Impact cratering: processes and products

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1.1 Introduction

Over the past couple of decades, it is has become widely recognized that impact cratering is a ubiquitous geological process that affects all planetary objects with a solid surface. Indeed, meteorite impact structures are one of the most common geological landforms on all the rocky terrestrial planets, except Earth, and many of the rocky and icy moons of Saturn and Jupiter. A unique result of the impact cratering process is that material from depth is brought to the surface in the form of ejecta deposits and central uplifts. Impact craters, therefore, provide unique windows into the subsurface on planetary bodies where drilling more than a few metres is not a viable scenario for the foreseeable future.

On many planetary bodies where planetary-scale regoliths can develop through micrometeorite bombardment, aeolian or cryogenic processes, the crater walls of fresh impact craters also provide unique sites where in situ outcrops can be found. It should not be surprising, therefore, that impact craters have been, and remain, high-priority targets for planetary exploration missions to the Moon, Mars and elsewhere.

The impact record on Earth remains invaluable for our understanding of impact processes, for it is the only source of ground-truth data on the three-dimensional structural and lithological character of impact craters. However, Earth suffers from active erosion, volcanic resurfacing and tectonic activity, which continually erase impact structures from the rock record. Despite this, 181 confirmed impact structures have been documented to date, with several more 'new' impact sites being recognized each year (Earth Impact Database, 2012). Although we lack ground truth, apart from a few lunar and Martian sites visited by human and robotic explorers, the results of planetary exploration missions continue to provide a wealth of new high-resolution data about the surface expression of impact craters. The driving paradigm is that impact cratering is governed by physics and the fundamental processes are the same regardless of the planetary target (Melosh, 1989). However, variations in planetary conditions permit the investigation of how different properties lead to slightly different end results. The Moon represents an end-member case with respect to the terrestrial planets. Low planetary gravity and lack of atmosphere result in cratering efficiency, for a given impact, that is higher than on the other terrestrial planets (Stöfler et al., 2006). The relatively simple target geology combined with the lack of post-impact modification by aqueous and aeolian processes makes the Moon an ideal natural laboratory for studying crater morphology and morphometry. Mercury is similar to the Moon, except for a higher impact velocity, and new data from the MESSENGER spacecraft (see Solomon et al. (2011) and references therein) are providing a wealth of new information on the mercurian impact cratering record (Strom et al., 2008). Venus is almost the antithesis of the Moon and Mercury. The relatively high planetary gravity and thick atmosphere reduce cratering efficiency for a given impact relative to these bodies (Schultz, 1993). Hotter surface and subsurface temperatures affect numerous aspects of the cratering process on Venus, the most spectacular outcome of which is the production of vast impact melt flows (Grieve and Cintala, 1995). The final terrestrial planet, Mars, has a thinner atmosphere, a more complex geology, including the presence of volatiles, and more endogenic geological processes to modify craters (Carr, 2006). It is more Earth-like in this respect, which comes with the associated complications, but its impact cratering record is vastly better preserved and exposed than on Earth (Strom et al., 1992).

Notwithstanding the prior discussion of the ubiquity of impact craters throughout the Solar System, it is important to recognize that, despite being first observed on the Moon by Galileo Galilei in 1609, it was not until the 1960s and 1970s that the importance of impact cratering as a geological process began to be recognized. In 1893, the American geologist Grove Gilbert proposed an impact origin for these lunar craters, but it was not until the 1900s that the first impact crater was recognized on Earth: Meteor or Barringer Crater in Arizona (Barringer,
Barringer Crater. It was not until the recognition of shock metamorphic criteria (French and Short, 1968; see Chapter 8), which resulted in the increased recognition of terrestrial impact sites, together with the impetus provided by the Apollo landings on the Moon in the 1960s and 1970s, that increasing awareness and a

Figure 1.1 Simple impact craters. (a) Panoramic image of the 1.2 km diameter Meteor or Barringer Crater, Arizona. (b) Schematic cross-section through a simple terrestrial impact crater. Fresh examples display an overturned flap of near-surface target rocks overlain by ejecta. The bowl-shaped cavity is partially filled with allochthonous unshocked and shocked target material. (c) A 2 m per pixel true-colour image of Barringer Crater taken by WorldView2 (north is up). Image courtesy of Livio L. Tornabene and John Grant. (d) Portion of Lunar Reconnaissance Orbiter Camera (LROC) image M122129845 of the 2.2 km diameter Linné Crater on the Moon (NASA/GSFC/Arizona State University). (See Colour Plate 1)
Discussion of the importance of meteorite impacts for Earth evolution finally entered the geological mainstream in 1980, with evidence for a major impact as the cause of the mass extinction event at the Cretaceous–Paleogene (K–Pg) boundary 65 Ma ago (Alvarez et al., 1980). The actual impact site, the approximately 180 km diameter Chicxulub crater, was subsequently identified in 1991, buried beneath approximately 1 km of sediments in the Yucatan Peninsula, Mexico (Hildebrand et al., 1991). The spectacular impact of comet Shoemaker–Levy 9 into Jupiter in July 1994 reminded us that impact cratering is a process that continues to the present day. The result is that it is now apparent that meteorite impact events have played an important role throughout Earth’s history, shaping the geological landscape, affecting the evolution of life and producing economic benefits. As summarized in Chapter 2, the evolutions of the terrestrial planets and the Earth’s moon have been strongly affected by changes in the population of impactors and in the impact cratering rate through time in the inner Solar System.

To summarize, our understanding of the impact cratering process has come a long way in the past century, but several fundamental aspects of the processes and products of crater formation remain poorly understood. One of the major reasons for this is that, unlike many other geological processes, there have been no historical examples of hypervelocity impact events (French, 1998). This is, of course, fortunate, as impacts release energies far in excess of even the most devastating endogenous geological events (Fig. 1.2). Our understanding is also hindered by the major differences between impact events and other geological processes, including (1) the extreme physical conditions (Fig. 1.3), (2) the concentrated nature of the energy release at a single point on the Earth’s surface, (3) the virtually instantaneous nature of the impact process (e.g. seconds to minutes) and (4) the strain rates involved (∼10⁴–10⁶ s⁻¹ for impacts versus 10⁻³–10⁻⁶ s⁻¹ for endogenous tectonic and metamorphic processes) (French, 1998). Impact events, therefore, are unlike any other geological process, and the goal of this chapter, and this book, is to provide a modern up-to-date synthesis as to our current understanding of the processes and products of impact cratering.

1.2 Formation of hypervelocity impact craters

The formation of hypervelocity¹ impact craters has been divided, somewhat arbitrarily, into three main stages (Gault et al., 1968): (1) contact and compression; (2) excavation; and (3) modification (Fig. 1.4). These are described below. A further stage of ‘hydrothermal and chemical alteration’ has also sometimes been

¹Hypervelocity impact occurs when a cosmic projectile is large enough (typically >50 m for a stony object and >20 m for an iron body) to pass through the atmosphere with little or no deceleration and so strike at virtually its original cosmic velocity (>11 km s⁻¹; French, 1998). This produces high-pressure shock waves in the target. Smaller projectiles lose most of their original kinetic energy in the atmosphere and produce small metre-size ‘penetration craters’, without the production of shock waves.
included as a separate, final stage in the cratering process (Kieffer and Simonds, 1980), and is also described below.

1.2.1 Contact and compression

The first stage of an impact event begins when the projectile, be it an asteroid or comet, contacts the surface of the target (Fig. 1.4) – see Chapter 3 for details. Modelling of the impact process suggests that the projectile penetrates no more than one to two times its diameter (Kieffer and Simonds, 1980; O’Keefe and Ahrens, 1982). The pressures at the point of impact are typically several thousand times the Earth’s normal atmospheric pressure (i.e. >100 GPa) (Shoemaker, 1960). The intense kinetic energy of the projectile is transferred into the target in the form of shock waves that occur at the boundary between the compressed and uncompressed target material (Melosh, 1989). These shock waves, traveling faster than the speed of sound, propagate both into the target sequence and back into the projectile itself. When this reflected shock wave reaches the ‘free’ upper surface of the projectile, it is reflected back into the projectile as a rarefaction or tensional wave (Ahrens and O’Keefe, 1972). The passage of this rarefaction wave through the projectile causes it to unload from high shock pressures, resulting in the complete melting and/or vaporization of the projectile itself (Gault et al., 1968; Melosh, 1989). The increase in internal energy accompanying shock compression and subsequent rarefaction also results in the shock metamorphism (see Chapter 8), melting (see Chapter 9) and/or vaporization of a volume of target material close to the point of impact (Ahrens and O’Keefe, 1972; Grieve et al., 1977). The point at which the projectile is completely unloaded is generally taken as the end of the contact and compression stage (Melosh, 1989; Chapter 3).

1.2.2 Excavation stage

The transition from the initial contact and compression stage into the excavation stage is a continuum. It is during this stage that the actual impact crater is opened up by complex interactions between the expanding shock wave and the original ground surface (Melosh, 1989) – see Chapter 4 for details. The projectile itself plays no role in the excavation of the crater, having been unloaded, melted and/or vaporized during the initial contact and compression stage.

During the excavation stage, the roughly hemispherical shock wave propagates out into the target sequence (Fig. 1.4). The centre of this hemisphere will be at some depth in the target sequence (essentially the depth of penetration of the projectile). The passage of the shock wave causes the target material to be set in motion, with an initial outward radial trajectory. At the same time, shock waves that initially travelled upwards intersect the ground surface and generate rarefaction waves that propagate back downwards into the target sequence (Melosh, 1989). In the near-surface region an ‘interference zone’ is formed in which the maximum recorded pressure is reduced due to interference between the rarefaction and shock waves (Melosh, 1989).

The combination of the outward-directed shock waves and the downward-directed rarefaction waves produces an ‘excavation flow-field’ and generates a so-called ‘transient cavity’ (Fig. 1.4 and Fig. 1.5) (Dence, 1968; Grieve and Cintala, 1981). The different trajectories of material in different regions of the excavation flow field result in the partitioning of the transient cavity into an upper ‘excavated zone’ and a lower ‘displaced zone’ (Fig. 1.5). Material in the upper zone is ejected ballistically beyond the transient cavity rim to form the continuous ejecta blanket (Oberbeck, 1975) – see Chapter 4. Experiments and theoretical
Figure 1.4 Series of schematic cross-sections depicting the three main stages in the formation of impact craters. This multi-stage model accounts for melt emplacement in both simple (left panel) and complex craters (right panel). For the modification stage section, the arrows represent different time steps, labelled 'a' to 'c'. Initially, the gravitational collapse of crater walls and central uplift (a) results in generally inwards movement of material. Later, melt and clasts flow off the central uplift (b). Then, there is continued movement of melt and clasts outwards once crater wall collapse has largely ceased (c). Modified from Osinski et al. (2011). (See Colour Plate 2)
Ejecta deposits represent one of the most distinctive features of impact craters on planetary bodies (other than Earth), where they tend to be preserved. It is notable that the continuous ejecta deposits vary considerably in terms of morphology on different planetary bodies. For example, on Mars, many ejecta blankets have a fluidized appearance that has been ascribed due to the effect of volatiles in the subsurface (Barlow, 2005). Indeed, the volatile content and cohesiveness of the uppermost target rocks will significantly affect the runout distance of the ballistically emplaced continuous ejecta blanket, with impact angle also influencing the overall geometry of the deposits (e.g. the production of the characteristic butterfly pattern seen in very oblique impacts) (Osinski et al., 2011). In terms of the depth of excavation $d_e$, few craters on Earth preserve ejecta deposits and/or have the distinct pre-impact stratigraphy necessary for determining depth of materials. Based on stratigraphic considerations, $d_e$ at Barringer Crater is greater than 0.08$D$, where $D$ is the final rim diameter. For complex craters and basins, the depth and diameter must be referred back to the ‘unmodified’ transient cavity to reliably estimate the depth of excavation (Melosh, 1989). The maximum $d_e$ of material in the ballistic ejecta deposits of the Haughton and Ries structures, the only terrestrial complex structures where reliable data are available, yield identical values of 0.035$D_e$, where $D_e$ is the apparent crater diameter (Osinski et al., 2011). If the initial final rim diameter $D$ is used, which is the parameter measured in planetary craters, a value 0.05$D$ is obtained for Haughton.

Based on experiments, it was generally assumed that material in the displaced zone remains within the transient cavity (Stöffler et al., 1975); however, observations from impact craters on all the terrestrial planets suggest that some of the melt-rich material from this displaced zone is transported outside the transient cavity rim during a second episode of ejecta emplacement (Osinski et al., 2011). This emplacement of more melt-rich, ground-hugging flows – the ‘surface melt flow’ phase – occurs during the terminal stages of crater excavation and the modification stage of crater formation (see Chapter 4). Ejecta deposited during the surface melt flow stage are influenced by several factors, most importantly planetary gravity, surface temperature and the physical properties of the target rocks. Topography and angle of impact play important roles in determining the final distribution of surface melt flow ejecta deposits, with respect to the source crater (Osinski et al., 2011). A critical consideration is that the upper layer of ejecta reflects the composition and depth of the displaced zone of the transient cavity (Fig. 1.4). At Haughton, this value is a minimum of 0.08$D_e$ or 0.12$D$.

A portion of the melt and rock debris that originates beneath the point of impact remains in the transient cavity (Grieve et al., 1977). This material is also deflected upwards and outwards parallel to the base of the cavity, but must travel further and possesses less energy, so that ejection is not possible. This material forms the crater-fill impactites within impact craters (see Chapter 7 for an overview of impactites). Eventually, a point is reached at which the motions associated with the passages of the shock and rarefaction waves can no longer excavate or displace target rock and melt (French, 1998). At the end of the excavation stage, a mixture of melt and rock debris forms a lining to the transient cavity.
Figure 1.6  Complex impact craters. (a) Landsat 7 image of the 23 km (apparent) diameter Haughton impact structure, Devon Island, Canada. (b) Portion of Apollo 17 metric image AS17-M-2923 showing the 27 km diameter Euler Crater on the Moon. Note the well-developed central peak. (c) Thermal Emission Imaging System (THEMIS) visible mosaic of the 29 km diameter Tooting Crater on Mars (NASA). Note the well-developed central peak and layered ejecta blanket. Scale bars for (a) to (c) are 10 km. (d) Schematic cross-section showing the principal features of a complex impact crater. Note the structurally complicated rim, a down-faulted annular trough and a structurally uplifted central area (SU). (e) Schematic cross-section showing an eroded version of the fresh complex crater in (d). Note that, in this case, only the apparent crater diameter can typically be defined. (See Colour Plate 3)
1.2.3 Modification stage

The effects of the modification stage are governed by the size of the transient cavity and the properties of the target rock lithologies (Melosh and Ivanov, 1999) – see Chapter 5 for an overview. For crater diameters less than 2–4 km on Earth, the transient cavity undergoes only minor modification, resulting in the formation of a simple bowl-shaped crater (Fig. 1.1). However, above a certain size threshold the transient cavity is unstable and undergoes modification by gravitational forces, producing a so-called complex impact crater (Fig. 1.4 and Fig. 1.6; Dence, 1965) – see Chapter 5. Uplift of the transient crater floor occurs, leading to the development of a central uplift (Fig. 1.4 and Fig. 1.6). This results in an inward and upward movement of material within the transient cavity. Subsequently, the initially steep walls of the transient crater collapse under gravitational forces (Fig. 1.4). This induces an inward and downward movement of large (~100 m to kilometre-scale) fault-bounded blocks. The diameter at which the transition occurs from simple to complex craters on Earth occurs at approximately 2 km for craters developed in sedimentary targets and approximately 4 km for those in crystalline lithologies. This transition diameter is dependent on the strength of the gravitational field of the parent body and increases with decreasing acceleration of gravity (Melosh, 1989). Thus, the transition from simple to complex craters occurs at approximately 5–10 km on Mars and at approximately 15–27 km on the Moon (Pike, 1980).

It is generally considered that the modification stage commences after the crater has been fully excavated (Melosh and Ivanov, 1999). However, numerical models suggest that the maximum depth of the transient cavity is attained before the maximum diameter is reached (e.g. Kenkmann and Ivanov, 2000). Thus, uplift of the crater floor may commence before the maximum diameter has been reached. As French (1998) notes, the modification stage has no clearly marked end. Processes that are intimately related to complex crater formation, such as the uplift of the crater floor and collapse of the walls (Chapter 5), merge into more familiar endogenous geological processes, such as mass movement, erosion and so on.

1.2.4 Post-impact hydrothermal activity

Evidence for impact-generated hydrothermal systems has been recognized at over 70 impact craters on Earth (Naumov, 2005; Osinski et al., 2012), from the approximately 1.8 km diameter Lonar Lake structure, India (Hagerty and Newsom, 2003), to the approximately 250 km diameter Sudbury structure, Canada (Ames and Farrow, 2007). Based on these data, it seems highly probable that any hypervelocity impact capable of forming a complex crater will generate a hydrothermal system, as long as sufficient H2O is present (see Chapter 6 for an overview). Thus, the recognition of impact-associated hydrothermal deposits is important in understanding the evolution of impact craters through time. There are three main potential sources of heat for creating impact-generated hydrothermal systems (Osinski et al., 2005a): (a) impact melt rocks and impact melt-bearing breccias; (b) physically elevated geothermal gradients in central uplifts; and (c) thermal energy deposited in central uplifts due to the passage of the shock wave. Interaction of these hot rocks with groundwater and surface water can lead to the development of a hydrothermal system. The circulation of hydrothermal fluids through impact craters can lead to substantial alteration and mineralization. It has been shown that there are six main locations within and around an impact crater where impact-generated hydrothermal deposits can form (Fig. 1.7): (1) crater-fill impact melt rocks and melt-bearing breccias; (2) interior of central uplifts; (3) outer margin of central uplifts; (4) impact ejecta deposits; (5) crater rim region; and (6) post-impact crater lake sediments.

1.3 Morphology and morphometry of impact craters

1.3.1 Simple craters

Impact craters are subdivided into two main groups based on morphology: simple and complex. Simple craters comprise a bowl-shaped depression (Fig. 1.1). When fresh, they possess an uplifted rim and are filled with an allochthonous breccia lens that comprises largely unshocked target material, possibly mixed with impact melt-bearing lithologies (Fig. 1.1b; Shoemaker, 1960). The overall low shock level of material in the breccia lens suggests that it formed due to slumping of the transient cavity walls, and is not ‘fallback’ material (Grieve and Cintala, 1981). Simple craters typically have depth-to-diameter ratios of approximately 1:5 to 1:7 (Melosh, 1989). It is important, however, to make the distinction between the true and apparent crater (Fig. 1.1b). Morphometric data from eight simple impact structures (i.e. Barringer, Brent, Lonar, West Hawk, Aouelloul, Tenoumer, Mauritania and Wolfe Creek) define the empirical relationships: \( d_a = D^{1.06} \) and \( d_a = 0.28D^{1.02} \), where \( d_a \) is the depth of the apparent crater, \( d_t \) is the depth of the true crater and \( D \) is the rim diameter of the structure (Grieve and Pilkington, 1996). As diameter increases, so-called ‘transitional craters’ form. Such craters have not been recognized on Earth, but on the Moon and Mars, where they are abundant, spacecraft observations show that, while they lack a central peak, they possess some of the other characteristics of complex craters (see below), such as a shallower profile and terraced crater rim. As such, they are neither simple nor complex and the exact mechanism(s) responsible for their appearance remain poorly understood.

1.3.2 Complex craters

Observations of lunar craters first revealed that, as diameter increases yet further, a topographic high forms in the centre of a transitional crater, signifying the progression to a so-called complex impact crater. Such craters generally have a structurally complicated rim, a down-faulted annular trough and an uplifted central area (Fig. 1.6). These features form as a result of gravitational adjustments of the initial crater during the modification stage of impact crater formation (Chapter 5). Owing to these late-stage adjustments, complex impact craters are shallower than simple craters, with depth-to-diameter ratios of
Figure 1.7  Distribution of hydrothermal deposits within and around a typical complex impact crater. Modified from Osinski et al. (2012). (See Colour Plate 4)
terrestrial complex structures (e.g. Haughton (Fig. 1.6a), Canada; Ries, Germany; Zhamanshin, Kazakhstan) lack an emergent central peak (Grieve and Therriault, 2004). These structures are in mixed targets of sediments overlying crystalline basement, with the lack of a peak most likely due to target strength effects. This highlights the problems with making direct comparisons between impact craters on Earth and those on other planetary bodies.

Based on observations of 24 impact craters developed in sedimentary rocks on Earth, the structural uplift of the target rocks in the centre of the crater (Fig. 1.6d) was determined to be $0.086D^{1.03}$, where $D$ is the crater ‘diameter’ (Grieve and Pilkington, 1996). According to this estimate, a good working hypothesis is that the observed structural uplift is approximately one-tenth of the rim diameter at terrestrial complex impact structures. It is important to note that no data exist on the amount of structural uplift in craters developed in crystalline targets for the obvious reason that stratigraphic markers, upon which this calculation relies upon, are lacking. Despite its widespread application, there is also currently no data to support the hypothesis that this formula for structural uplift holds for craters on other planetary bodies, at least in its current form.

Figure 1.8 Series of images of lunar craters depicting the change in crater morphology with increasing crater size. (a) The 27 km diameter Euler Crater possesses a well-developed central peak. Portion of Apollo 17 metric image AS17-M-2923 (NASA). (b) The 165 km diameter Compton Crater is one of the rare class of central-peak basin craters on the Moon. Clementine mosaic from USGS Map-A-Planet. (c) Clementine mosaic of the 320 km diameter Schrödinger impact crater, which displays a peak ring basin morphology (NASA). (d) The approximately 950 km diameter Orientale Basin is the youngest multi-ring basin on the Moon (NASA/GSFC/Arizona State University).

approximately 1:10 to 1:20 (Melosh, 1989). The so-called annular trough in complex craters is filled with a variety of impact-generated lithologies (impactites) that will be introduced in Section 1.4.

A unique result of complex crater formation is that material from depth is brought to the surface. As noted above, for many impact sites, these ‘central uplifts’ provide the only samples of the deep subsurface. This is particularly important on other planetary bodies, but even on Earth they provide vital clues as to the structure of the crust. For example, the central uplift of the approximately 250 km diameter Vredefort impact structure, South Africa, provides a unique profile down to the lower crust (Tredoux et al., 1999). On the Moon and other planets, where post-impact modification of craters is generally minimal, there is a progression with increasing crater size from central peak, central-peak basin (i.e. a fragmentary ring of peaks surrounding a central peak), to peak-ring basins (i.e. a well-developed ring of peaks but no central peak) (Fig. 1.8; Stöffler et al., 2006). On Earth, erosion has modified the surface morphology of all impact craters and it is, therefore, typically not possible to ascertain the original morphology. As such, the term central uplift is preferred. Related to this is the fact that a number of relatively young (i.e. only slightly eroded) terrestrial complex structures (e.g. Haughton (Fig. 1.6a), Canada; Ries, Germany; Zhamanshin, Kazakhstan) lack an emergent central peak (Grieve and Therriault, 2004). These structures are in mixed targets of sediments overlying crystalline basement, with the lack of a peak most likely due to target strength effects. This highlights the problems with making direct comparisons between impact craters on Earth and those on other planetary bodies.
A key descriptor for complex craters is ‘diameter’. As noted above, defining the size, or diameter, of a crater is critical for estimating stratigraphic uplift, in addition to energy scaling and numerical modelling of the cratering process. Unfortunately, there is considerable confusion about crater sizes within the literature. This arises largely from the fact that most craters on Earth are eroded to some degree, whereas most craters on other planetary bodies are relatively well preserved. For a discussion of what crater diameter represents, the reader is referred to Turtle et al. (2005) and the nomenclature recommended here comes from this synthesis paper. In short, the rim (or final crater) diameter is defined as the diameter of the topographic rim that rises above the surface for simple craters, or above the outermost slump block not concealed by ejecta for complex craters (Fig. 1.6d). This is relatively easy to measure on most planetary bodies, where the topographic rim is usually preserved due to low rates of erosion (e.g. Fig. 1.6b,c). On Earth, however, such pristine craters are rare and the rim region is typically eroded away (e.g. Fig. 1.6a). The apparent crater diameter, in contrast, is defined as the diameter of the outermost ring of (semi-) continuous concentric normal faults (Fig. 1.6e). For the majority of impact structures on Earth this will be the only measurable diameter. It is not always clear how the apparent diameter is related to the rim diameter, although one would expect the rim diameter to be smaller than the apparent crater diameter. This is consistent with observations at the Haughton impact structure, where an apparent crater diameter of 23 km and a rim diameter of 16 km have been reported (Osinski and Spray, 2005).

Returning to the previous discussion on stratigraphic uplift and its application to planets other than Earth, $D$ in Grieve and Pilkington’s (1996) formula actually is predominantly based on apparent crater diameter estimates (R. A. F. Grieve, personal communication, 2012) and not rim diameter estimates, further complicating the discussion about its application to other planets.

1.3.3 Multi-ring basins

The largest impact ‘craters’ in the Solar System are typically surrounded by one or more concentric scarps or fractures and are known as multi-ring basins (Fig. 1.8d). Multi-ring basins are best studied on the Moon and Callisto, where a large number exist, although these structures remain the least understood crater morphology. There are two basic morphological types (e.g. Melosh and McKinnon, 1978). The first type, as exemplified by the Orientale basin on the Moon, exhibits a few to several inward-facing scarps with gentle outward slopes. The second type exhibit tens to hundreds of closely spaced rings consisting of a graben or outward-facing scarps surrounding a central, flat basin (e.g. Valhalla, Callisto). An important observation is that very few multi-ring basins have been documented on Ganymede, despite the obvious similarities with Callisto, and there is no clear evidence for their existence on Mercury, Mars or Venus (Melosh, 1989). In this respect, it is critical to understand that just because an impact crater is very ‘large’ (e.g. Hellas, Mars; South Pole-Aitken, Moon), this does not necessarily mean that a structure is a multi-ring basin; to be categorized as such, multiple rings must be clearly observable. It is also important to note that the rings that define multi-ring craters are distinct from the peak rings described in the previous section. In particular, it is thought that rings characteristic of multi-ring craters form outside the final crater. Several mechanisms have been proposed to account for the formation of multi-ring basins, but no agreement exists in the literature to date – see Melosh (1989) for a discussion. Melosh (1989) preferred the so-called ring tectonic theory, where the thickness of the lithosphere plays a dominant role in determining whether or not a ring forms. More recently, a nested melt-cavity model has been proposed to account for transition from complex craters to multi-ringed basins on the Moon (Head, 2010).

Complications arise, as external rings have been documented around much smaller impact structures, such as the proposed (but not confirmed) approximately 20 km diameter Silverpit structure in the North Sea (Fig. 1.9; Stewart and Allen, 2002). Numerical modelling suggests this morphology formed due to an impact into a layer of brittle chalk overlying weak shales (Stewart and Allen, 2002; Collins et al., 2003). Whether multi-ring basins exist on the Earth also remains a topic of debate. Of the three largest structures on Earth (Chicxulub, Sudbury and Vredefort), Chicxulub is the best-preserved large terrestrial impact structure, due to burial. As such, however, the definition of its morphological elements depends on the interpretation of geophysical data. It has an interior topographic ‘peak-ring’, a terraced rim area and exterior ring faults and, therefore, appears to correspond to the definition of a multi-ring basin, as on the Moon (Grieve et al., 2008).

Figure 1.9 Perspective view of the top chalk surface at the Silverpit structure, North Sea, UK, a suspected meteorite impact structure. The central crater is 2.4 km wide and is surrounded by a series of concentric faults, which extend to a radial distance of approximately 10 km from the crater centre. False colours indicate depth (yellow: shallow; purple: deep). Image courtesy of Phil Allen and Simon Stewart. (See Colour Plate 5)
1.4 Impactites

In terms of the products of meteorite impact events, the above considerations of the impact cratering process reveal that pressures and temperatures that can vaporize, melt, shock metamorphose and/or deform a substantial volume of the target sequence can be generated. The transport and mixing of impact-metamorphosed rocks and minerals during the excavation and formation of impact craters produces a wide variety of distinctive impactites that can be found within and around impact craters (see Fig. 1.10; ‘rock affected by impact metamorphism’) (Stöffler and Grieve 2007) – see Chapter 7. Much of our knowledge of impactites comes from impact craters on Earth and, to a lesser extent the Moon, where large numbers and volumes of samples from known locations are available for study.

The transient compression, decompression and heating of the target rocks lead to shock metamorphic effects (see Chapter 8 for an overview), which record pressures, temperatures and strain rates well beyond those produced in terrestrial regional or contact metamorphism (Fig. 1.8 and Fig. 1.9). Given the highly transient nature of shock metamorphic processes, disequilibrium and metastable equilibrium are the norm. The only megascopic shock morphic feature are shatter cones, which are distinctive, striated and horse-tailed conical fractures ranging in size from millimetres to tens of metres (Fig. 1.11a). The most-documented shock metamorphic feature is the occurrence of so-called planar deformation features, particularly in quartz (Fig. 1.11b), although they do occur in other minerals (e.g. feldspar and zircon). When fresh, the planar deformation features are parallel planes of glass, with specific crystallographic orientations as a function of shock pressures of approximately 10–35 GPa. At higher pressures, the shock wave can destroy the internal crystallographic order of feldspars and quartz and convert them to solid-state glasses, which still have the original crystal shapes. These are ‘diaplectic’ glasses (Fig. 1.11c,d), with the required pressures being 30–45 GPa for plagioclase feldspar (also known as maskelynite) and 35–50 GPa for quartz. The extremely rapid compression and then decompression also produces metastable polymorphs, including coesite and stishovite from quartz and diamond and lonsdaleite from graphite (Chapter 8).

1.4.1 Classification of impactites

As part of the IUGS Subcommission on the Systematics of Metamorphic Rocks, a study group formulated a series of recommendations for the classification of impactites (Stöffler and Grieve, 2007). This group suggested that impactites from a single impact should be classified into three major groups irrespective of their geological setting: (1) shocked rocks, which are non-breciated, melt-free rocks displaying unequivocal effects of shock metamorphism; (2) impact melt rocks (Fig. 1.10a–c), which can be further subclassified according to their clast content (i.e. clast-free, -poor or -rich) and/or degree of crystallinity (i.e. glassy, hypocrystalline or holocrystalline); (3) impact breccias (Fig. 1.10d,e), which can be further classified according to the degree of mixing of various target lithologies and their content of melt particles (e.g. lithic breccias and ‘suevites’).

It is apparent from the literature that substantial problems exist with the current IUGS nomenclature of impactites, particularly those including impact melt products (see Chapters 7 and 9 for detailed discussions). This is due to several reasons, including the erosional degradation of many impact structures on Earth such that outcrops of impact melt-bearing lithologies preserving their entire original context are relatively rare (Grieve et al., 1977). Other complicating factors are introduced due to inconsistent nomenclature and unqualified use of terms (such as ‘suevite’ – Fig. 1.10d) for several types of impactites with somewhat different genesis; for example, impactites with glass contents ranging up to approximately 90 vol.% have been termed suevites at the Popigai impact structure (Masaitis, 1999). It is also important to note that the framework for the IUGS classification scheme was developed in the 1990s and remained little changed up to its publication in 2007, despite several major discoveries and advancements in our understanding of impactites. In particular, in recent years, the effect(s) of target lithology on various aspects of the impact cratering process, in particular the generation and emplacement of impactites, has emerged as a major research topic (Oskinia et al., 2008a).

1.4.2 Impact melt-bearing impactites

The production of impact melt rocks and glasses is a diagnostic feature of hypervelocity impact, and their presence, distribution and characteristics have provided valuable information on the cratering process (Dence et al., 1977; Grieve et al., 1977; Grieve and Cintala, 1992) – see Chapter 9. Within complex impact structures formed entirely in crystalline targets, coherent impact melt rocks or ‘sheets’ are formed. These rocks can display classic igneous structures (e.g. columnar jointing) and textures (Fig. 1.10a–c). Impact craters formed in ‘mixed’ targets (e.g. crystalline basement overlain by sedimentary rocks) display a wide range of impact-generated lithologies, the majority of which were typically classed as ‘suevites’ (Fig. 1.10d; Stöffler et al., 1977; Masaitis, 1999); the original definition of a suevite is a polymict impact breccia with a clastic matrix/groundmass containing fragments and shards of impact glass and shocked mineral and lithic clasts (Stöffler et al., 1977). Minor bodies of coherent impact melt rocks are also sometimes observed, often as lenses and irregular bodies within larger bodies of suevite (e.g. Masaitis, 1999). In impact structures formed in predominantly sedimentary targets, impact melt rocks were not generally recognized, with the resultant crater-fill deposits historically referred to as clastic, fragmental or sedimentary breccias (Masaitis et al., 1980; Redeker and Stöffler 1988; Masaitis 1999). These observations led to the conclusion

\(^2\)Shock metamorphism is defined as the metamorphism of rocks and minerals caused by shock wave compression and decompression due to impact of a solid body or due to the detonation of high-energy chemical or nuclear explosives (Stöffler and Grieve, 2007).

\(^1\)Impact metamorphism is essentially the same as shock metamorphism except that it also encompasses the melting and vaporization of target rocks (Stöffler and Grieve, 2007).
Figure 1.10  Field images of impactites. (a) Oblique aerial view of the approximately 80 m high cliffs of impact melt rock at the Mistastin impact structure, Labrador, Canada. Photograph courtesy of Derek Wilton. (b) Close-up view of massive fine-grained (aphanitic) impact melt rock from the Discovery Hill locality, Mistastin impact structure. Camera case for scale. (c) Coarse-grained granophyre impact melt rock from the Sudbury Igneous Complex, Canada. Rock hammer for scale. (d) Impact melt-bearing breccia from the Mistastin impact structure. Note the fine-grained groundmass and macroscopic flow-textured silicate glass bodies (large black fragments). Steep Creek locality. Marker/pen for scale. (e) Polymict lithic impact breccias from the Wengenhausen quarry, Ries impact structure, Germany. Rock hammer for scale. (f) Carbonate melt-bearing clast-rich impact melt rocks from the Haughton impact structure. Penknife for scale. This lithology was originally interpreted as a clastic or fragmental breccia (Redeker and Stöffler, 1988), but was subsequently shown to be an impact melt-bearing impactite (Osinski and Spray, 2001; Osinski et al., 2005b). (See Colour Plate 6)
1.5 Recognition of impact craters

Several criteria may be used to recognize hypervelocity impact structures, including the presence of a crater form and/or unusual rocks, such as breccias, melt rocks and pseudotachylite; however, on their own, these indicators do not provide definitive evidence for a meteorite impact structure. Geophysics can also provide clues (see Chapter 14), and a geophysical anomaly is often the first indicator of the existence of buried structures. The most common geophysical anomaly is a localized low in the regional gravity field, due to lowering of rock density from brecciation and fracturing (Pilkington and Grieve, 1992). Larger complex impact structures tend to have a central, relative gravity high, which can extend out to approximately half the diameter of the structure. In terms of magnetics, the most common expression is a magnetic...
low, with the disruption of any regional trends in the magnetic field. This is due to an overall lowering of magnetic susceptibility and the randomizing of pre-impact lithologic trends in the target rocks (Pilkington and Grieve, 1992). Seismic velocities are reduced at impact structures, due to fracturing, and reflection seismic images are extremely useful in characterizing buried structures in sedimentary targets. There is, however, no geophysical anomaly that can provide definitive evidence for a meteorite impact structure.

The general consensus within the impact community is that unequivocal evidence for hypervelocity impact takes the form of shock metamorphic indicators (French and Koeberl, 2010), either megascopic (e.g. shatter cones) or microscopic (e.g. planar deformation features, diapeptic glass) (Fig. 1.11), and the presence of high-pressure polymorphs (e.g. coesite and stishovite) – see Chapter 8 for an overview of shock metamorphism. Unfortunately, this requires investigation and preservation of suitable rocks within a suspected structure. However, this is often not possible for eroded and/or buried structures and/or structures presently in the marine environment (e.g. the Eltanin feature in the South Pacific; Kyte et al., 1988), even though there is strong evidence for an impact origin.

A prime example is the controversy surrounding the Silverpit structure in the North Sea. Stewart and Allen (2002) originally proposed that this structure was an impact crater based on high-resolution three-dimensional (3D) seismic data (Fig. 1.9); and despite some opposition (Thomson et al., 2005), most impact workers accept this; however, without drilling to retrieve samples, this structure is currently relegated to the list of ‘possible’ impact structures. This is unfortunate, as the seismic dataset for this structure surpasses that available for any known impact structure and may provide important insights into complex crater formation. In order to try to address this issue, Stewart (2003) proposed a framework for the identification of impact structures based on 3D seismic data, but this has received little attention to date within the impact community.

### 1.6 Destructive effects of impact events

In 1980, Luis and Walter Alvarez and colleagues published a paper in Science outlining evidence for an extraterrestrial origin for the most recent of the ‘big five’ mass extinctions: the Cretaceous–Tertiary (now the Cretaceous–Paleogene) mass extinction event at approximately 65 Ma (Alvarez et al., 1980). A decade later, the source crater – the approximately 180km diameter Chicxulub impact structure – was found lying beneath approximately 1 km of sediment below and half offshore the present-day Yucatan Peninsula, Mexico (Hildebrand et al., 1991). As outlined in Chapter 10, the Chicxulub impact caused severe environmental effects that ranged from local to global and that lasted from seconds to tens of thousands of years. The local and regional effects of the impact event include the air blast and heat from the impact explosion, tsunamis and earthquakes. Global effects included forest fires ignited by impact ejecta re-entering the Earth’s atmosphere, injection of huge amounts of dust in the upper atmosphere, which may have inhibited photosynthesis for as much as 2 months, and the production of vast quantities of N₂O from the shock heating of the atmosphere (Chapter 10). However, one of the most important findings has been that, in terms of global effects, the severity of the Chicxulub impact was due, in part, to the composition of the target rocks: approximately 3 km of carbonates and evaporites overlaying crystalline basement.

While it was initially thought that the vaporization and decomposition of carbonates – producing CaO and releasing CO₂, resulting in global warming – was important (O’Keefe and Ahrens, 1989), it appears that the most destructive effect(s) came from the release of sulfur species from the evaporite target rocks (Pope et al., 1997). We know from studies of sulfur-rich volcanic eruptions, such as Mount Pinatubo in 1992, that sulfur aerosols can significantly reduce the amount of sunlight that reaches the Earth’s surface, resulting in short-term global cooling. Estimates for Chicxulub suggest as much as a 15°C decrease in the average global temperatures, which when coupled with the other effects of the impact event would have resulted in severe environmental consequences (see Chapter 10 for an overview).

### 1.7 Beneficial effects of impact events

#### 1.7.1 Microbiological effects

As noted in Section 1.6, ever since the proposal of a link between meteorite impacts and mass extinctions, the deleterious effects of impact events have received much attention (Schulte et al., 2010). However, research conducted over the past few years indicates that, although meteorite impacts are indeed destructive, catastrophic events, there are several potential beneficial effects, particularly in terms of providing new habitats for microbial communities (Cockell and Lee, 2002) – see Chapter 11 for an overview. This may have important implications for understanding the origin and evolution of life on Earth and other planets such as Mars.

One of the most important beneficial effects is the generation of a hydrothermal system within an impact crater immediately following its formation. As noted in Section 1.2.4, recent work suggests that impact-associated hydrothermal systems will form following impacts into any H₂O-bearing solid planetary body, with exceptions for small impacts and those in extremely arid regions (Naumov, 2005; Osinski et al., 2012) – see Chapter 6. Numerical models of these hydrothermal systems suggest that they may last several million years for large, 100 km size, impact structures (Abramov and Kring, 2004, 2007). This may have important astrobiological implications, as many researchers believe that hydrothermal systems in general might have provided habitats or ‘cradles’ for the origin and evolution of early life on Earth (Farmer, 2000) and possibly other planets, such as Mars. Excitingly, the first clear evidence for impact-generated hydrothermal systems on Mars has recently been discovered (Marzo et al., 2010).

Other potential habitats exclusive to impact craters include impact-processed crystalline rocks (Cockell et al., 2003), which
have increased porosity and translucence compared with unshocked materials, improving microbial colonization, impact-generated glasses (Sapers et al., 2010), and impact crater lakes, which form protected sedimentary basins that can provide protective environments and increased preservation potential of fossils and organic material (Cockell and Lee, 2002).

1.7.2 Economic effects

One of the less well-known aspects of meteorite impact craters, at least in the general scientific community, is the potential association of economic mineral and hydrocarbon deposits, and thus their suitability as exploration targets (see Chapter 12 for an overview). This is exemplified by the large, approximately 200–250 km diameter Sudbury (Canada) and Vredefort (South Africa) impact structures, which host some of the world’s largest and most profitable mining camps (Grieve, 2005). As outlined in Chapter 12, economic resources associated with impact craters can be classified as either pre-, syn- or epigenetic with respect to the impact event. At Vredefort, the impact event led to the preservation of pre-impact (progenetic) gold and uranium deposits in the Witwatersrand Basin and their subsequent mobilization and concentration during impact-induced hydrothermal alteration, producing the world’s richest gold province. In contrast, at Sudbury, the world’s largest nickel–copper ore deposits occur at the base of the impact melt sheet and in radial dikes. These ore deposits are syngenetic and formed through the separation of immiscible sulfide liquids from the silicate impact melt. Subsequent post-impact hydrothermal activity also led to the formation of copper–platinum group element-rich and zinc–copper–lead economic ore deposits at Sudbury (Ames and Farrow, 2007). Economic ore deposits also occur at a number of other smaller terrestrial impact structures, and the lack of detailed studies of many impact sites leaves room for further discoveries.

In addition to economic metalliferous ore deposits, several meteorite impact structures have been exploited for hydrocarbons (Donofrio, 1998). The fracturing and faulting of rocks in central uplifts and faulted crater rims, results in enhanced porosity and permeability, providing valuable reservoirs for oil and gas, even in rocks such as granites that are typically not suitable hydrocarbon reservoirs (e.g. Ames structure, USA; Chapter 12). Post-impact sedimentary crater-fill deposits can also generate suitable source rocks.

1.8 When a crater does not exist: other evidence for impact events

The subject of this book is impact cratering, which implies that the emphasis is on cratering. However, it is important to note that not all impact events result in the formation of an impact crater. This is obvious for impacts into the Jovian planets, such as the collision of comet Shoemaker–Levy 9 with Jupiter in 1992, but perhaps less so for Earth. The documentation of spherule beds and tektites is a topic that is discussed in chapters on impact ejecta (Chapter 4) and impact melting (Chapter 9). While many of these occurrences have been linked to source craters, the majority, particularly spherule beds in Archaean-age rocks, have not (Simonson and Glass, 2004). These distal ejecta deposits do, however, provide important information regarding the impact cratering process and should not be overlooked. These Archaean-age spherule beds found in South Africa and Australia are also the most ancient part of the impact record on Earth.

In addition to distal ejecta deposits, there is an increasing discovery of occurrences of natural glasses around the world that are neither spherules nor tektites (e.g. Fig. 1.12). These glasses are either confirmed or suspected as being of impact origin but for which no source crater has been recognized. Some of these glass occurrences are well known and widely accepted as being of impact origin; for example, Libyan Desert Glass (Weeks et al., 1984), Darwin Glass (Meisel et al., 1990), urengoiites or South Ural Glass (Deutsch et al., 1997) and Dakhleh Glass (Osinski et al., 2007; Fig. 1.12). Others remain more enigmatic (Haines et al., 2001; Schultz et al., 2006). Several of these occurrences have been ascribed to large aerial bursts or airbursts.

The 1908 Tunguska event represents the largest recorded example of an airburst event on Earth to date (Vasileyv, 1998), with estimated magnitude estimates ranging from 3–5 Mt up to approximately 10–40 Mt. Theoretical calculations coupled with ground- and satellite-based observations of airbursts suggest that the Earth is struck annually by objects of energy 2–10 kt with Tunguska-size events occurring once every 1000 years (Brown et al., 2002). Recent numerical modelling suggests that substantial amounts of glass can be formed by radiative/convective heating of the surface during greater than 100 Mt low-altitude airbursts (Boslough, 2006). When coupled with the observations of natural glasses described above, there is, therefore, growing evidence to suggest that airbursts – and the glass produced by such events – should occur more frequently than has been previously recognized in the geological record.

1.9 Concluding remarks

The recognition of impact cratering as a fundamental geological process represents a revolution in Earth and planetary sciences. The study of impact craters and related phenomena is relatively young when compared with other fields of geological study. It is clear that the formation of meteorite impact structures is unlike any other geological process; however, this should not hinder their study. Far from it: coming to terms with understanding a geological process that takes place in only a few seconds to minutes, with energies that can be greater than the total annual internal energy release from the Earth, provides a stimulating framework for research and teaching. What is more, the study of impact craters requires a multi- and inter-disciplinary approach and must take into account observations from throughout the Solar System. Unlike many areas in the geological sciences, there is, therefore, still considerable potential for new and exciting contributions and areas of study.

As outlined above and in other chapters in this book, basic processes such as the mechanics of complex crater formation
Figure 1.12 Field images of Dakhleh Glass, the potential product of an airburst event (Osinski et al., 2007; Osinski et al., 2008c). (a) Area of abundant Dakhleh Glass lagged on the surface of Pleistocene lacustrine sediments; the arrows point to some large Dakhleh Glass specimens. (b) Upper surfaces of many large Dakhleh Glass lag samples appear to be in place and are highly vesiculated. This contrasts with the smooth, irregular lower surfaces. (c) In cross-section, it is clear that there is an increase in the number of vesicles towards the upper surface. Together, these features are indicative of ponding of melt and volatile loss through vesiculation. (d) Highly vesicular pumice-like Dakhleh Glass sample. 7 cm lens cap for scale. (See Colour Plate 7)

(see Chapter 5) and the production of vapour plumes and ejecta deposits (see Chapters 3 and 4) are still not fully understood. Major questions concerning the effect of target properties (e.g. volatiles, porosity, layering) on the impact cratering process and the environmental effects of impact events still remain to be resolved. As the exploration of our Solar System continues, furthering our understanding of impact cratering will become even more important.

References


