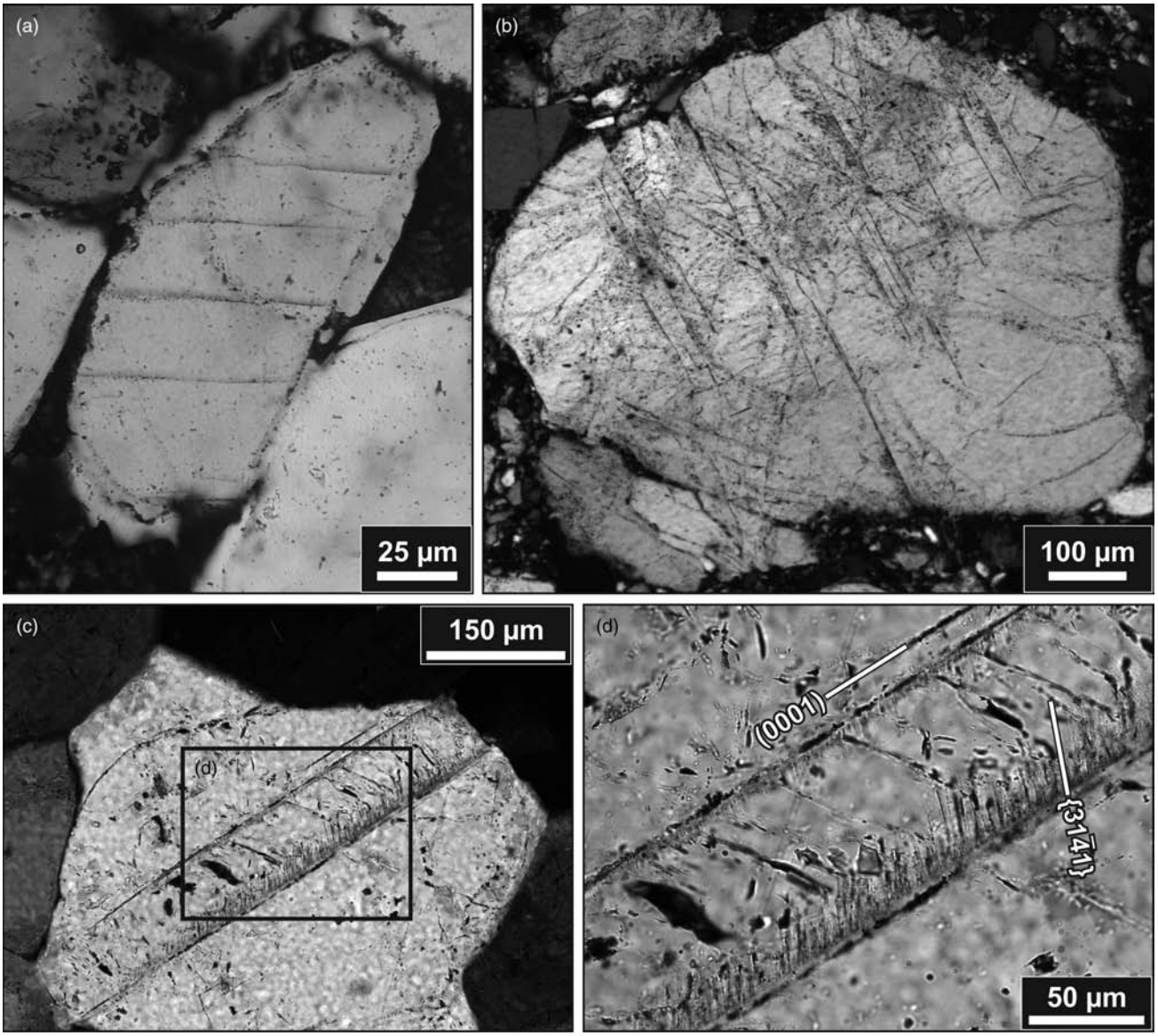

Figure 8.3 Microphotographs (crossed polars) of quartz grains with PFs. (a) Authigenic quartz grain with one set of typically spaced PFs (sandstone sample from the Libyan Desert Glass strewn field, Egypt). (b) Quartz grain with one prominent set of planar fractures oriented NW–SE; other irregular fractures are also visible in the grain (sample OR-10; Aorounga structure, Chad). (c) Quartz grain showing one set of *c*(0001) PFs (oriented NE–SW) with feather features that branch off the PF (sandstone sample HMP-04-055; Haughton, Canada). (d) Enlarged part of (c) showing 'feather features' with {31$\bar{4}$1}-equivalent orientation. Directions (e.g. NW–SE) are in relation to an arbitrary 'north' at the top of the image.

PDFs developing parallel to the basal plane (0001) represent mechanical Brazil twins. These Brazil twins are generated at high shear stresses on the basal plane and result from glide motion of the partial dislocations (McLaren *et al.*, 1967).

Planar deformation features form in both crystalline and sedimentary target rocks, although in the latter they have not received much attention and systematic studies are lacking. As such, there are several unanswered questions concerning PDF development in sedimentary rocks. For example, it is clear that PDFs are missing or rare in some craters (e.g. Meteor Crater) but are common in others (e.g. the B.P. and Oasis structures; Grieve *et al.*, 1996). The development of PDFs in sedimentary rocks seems to depend on the grain size of quartz grains, as PDFs are preferentially observed in grains with larger diameters (Grieve *et al.*, 1996). However, a similar grain size effect is reported for leucosome samples from the Ries crater (Walzebuck and von Engelhardt, 1979) and for gneisses from the Charlevoix structure (Trepmann and Spray, 2006). In addition, the orientation of PDFs appears to differ in sedimentary rocks, with some suppression or reduction of some orientations (e.g. {10$\bar{1}$3}) in favour of others (e.g. {10$\bar{1}$1}, {11$\bar{2}$2}) (Robertson, 1980). To our knowledge, unequivocal PDFs have not been recognized at impact sites developed

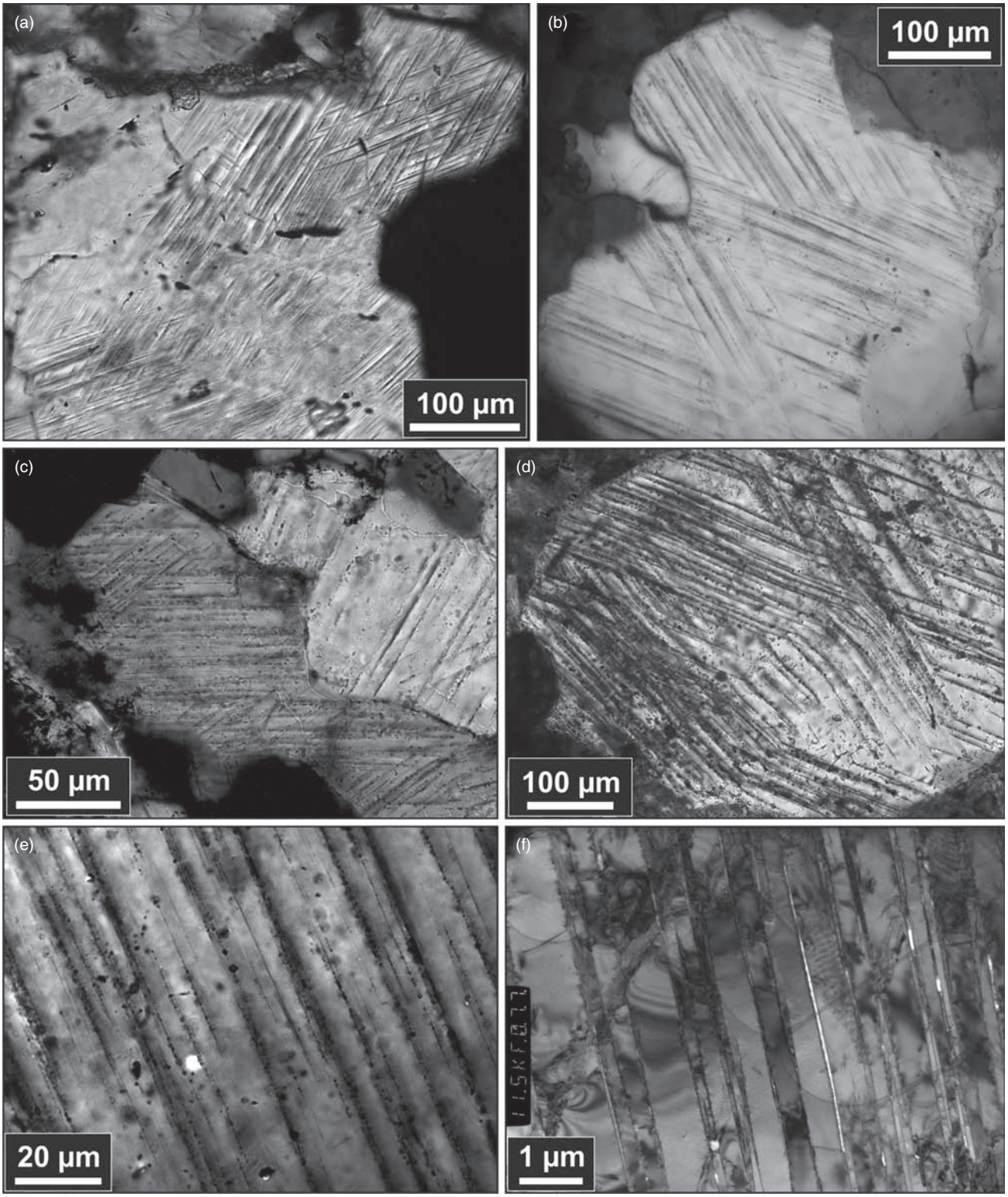

Figure 8.4 Microphotographs of quartz grains with sets of PDFs. (a) Quartz grain with two relatively non-decorated PDF sets. Quartzite clast in suevite from the Bosumtwi crater (sample KR8-006; depth: 240.36 m). (b) Quartz grain with two PDF sets. Meta-greywacke sample from the Bosumtwi crater (sample KR8-080; depth: 384.54 m). (c) Two sets of decorated PDFs in a quartz grain from a meta-greywacke sample from the Bosumtwi crater (sample KR8-070; depth: 364.12 m). (d) Quartz grain with two PDF sets. Fragmental dike breccia sample from the Manson structure (USA; sample M8-516.3; depth: 157.37 m). (e) One set of decorated (with numerous tiny fluid inclusions) PDFs in a quartz grain from a meta-greywacke sample from the Bosumtwi crater (sample KR8-056; depth: 326.78 m). Microphotographs (a–c) taken in crossed polars and (d, e) in plane-polarized light. (f) TEM bright-field microphotograph of one set of irregularly spaced PDFs in quartz (Bosumtwi sample KR8-006). Note that the light grey network shown in the background corresponds to the carbon net supporting the specimen. (See Colour Plate 20)

in unconsolidated sedimentary materials (e.g. Henbury (Taylor, 1967) and Wabar (Hörz *et al.*, 1989)), although the lack of detailed studies leaves this question open.

8.2.2.5 Mosaicism

Mosaicism or mosaic structure is characterized by an irregular or mottled optical extinction pattern, which is distinctly different from undulatory extinction (commonly developed in tectonically deformed quartz). A crystal showing mosaicism comprises several sub-domains with slightly different optical axes (e.g. Dachille *et al.*, 1968; Stöffler, 1972; Stöffler and Langenhorst, 1994; French and Koeberl, 2010), as a result of the distortion (i.e. plastic deformation) of the lattice into small domains that are rotated by low angles against each other. Deformation bands (i.e. bands generally <20 μm across showing extinction directions different from those of the host; also called 'kink bands; e.g. see French and Koeberl (2010)) and/or PFs and PDFs are generally associated with mosaicism (e.g. Stöffler, 1972). Mosaicism can be semi-quantitatively characterized by the X-ray diffraction study of the degree of asterism (i.e. broadening of the normally sharp diffraction spots of lattice planes into elongate spots) in a single crystal, which can be useful for the characterization of shock pressure recorded by minerals (e.g. Hörz and Quaide, 1973). However, no correlation between pressure and degree of mosaicism is valid if minerals were recrystallized during post-shock annealing or subsequent thermal metamorphism (Stöffler, 1972). Even if mosaicism is definitely induced by shock during an impact event, a structure somewhat resembling a mosaic can also be produced by endogenic processes (e.g. Spry, 1969); thus, it cannot be used as a unique diagnostic indicator of shock metamorphism.

8.2.2.6 Refractivity, birefringence and density

Optical properties, such as refractivity and birefringence of quartz, have been intensively investigated in the past – see reviews by Stöffler (1974) and Stöffler and Langenhorst (1994). It was shown that, with increasing shock pressure, the birefringence and the refractive index both decrease simultaneously, until the amorphous state (i.e. diaplectic glass) is reached. Similarly, there is a decrease of the density of quartz with increasing shock pressure, from its normal value of $2.650 \pm 0.002\,\mathrm{g\,cm^{-3}}$ to values as low as $2.280 \pm 0.002\,\mathrm{g\,cm^{-3}}$ for shocked quartz (Langenhorst and Deutsch, 1994). The same authors show that this significant drop in density, between 25 and 35 GPa, depends on the pre-shock temperature, as well as on the orientation of the shock wave relative to the *c*-axis of the quartz crystal.

A unique feature of sandstones is the formation of the characteristic 'jigsaw' texture at pressures less than 10 GPa (Kieffer, 1971). This texture is thought to form due to rotation and shear at quartz grain boundaries, which leaves the interiors of grains relatively undamaged (Kieffer *et al.*, 1976).

8.2.3 Deformation in other minerals

Shock-induced deformation occurs in all minerals; deformation that largely depends on the crystal structure and on the composition of the mineral (Stöffler, 1972; Langenhorst, 2002). Two main types of microstructures are formed; namely planar microstructures (i.e. PFs and PDFs) and deformation bands (i.e. kink bands and mechanical twins). Mosaicism, while mainly described for quartz, is also observed in several minerals, such as olivine and pyroxene (e.g. Reimold and Stöffler, 1978; Bauer, 1979; Rubin *et al.*, 1997). These features have been poorly investigated and characterized in minerals other than quartz, with the notable exception of olivine in meteorites (Reimold and Stöffler, 1978; Bauer, 1979; Stöffler *et al.*, 1991; Bischoff and Stöffler, 1992; Schmitt, 2000; and references therein), possibly due to the complexity of such features in other minerals and/or because of the obliteration of these features by secondary alteration.

8.2.3.1 Planar microstructures

Aside from quartz, deformation in feldspar is most commonly reported in the literature (e.g. Chao, 1967; Stöffler, 1967, 1972; French, 1998). With increasing shock pressure, fracturing, plastic deformation and PFs, or more frequently PDFs, in both plagioclase and alkali-feldspar occur (Fig. 8.5). Similarly, PDFs have been observed in olivine, pyroxene, amphibole, sillimanite, garnet and apatite (e.g. Stöffler, 1972; French, 1998; Langenhorst, 2002; and references therein). It is notable that PFs and PDFs have not been documented to date in carbonate or sulfate minerals. Stöffler (1970) showed that PFs and PDFs in sillimanite are oriented parallel to bipyramids, prisms and pinacoids. Decorated PDFs have also been observed in feldspar and in amphibole (Stöffler, 1972), and even in zircon grains (e.g. Wittmann *et al.*, 2006 and references therein). It is likely that, as for quartz, PDFs are preferentially oriented parallel to rational crystallographic planes for most rock-forming minerals; however, the lack of studies for minerals other than quartz leaves this question open. In addition, it is unclear as to whether there are clear relationships between shock pressure and specific orientations of resultant PDFs in minerals other than quartz.

Olivine, rarely present in terrestrial impactites, is an important diagnostic mineral used for the classification of shock metamorphism in stony meteorites (e.g. Stöffler *et al.*, 1991; Scott *et al.*, 1992). Olivine with PFs typically parallel to low-index planes is abundant in shocked meteorites (see Langenhorst (2002) and references therein). Importantly, in contrast to quartz, the presence of PFs in olivine is accepted as being indicative of shock, as they are oriented parallel to rational crystallographic planes that do not correspond to the normal cleavage planes of olivine (Langenhorst, 2002).

Zircon grains with shock-induced deformation are also reported in impactites that were subject to shock pressures higher than 20 GPa (e.g. Bohor *et al.*, 1993; Kamo *et al.*, 1996; Wittmann *et al.*, 2006), including mainly planar microdeformation features, such as pervasive micro-cleavage (Fig. 8.6) and dislocation patterns (see Leroux *et al.* (1999)), and granular (or 'strawberry') textures (e.g. Bohor *et al.*, 1993; Gucsik *et al.*, 2004; Wittmann *et al.*, 2006).

8.2.3.2 Kind bands

Kink bands are frequently observed in mica, for example in muscovite and in biotite (Cummings, 1965; Chao, 1967; Hörz, 1970;

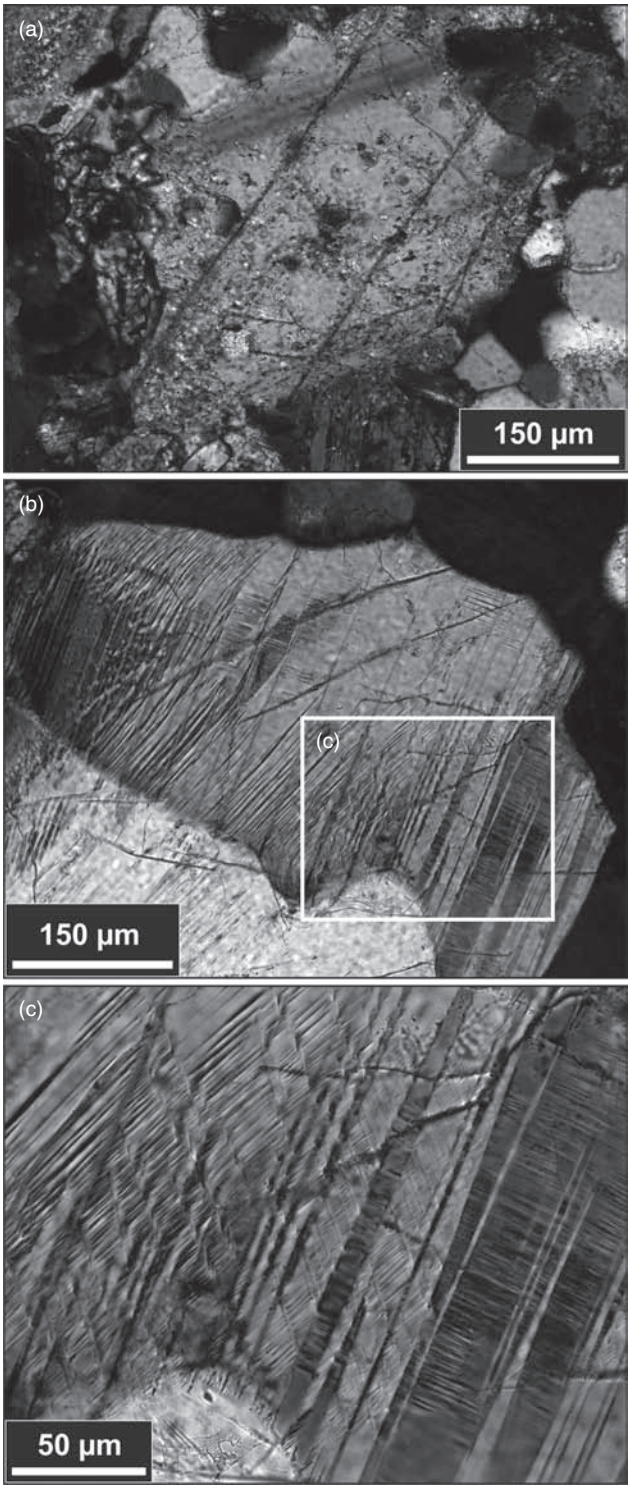

Figure 8.5 Thin-section photomicrographs (crossed polars) of shocked feldspar grains in shatter cone samples from (a) the Keurusselkä (Finland) and (b, c) the Manicouagan (Canada) impact structures. (a) Plagioclase grain with one prominent set of PFs oriented NE–SW (sample KE2; from Jylhänniemi; from Ferrière *et al.* (2010a: fig. 5a)). (b) Plagioclase grain with three sets of planar microstructures. (c) Enlarged part of (b) showing details of the three sets of planar microstructures. Two sets of closely spaced microstructures (possible PDFs?) are visible; one prominent set trending NE–SW is visible on the left part of the photograph; the other possible PDF set is perpendicular to polysynthetic twinning (visible on the right part of the photograph). The third set, most likely PFs, is barely visible on the central part of the photograph (oriented NW–SE). Directions (e.g. NE–SW) are in relation to an arbitrary 'North' at the top of the image.

see Fig. 8.7), but also in other minerals, such as graphite (Stöffler, 1972; El Goresy *et al.*, 2001a). Typically, kink bands form in sheet silicates, without specific orientation relative to the rational crystallographic planes. As kink bands are also observed in minerals from non-impact settings (such as in tectonically deformed rocks) they cannot be used as a diagnostic criterion for the impact origin of a structure. In the case of graphite, in addition to kink bands and with increasing shock pressure, narrowly spaced twin lamellae develop and a partial to complete degradation of the typical prominent bireflection and birefringence is observed (El Goresy *et al.*, 2001a).

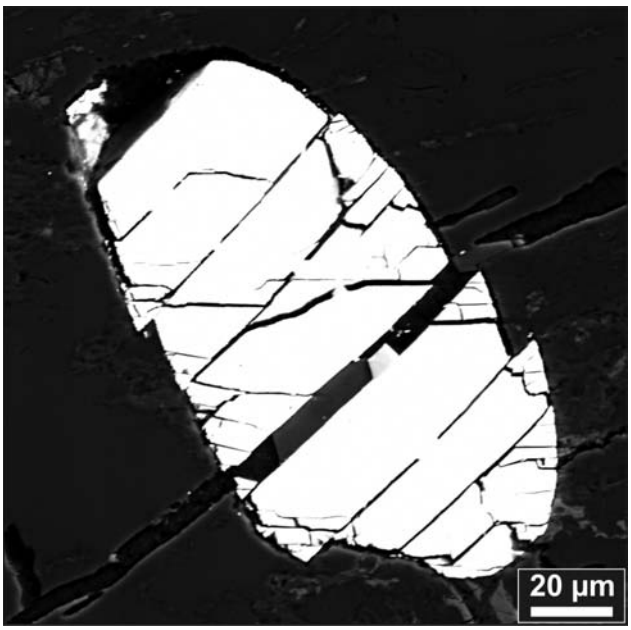

Figure 8.6 Backscattered electron image of a zircon grain with one set of planar fractures (impactite sample from the Haughton impact structure, Canada). Image kindly provided by Alaura C. Singleton (London, Canada).

8.2.3.3 Mechanical twins

Mechanical twins have been observed in a variety of minerals, including pyroxene, amphibole, titanite and ilmenite, and more rarely in plagioclase. These twins appear as sets of parallel bands, submicroscopic to some 10 µm in width (Stöffler, 1972).

Carbonates are known from a variety of impact structures, but relatively little is actually known about shock deformation of calcite and even less for dolomite. One of the few known shock effects is the development of mechanical twins (e.g. Turner *et al.*, 1954; Barber and Wenk, 1973; Robertson and Grieve, 1978; Langenhorst *et al.*, 2002). The first detailed application of twin analysis in calcite has recently been used to quantify shock pressures in calcite from the Serpent Mound impact structure, USA (Schedl, 2006). Given that low shear stresses of approximately 10 MPa are required for twinning in calcite (Schedl, 2006); this technique may be a useful shock indicator for carbonates shocked to relatively low shock levels.

8.2.4 Diaplectic glasses

Diaplectic glass forms without melting, by solid-state transformation (De Carli and Jamieson, 1959), generally from framework minerals, such as quartz or feldspar. Other minerals, such as biotite and pyroxene, tend to oxidize or to decompose without forming diaplectic glass. A fundamental diagnostic feature of diaplectic glass is that, although it is amorphous, the pre-shock morphology and texture of the mineral are preserved and flow

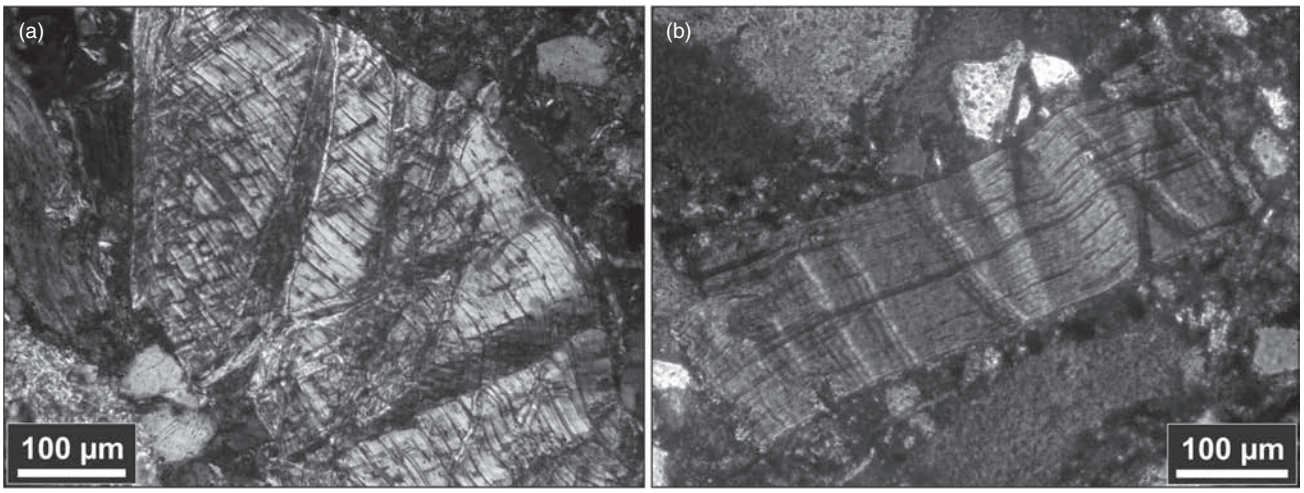

Figure 8.7 Microphotographs (crossed polars) of kink-banding in mica. (a) Kink bands in a muscovite grain; suevite sample from the Chesapeake Bay structure (USA; sample CB6-100; depth: 1427.01 m). (b) Kink bands in a biotite grain; polymict impact breccia sample from the Rochechouart structure (France). Note that kink bands in mica are not a diagnostic criterion for shock metamorphism, as they can form also in other (i.e. non-impact) settings (see text). (See Colour Plate 21)

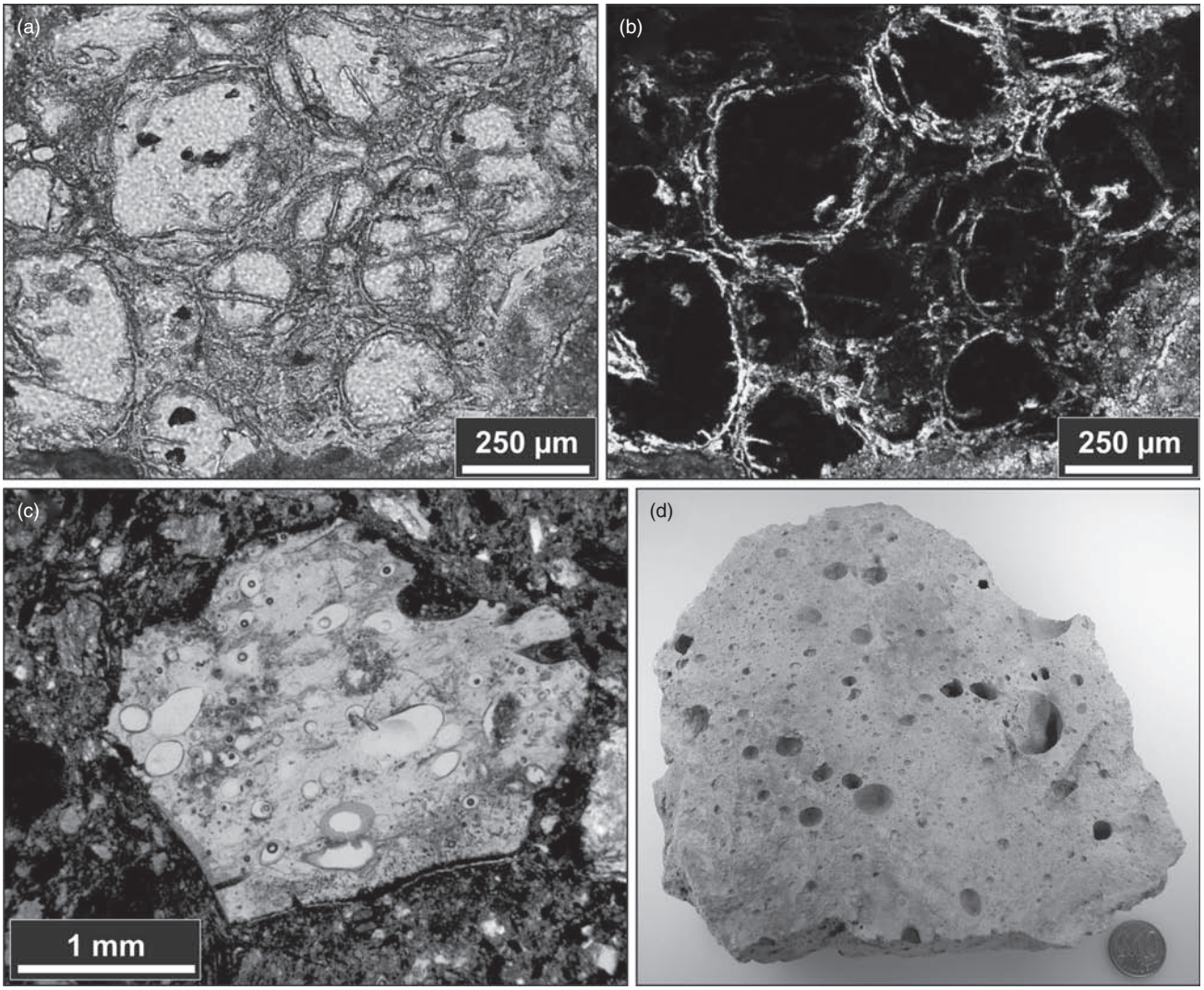

Figure 8.8 Photographs of impact metamorphism features formed by solid-state transformation (a, b) and by melting (c, d). (a) Plane-polarized light and (b) crossed-polars microphotographs of diaplectic quartz glass from a shocked sandstone clast in impact breccia from the Haughton impact structure (sample HMP-02-092). Note the occurrence within and between the idiomorphic quartz grains (i.e. now transformed to diaplectic quartz glass) of coesite and some alteration products. (c) Highly vesicular, shard-like, melt glass fragment in impactite from the Bosumtwi crater (sample LB-39a; from outside the crater rim; plane-polarized light). (d) Macrophotograph of a hand specimen of vesicular impact melt rock from the Rochechouart impact structure (France). (See Colour Plate 22)

structures or vesicles are absent (Stöffler and Langenhorst, 1994). Two types of diaplectic glasses are often reported in the literature; namely, diaplectic quartz glass and maskelynite (i.e. diaplectic plagioclase feldspar glass; Tschermak, 1872; Milton and De Carli, 1963; Bunch *et al.*, 1967). The latter has only been documented in crystalline rocks.

Diaplectic glass starts to form in the high-pressure regime, at shock pressures higher than approximately 35 GPa for quartz in the case of dense non-porous crystalline rocks (e.g. Stöffler, 1972; Stöffler and Langenhorst, 1994; French, 1998; and references therein) and at somewhat lower shock pressures, between 28 and 35 GPa, for An-rich feldspar (Stöffler *et al.*, 1986). In sandstones, diaplectic quartz glass starts to form at pressures as low as approximately 5.5 GPa (Kieffer *et al.*, 1976), and between approximately 10 and 20 GPa almost complete conversion of quartz to diaplectic glass has been observed (Osinski, 2007; Fig. 8.8a,b). Importantly, the pressure limit for complete transformation to diaplectic glass decreases with increasing pre-shock temperature (e.g. Langenhorst and Deutsch, 1994).

Typically, diaplectic glasses have a refractive index and a density that decrease with increasing shock intensity. Thermal annealing experiments have shown that, at temperatures above approximately 1200 °C, diaplectic quartz glass starts to recrystallize (e.g. Rehfeldt-Oskierski, 1986), forming ballen α-cristobalite

(see Section 8.3.2 for a discussion of ballen silica). As a result, in some impactites, diaplectic quartz glass is missing, but either ballen α-cristobalite or ballen α-quartz is present.

8.2.5 Mineral and whole-rock melt

Melting of individual minerals starts at around 50 GPa (Stöffler, 1972) and at around 60 GPa for the whole-rock in the case of non-porous crystalline rocks, while for sandstones the melting of individual quartz grains starts at pressures as low as approximately 20 GPa and whole-rock melting occurs above approximately 30–35 GPa (Kieffer *et al.*, 1976). Somewhat similar observations as the one by Kieffer *et al.* (1976) on Coconino Sandstone from Meteor Crater were reported recently by Osinski (2007) for sandstone samples from the Haughton impact structure. At both these sites it is apparent that melting is localized in original pore spaces and along grain boundaries.

Mineral and whole-rock melts (Fig. 8.8c,d) have approximately the same composition as the original minerals or mixture of minerals. Impact melt is present in many different forms and settings associated with impact structures, and the reader is referred to Chapter 9 for a detailed overview and discussion of the processes and products of impact melting. Commonly observed in impactites, lechatelierite is an SiO_2 melt that forms at very high temperatures (above 1700 °C) without necessarily requiring high shock pressures. Lechatelierite occurs in nature only in fulgurites and in impactites (e.g. Stöffler and Langenhorst, 1994 and references therein).

8.2.6 High-pressure polymorphs

High-pressure phases are commonly reported in impactites; for example, coesite and stishovite (from quartz), diamond (from graphite) and reidite (from zircon). However, coesite and diamond are not exclusively formed during shock metamorphism as they are also products of endogenic processes (Schreyer, 1995). Other high-pressure polymorphs, such as jadeite (from plagioclase), majorite (from pyroxene) and wadsleyite and ringwoodite (from olivine), are commonly reported in meteorites (e.g. Ohtani *et al.*, 2004; Fritz and Greshake, 2009; and references therein) – where they are interpreted as products of shock – but these phases have not yet been documented in impactites from terrestrial impact structures. Recently, Stähle *et al.* (2004) presented detailed observations documenting the shock-induced formation of kyanite (Al_2SiO_5) from sillimanite, arguing that it represents a third shock-induced high-pressure silicate polymorph found in the Ries impact crater, in addition to coesite and stishovite. Finally, two rutile high-pressure phases, namely TiO_2-II and akaogiite (El Goresy *et al.*, 2001b, 2010) were discovered in shocked gneisses from the Ries structure; TiO_2-II was also found in breccias from the Chesapeake Bay impact structure (Jackson *et al.*, 2006).

8.2.6.1 Coesite and stishovite

The high-pressure polymorphs of quartz, namely coesite and stishovite, form upon shock compression and are preserved as metastable phases in dense non-porous crystalline rocks that experienced peak pressure ranges between 30 and 60 GPa and between 12 and 45 GPa respectively (e.g. Stöffler and Langenhorst, 1994). Interestingly, recent investigations (Lakshtanov *et al.*, 2007) have shown that aluminium and possibly hydrogen content in these phases have a large effect on the stability fields of coesite and stishovite. Coesite was named after Loring Coes, who first produced experimentally this high-shock pressure polymorph (Coes, 1953), whereas stishovite was named after Sergei M. Stishov, who first discovered stishovite during high-pressure experiments (Stishov and Popova, 1961). Both polymorphs generally occur in impactites within diaplectic quartz glass (see Fig. 8.9), along grain boundaries or in association with PDFs (Stöffler, 1971; Kieffer *et al.*, 1976, Stähle *et al.*, 2008; and references therein).

In sedimentary rocks, coesite is known to form at pressures as low as approximately 5.5 GPa and is common at pressures above 10 GPa (Kieffer *et al.*, 1976). In sandstones from Meteor Crater and the Haughton impact structure, coesite is typically localized within so-called 'symplectic regions' that are distinguished in transmitted light by their very high relief and the presence of opaque regions. These symplectic regions represent microscopic intergrowths of quartz, coesite and diaplectic glass and form around the rims of sand grains and along fractures (Kieffer, 1971; Kieffer *et al.*, 1976; Osinski, 2007). The generation of coesite (and other shock metamorphic effects) in sandstones at pressures substantially lower than in crystalline rocks has been explained due to the effect of porosity and complex interactions during the collapse of pore spaces (Kieffer *et al.*, 1976). The porosity leads to much higher shock- and post-shock-temperatures within these rocks, especially at pores and fractures, which is preferably the cause for the onset of high-pressure phases in porous rocks at distinctly lower shock pressures. Small polycrystalline aggregates of coesite (up to approximately 20–25 μm in size), colourless to brownish, have been characterized in impactites from a few impact structures, including Meteor Crater (Chao *et al.*, 1960), the Ries (Shoemaker and Chao, 1961), Wabar (Chao *et al.*, 1961), Bosumtwi (Littler *et al.*, 1961) and Haughton (Osinski, 2007). At Meteor Crater and at the Ries, stishovite was also detected. Indeed, the identification of these two polymorphs was used for establishing the impact origin of the Ries crater (Shoemaker and Chao, 1961).

Stishovite has so far only been identified in meteorites and in impact-related rocks (e.g. see the review by Gillet *et al.* (2007)), but post-stishovite is one of the main constituents of the basaltic layer of subducting slabs and it may also occur in the lower mantle and at the core–mantle boundary (e.g. Lakshtanov *et al.*, 2007; Liu *et al.*, 2007). The presence of coesite is frequently reported in kimberlites or in ultra-high-pressure metamorphic rocks, but it can be distinguished easily from coesite in impactites by the paragenesis and by the different geological setting of occurrence. It is important to note that, under static equilibrium conditions, where reaction rates are slower and kinetic factors less important, coesite forms at pressures greater than approximately 2 GPa and stishovite starts to form around 7–8 GPa (Heaney *et al.*, 1994 and references therein).

8.2.6.2 Impact diamonds

Impact diamonds were first discovered in impactites from the Popigai structure, Siberia (Masaitis *et al.*, 1972), and were later

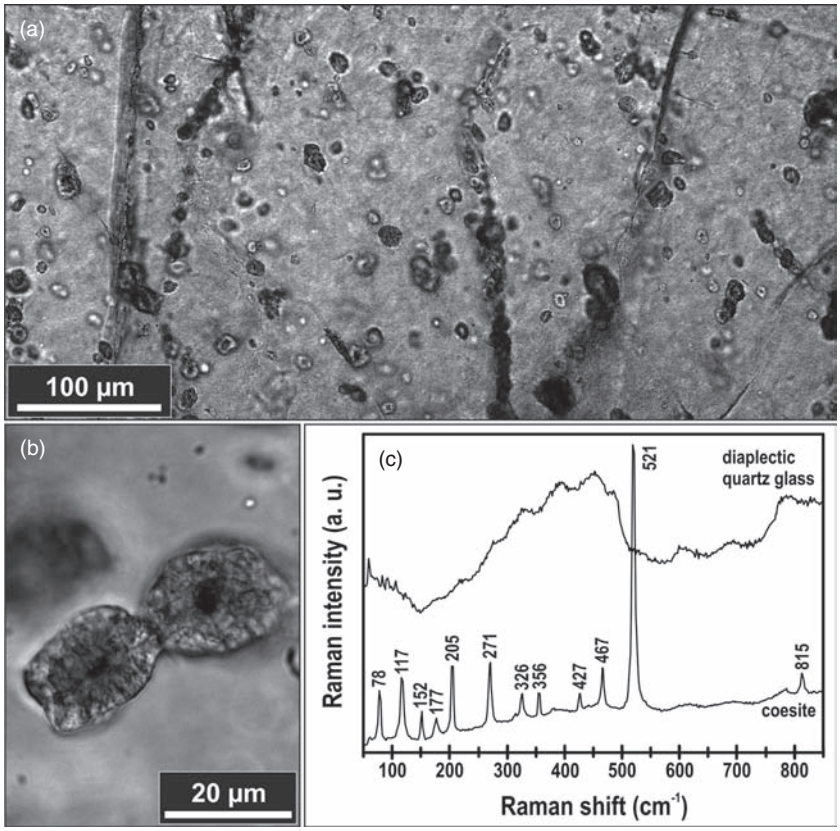

Figure 8.9 Microphotographs (a, b; in plane-polarized light) and associated micro-Raman spectra (c) of coesite in diaplectic quartz glass. (a) Coesite aggregates (green to brown clusters) within diaplectic quartz glass from suevite (Bosumtwi crater; sample LB-44B; from outside the crater rim). (b) Enlarged view of two aggregates of coesite from the same sample. (c) Typical micro-Raman spectra of one of the illustrated aggregates of coesite and of the host diaplectic quartz glass.

found at the Ries crater (Rost *et al.*, 1978) and in some other impact structures (see Vishnevsky *et al.* (1997) and El Goresy *et al.* (2001a) and references therein). Impact diamonds are usually polycrystalline and defect rich, and some features of the precursor graphite are inherited (e.g. Koeberl *et al.*, 1997; Langenhorst, 2002). Based on graphite–diamond textural relations, El Goresy *et al.* (2001a) concluded that impact diamonds are formed from graphite by shock-induced solid-state phase transformation during the very short time of the shock compression.

Schmitt *et al.* (2005) discussed two distinct formation mechanisms for impact diamond: (a) diamonds are produced by thermally activated, diffusion-controlled, mechanisms (resulting in the formation of aggregates of small diamond grains of single-crystal structure in paragenesis with shocked graphite or other high-temperature phases of carbon); (b) solid-state martensitic phase transformation of graphite or other carbon phases to diamond (resulting in the formation of diamond paramorphs after graphite or other carbon phases). El Goresy *et al.* (2001a) estimated that impact diamonds from the Ries crater are formed at a peak-shock pressure between 30 and 40 GPa; however, impact diamonds could form at lower pressure, depending on the setting and on the initial crystallinity of the graphite precursor.

Another carbon high-pressure phase with a hexagonal stacking sequence, called lonsdaleite, has been detected using X-ray diffraction techniques (Frondel and Marvin, 1967; Hanneman *et al.*, 1967); however, it is not yet clear if this phase exists as an independent mineral phase or if it is an effect of stacking faults within impact diamond crystals.

8.2.6.3 Reidite

At around 20 GPa, zircon is transformed onto the scheelite-structured high-pressure polymorph reidite. This solid-state transition of zircon to reidite is completed at 52 GPa according to Fiske *et al.* (1994). Reidite was first characterized in zircons from an upper Eocene impact ejecta layer (Glass *et al.*, 2002), and it was named after Alan F. Reid, who first produced experimentally this high-shock pressure polymorph of zircon (Reid and Ringwood, 1969). Subsequently, reidite was also identified in impactites from the Ries impact structure by Gucsik *et al.* (2004). According to Wittmann *et al.* (2006), the transformation of zircon to reidite may be hampered by impurities and pre-impact metamictization. Because reidite is refractory, surviving to a temperature up to approximately 1000 °C (Fiske *et al.*, 1994), and also resistant to alteration, it can be used as an indicator of peak shock pressure in impactites (Glass *et al.*, 2002). Based on the paragenetic relationships of reidite with ballen quartz observed in samples from the Popigai structure, Wittmann *et al.*

(2006) concluded that reidite decomposes to granular-textured zircon at temperatures greater than approximately 1100 °C.

8.3 Post-shock thermal features

In several impactites, the presence of post-shock (thermal) features, such as toasted quartz and ballen silica, are also commonly present.

8.3.1 Toasted quartz

The term 'toasted' quartz was used first by Short and Gold (1993) in an abstract, for the description of quartz grains showing an orange–brown to greyish–reddish brown colour, and in reference to the perceived aspect of a 'toasted bread' appearance (see Fig. 8.10). Toasted quartz was then studied in more detail by Short and Gold (1996) and Whitehead *et al.* (2002) and has now been documented at many terrestrial impact sites developed in both crystalline and sedimentary target rocks. Toasted quartz is considered to be a post-shock feature, either resulting from 'hydrothermal or other post-shock modification' (Short and Gold, 1996) or resulting from 'the exsolution of water from glass, primarily along PDFs, during heat-driven recrystallization' (Whitehead *et al.*, 2002). Both studies have also shown that toasted quartz has a higher albedo than untoasted quartz, visible in a hand specimen. In addition, Whitehead *et al.* (2002) noted that 'no compositional origin for the browning is evident'. The brownish appearance of quartz is, according to Whitehead *et al.* (2002), caused by a high proportion of very tiny fluid inclusions that are principally located along decorated PDFs. More recently, shocked quartz with toasted appearance has also been reported from the late Eocene impact ejecta layer at Massignano, Italy (Glass *et al.*, 2004) and within an impact ejecta layer containing Australasian microtektites (Glass and Koeberl, 2006). In a recent abstract, Ferrière *et al.* (2009b) confirm that toasted quartz contains a high proportion of small vesicles and that these vesicles probably enhance scattering of transmitted light. However, Ferrière *et al.* (2009b) note that the vesicles observed in toasted quartz, with highly variable sizes, are likely not related to recrystallization of PDF glass, as previously suggested by Whitehead *et al.* (2002). Using electron microprobe techniques, Ferrière *et al.* (2009b) showed that some trace elements (and inclusions), originally present in quartz, were removed from the quartz structure. They suggest that toasted quartz is formed by vesiculation after pressure release, at high

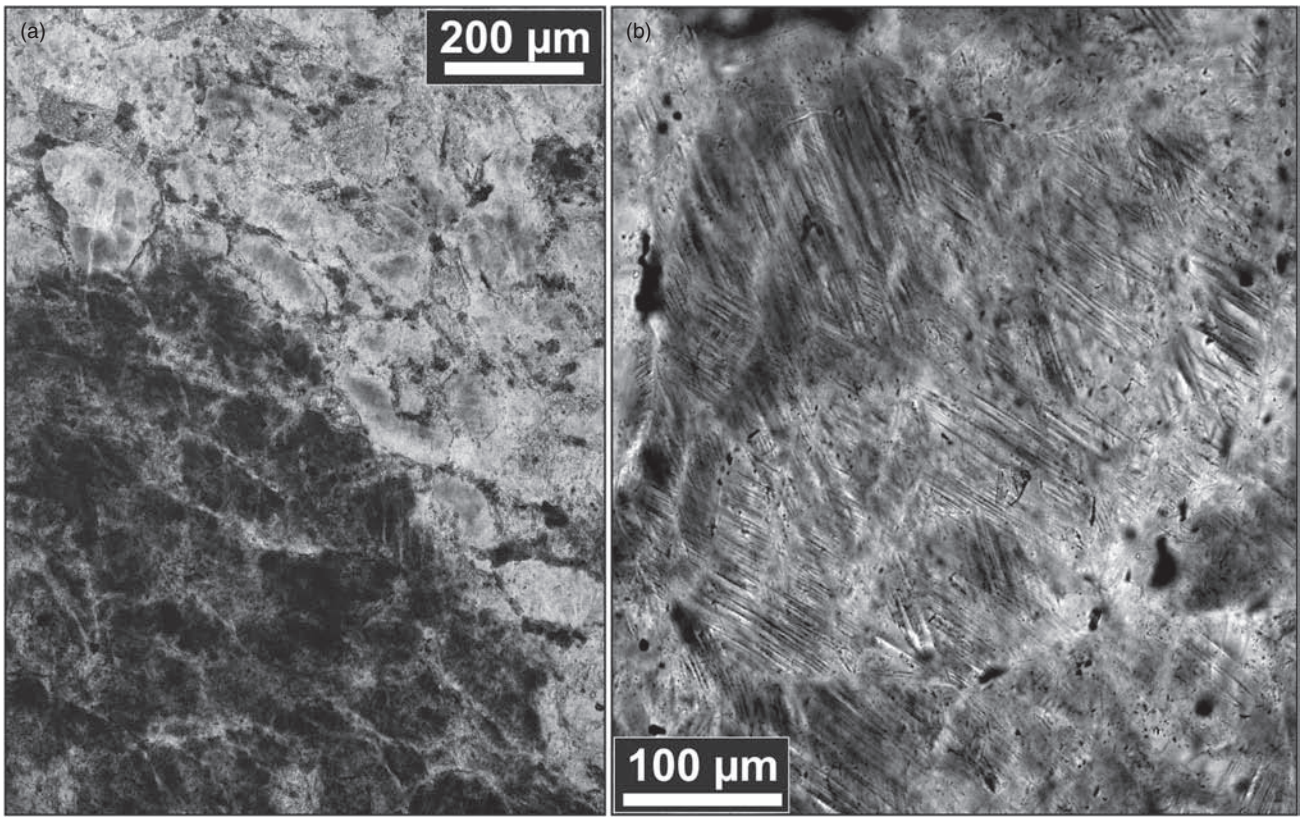

Figure 8.10 Microphotographs (in plane-polarized light) of quartz grains with toasted appearance. (a) Detail of a toasted region (lower left side of the photograph) and untoasted region of a sandstone clast; conglomerate sample from the Chesapeake Bay structure (USA; sample CB6-125b; depth: 1522.72 m). (b) Toasted quartz grain with two PDF sets; quartzite clast in suevite from the Bosumtwi crater (sample KR8-006; depth: 240.36 m).

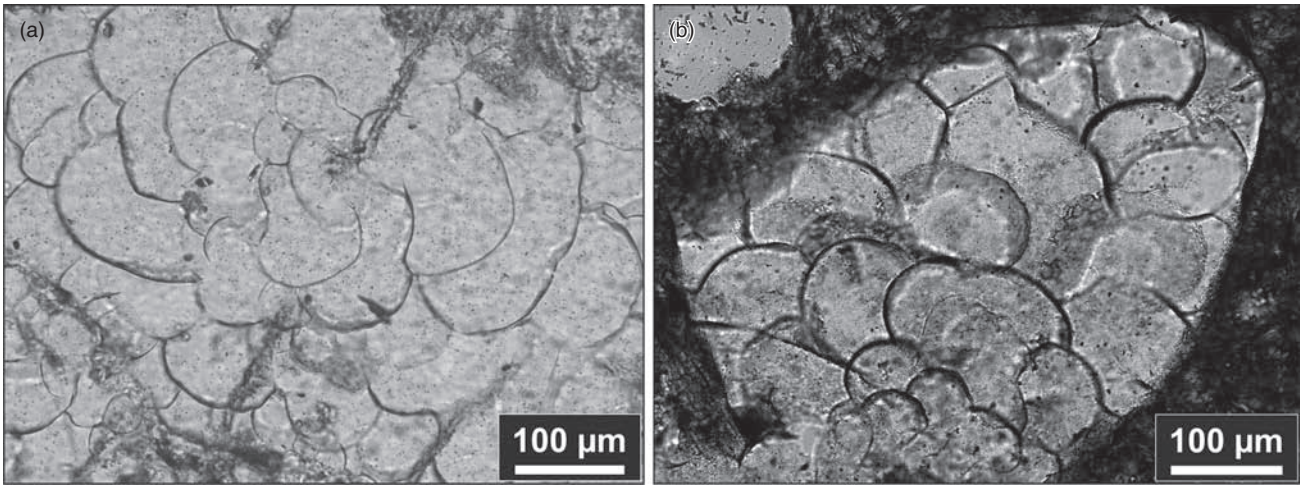

Figure 8.11 Microphotographs (in plane-polarized light) of ballen silica. (a) Elongate ovoid (crescent) to roundish α-cristobalite ballen in suevite from the Bosumtwi crater (sample BH1-0790; from Ferrière *et al.* (2009c: fig. 1a). (b) Ballen α-quartz with varied ballen sizes (smaller ones on the lower part of the photograph) in a glassy impact melt rock from the Wanapitei structure (Canada; from Ferrière *et al.* (2010b: fig. 1a).

post-shock temperatures and, thus, represents the beginning of quartz breakdown due to heating.

8.3.2 Ballen quartz and cristobalite

Ballen silica, with either an α-quartz or α-cristobalite structure, occurs in impactites as independent clasts or within diaplectic quartz glass or lechatelierite inclusions (see Fig. 8.11). Ballen are more or less spheroidal, in some cases elongate (ovoid), bodies that range in size from approximately 8 to 215 µm; they can intersect or penetrate each other or abut each other (Ferrière *et al.*, 2009c). The occurrence of so-called ballen quartz has been reported from approximately one-fifth of the known terrestrial impact structures, mostly from clasts in impact melt rock and, more rarely, in suevite (e.g. from the Chicxulub crater; Ferrière *et al.*, 2009c). Different types of ballen silica have been described and characterized (Carstens, 1975; Bischoff and Stöffler, 1984; Ferrière *et al.*, 2009c). Several mechanisms have been proposed for the formation of ballen. Most recently, it has been suggested that the formation of ballen can occur via two processes (Ferrière *et al.*, 2009c): (1) impact-triggered solid–solid transition from α-quartz to diaplectic quartz glass, followed by the formation (at high temperature) of ballen of β-cristobalite and/or β-quartz, and finally back-transformation to α-cristobalite and/or α-quartz; or (2) a solid–liquid transition from quartz to lechatelierite followed by nucleation and crystal growth at high temperature. Given that ballen quartz/cristobalite results from back-transformations from shock-induced states, ballen silica cannot be considered direct evidence of shock metamorphism, but represents indirect evidence. For more information, the reader is invited to read the recent detailed observations and review on ballen silica by Ferrière *et al.* (2009c).

It is notable that ballen silica has only been documented in impactites at impact structures developed in entirely crystalline target rocks (e.g. Carstens, 1975; Grieve, 1975; Bischoff and Stöffler, 1984; Ferrière *et al.*, 2009c) or in mixed crystalline–sedimentary targets but where the crystalline rocks comprise the bulk of the shocked and shock-melted lithologies (e.g. the Ries crater; von Engelhardt, 1972). Indeed, at the Ries crater, ballen silica only occurs in coherent impact melt rocks derived entirely from the crystalline basement (Osinski, 2004), but is not present in the suevite deposits (Osinski *et al.*, 2004), which are derived from a combination of sedimentary and crystalline target rocks. At impact structures formed predominantly in sedimentary rocks, such as the Haughton structure, ballen silica has not been documented to date (Osinski, 2007). This is consistent with the models for ballen silica formation (Ferrière *et al.*, 2009c), which involve high-temperature transitions; sediment-derived impact melts are typically rapidly quenched compared with melts generated from crystalline targets, such that cooling rate may be an important constraint in the formation of ballen silica (Osinski, 2007).

8.4 Concluding remarks

After decades of intense investigation of tectosilicates, mostly quartz in terrestrial impactites, the time has arrived to have a closer look at shock effects in nesosilicates, inosilicates and phyllosilicates, and also at carbonate and sulfate minerals in respect to being able to identify impact structures formed in non-silicate target rocks, and also to better evaluate consequences of impact in such types of rock-forming minerals on the environment and atmosphere. We encourage both, field and experimental studies, as we cannot have one without the other if we want to associate proper numbers with observations, even we should be aware that nature is hard to model. Such a type of research is not only needed to expand our knowledge on impact effects on Earth, but also to better understand the formation and evolution of our Solar System.

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