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The recognition of terrestrial impact structures

ANN M. THERRIAULT – RICHARD A. F. GRIEVE – MARK PILKINGTON

Natural Resources Canada, Booth Street, Ottawa, Ontario, KIA 0ES Canada; e-mail: ATherria@NRCan.gc.ca

Abstract. The Earth is the most endogenically active of the terrestrial planets and, thus, has retained the poorest sample of impacts that have occurred throughout geological time. The current known sample consists of approximately 160 impact structures or crater fields. Approximately 30% of known impact structures are buried and were initially detected as geophysical anomalies and subsequently drilled to provide geologic samples. The recognition of terrestrial impact structures may, or may not, come from the discovery of an anomalous quasi-circular topographic, geologic or geophysical feature. In the geologically active terrestrial environment, anomalous quasi-circular features, however, do not automatically equate with an impact origin. Specific samples must be acquired and the occurrence of shock metamorphism, or, in the case of small craters, meteoritic fragments, must be demonstrated before an impact origin can be confirmed. Shock metamorphism is defined by a progressive destruction of the original rock and mineral structure with increasing shock pressure. Peak shock pressures and temperatures produced by an impact event may reach several hundreds of gigaPascals and several thousand degrees Kelvin, which are far outside the range of endogenic metamorphism. In addition, the application of shock-wave pressures is both sudden and brief. Shock metamorphic effects result from high strain rates, well above the rates of normal tectonic processes. The well-characterized and documented shock effects in quartz are unequivocal indicators and are the most frequently used indicator for terrestrial impact structures and lithologies.

Key words: Earth, impact structures, shock metamorphism, melting, glasses, shatter cones, geophysical anomaly

Introduction

On Earth, compared to the other terrestrial planets, the very active geologic environment tends to modify and destroy the impact crater record. Approximately 160 impact structures or crater fields are currently known on Earth. Impact involves the transfer of massive amounts of energy to a spatially limited area of the Earth's surface, in an extremely short time interval. As a consequence, local geology of the target area is of secondary importance. The effects of impact are, however, scale-dependent and show progressive changes with increasing energy of the impact event. The net result is that, impacts of similar scale produce similar first-order geological and geophysical effects. Thus, general observations can be derived with respect to the appearance and geological and geophysical signatures of terrestrial impact structures, in specific size ranges.

Terrestrial impact structures were first recognized by their bowl-like shape and meteorite fragments found in their vicinity or within them (the classic example being Meteor or Barringer Crater, Arizona). In the 1960's, petrographic studies of rocks from impact structures defined a series of unique characteristics produced by a style of deformation called shock metamorphism (e.g. French and Short 1968). Shock metamorphic effects include shatter cones (e.g. Dietz 1947, Milton 1977), the only macroscopic diagnostic shock effect observed at terrestrial impact structures, a number of microscopic effects in minerals, some of which are diagnostic of shock, and impact melting.

The aim of this paper is to summarize the morphology and geoscientific aspects of terrestrial impact structures and provide a general description of shock-metamorphic effects.

Morphology

On most planetary bodies, well-preserved impact structures are recognized by their characteristic morphology and morphometry. The basic shape of an impact structure is a depression with an upraised rim. Detailed appearance, however, varies with crater diameter. With increasing diameter, impact structures become proportionately shallower and develop more complicated rims and floors, including the appearance of central peaks and interior rings. Impact craters are divided into three basic morphologic subdivisions: simple craters, complex craters, and basins (Dence 1972, Wood and Head 1976).

Small impact structures have the form of a bowl-shaped depression with an upraised rim and are known as simple craters (Fig. 1). The exposed rim, walls, and floor define the so-called apparent crater. At the rim, there is an overturned flap of ejected target materials, which displays inverted stratigraphy, with respect to the original target materials. Beneath the floor is a lens of brecciated target material that is roughly parabolic in cross-section (Fig. 2). This breccia lens is allochthonous and polymict, with fractured blocks of various target materials. In places, near the top and the base, the breccia lens may contain highly shocked, and possibly melted, target materials. Beneath the breccia lens, parautochthonous, fractured rocks define the walls and floor of what is known as the true crater. In the case of terrestrial simple craters, the depth to the base of the breccia lens (i.e., the base of the true crater) is roughly twice that of the depth to the top of the breccia lens (i.e., the base of the apparent crater, Fig. 2). Shocked rocks in the parautochthonous materials of the true crater floor are confined to a small central volume at the base of the true crater.

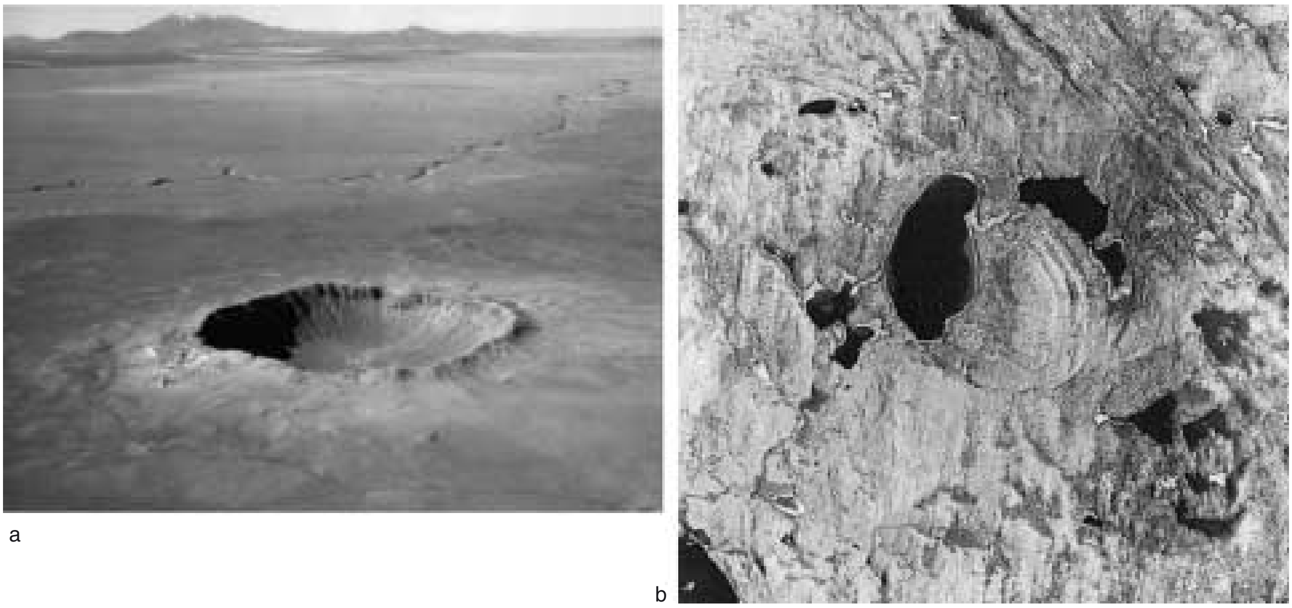


Figure 1. (a) Oblique aerial view of 1.2 km diameter, 50,000 years old simple crater, Meteor or Barringer Crater, Arizona, U.S.A. (b) Vertical aerial view of 3.8 km diameter, 450 ± 30 million years old, Brent Crater, Ontario, Canada. Note how this ancient crater has no rim, has been filled by sediments and lakes and is a generally subtle topographic feature.

With increasing diameter, simple craters show increasing evidence of wall and rim collapse and evolve into complex craters (Fig. 3). The transition diameter varies between planetary bodies and is, to a first approximation, an inverse function of planetary gravity (Pike 1980). Other variables, such as target material and possibly projectile type and velocity, play a lesser role, so that the transition diameter varies over a small range. The most obvious effect of secondary variables appears on Earth, where there are major areas of both sedimentary and crystalline rocks at the surface. Complex craters on Earth first occur at diameters greater than 2 km in layered sedimentary target rocks but not until diameters of 4 km or greater in stronger, more coherent, igneous or metamorphic, crystalline target rocks (Dence 1972).

With a central topographic peak or peaks, a broad, flat floor, and terraced, inwardly slumped rim areas (Fig. 4), complex craters are a highly modified craterform compared to simple craters. The rim of a typical complex crater is a structural feature corresponding to a series of fault terraces. Interior to the rim lays an annular trough, which is partially filled by a sheet of impact-melt rock and/or polymict allochthonous breccia (Fig. 4). Only in the central area of the crater is there evidence of substantial excavation of target materials. This region is structurally complex and, in large part, occupied by a central peak, which is the topographic manifestation of a much broader and extensive area of uplifted rocks that occurs beneath the center of complex craters. Readers interested in the details of cratering mechanics at simple and complex structures are referred to Melosh (1989) and references therein.

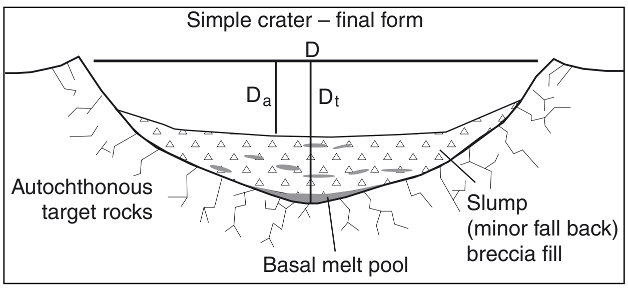
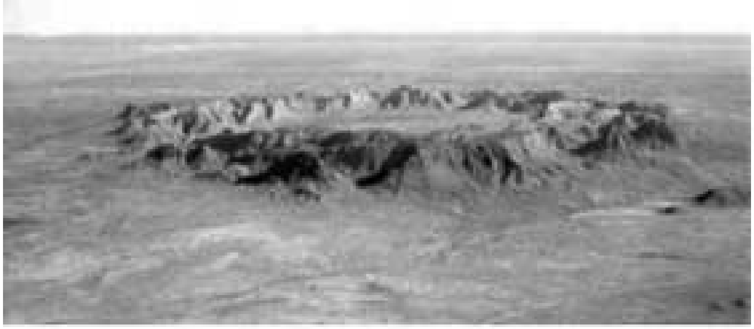


Figure 2. Schematic cross-section of a simple crater. D is the diameter and d_a and d_t are the depths of the apparent and true crater, respectively. See text for details.

With increasing diameter, a fragmentary ring of interior peaks appears, marking the transition from craters to basins. While a single interior ring is required to define a basin, basins have been subdivided, with increasing diameter, on other planetary bodies, into central-peak basins, with both a peak and ring; peak ring basins, with only a ring; and multi-ring basins, with two or more interior rings (Wood and Head 1976). There have been claims that the largest known terrestrial impact structures have multi-ring forms, e.g. Chicxulub, Mexico (Sharpton et al. 1993), Sudbury, Canada (Stöffler et al. 1994, Spray and Thompson 1995) and Vredefort, South Africa (Therriault et al. 1997). Although certain of their geological and geophysical attributes form annuli, it is not clear that these correspond, or are related in origin, to the obvious topographical rings observed, for example, in lunar multi-ring basins (Spudis 1993, Grieve and Therriault 2000).

Most terrestrial impact structures are affected by erosion. In extreme cases, the craterform has been completely removed. In such cases, recognition of structural and



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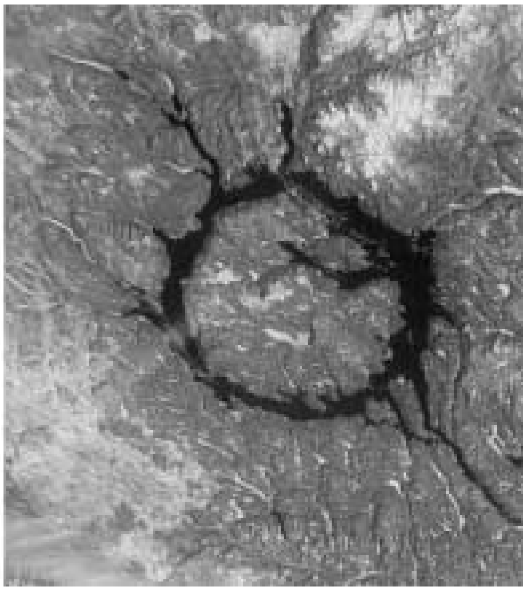

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Figure 3. (a) Oblique aerial photograph of the Gosses Bluff impact structure, Australia. Note that all that is visible of this originally 22 km, 142.5 ± 0.8 million years old structure is a 5 km annulus of hills, representing the eroded remains of a central uplift. See text for details. (b) Shuttle photograph of the Manicouagan impact structure, Canada, 100 km in diameter and 214 ± 1 million years old. Note that the annular trough (with a diameter of ~ 65 km) is filled by water.

geological effects of impact in the target rocks is essential to the identification of an impact structure rather than the presence of a characteristic craterform. For example, Gosses Bluff, Australia has a positive topographical form consisting of an annular ring of hills, approximately 5 km in diameter (Fig. 3). The ring consists of erosionally resistant beds from within the original central uplifted area of a complex impact structure. The original craterform, which has an estimated diameter of approximately 22 km (Milton et al. 1996), has been removed by erosion. There are several other impact structures, which have some form of rings, e.g. Manicouagan, Canada (Floran and Dence 1976), Haughton, Canada (Robertson and Sweeney 1983), but it is not clear whether these are primary forms or secondary features, with some relation to primary structural features (Grieve and Head 1983).

There are also other subtleties to the character of craterforms in the terrestrial record that do not appear on the other terrestrial planets. A number of relatively young, and, therefore, only slightly eroded, complex impact structures (e.g. Haughton, Canada; Ries Germany; Zhamanshin, Kazakhstan) do not have an emergent central peak or other interior topographical expression of a central uplift (Garvin and Schnetzler 1994). These structures are in mixed targets of platform sediments overlying crystalline basement. Although there are no known comparably young complex structures entirely in crystalline targets, the buried and well-preserved Boltysh structure, Ukraine, which is of comparable size, has a central peak (emergent from the crater-fill), similar to the appearance of lunar central peak craters. This difference in form is probably a target rock effect but it has not been studied in detail.

The morphology of impact craters formed in marine environment is also quite distinct. These impact structures are characterized by a broad and shallow brim at the periphery of the crater, extensive infilling, and prominent fault blocks floored by apparent low-angle décollement surfaces at the periphery of the crater (e.g. Tsikalas et al. 1999, Ormö and Lindström 2000). The extensive infilling is most likely due to large amounts of ejecta and crater wall material transported into the excavated crater by the collapse of the impact-induced water cavity and the subsequent rapid surge of sea water (Tsikalas et al. 1999, Ormö and Lindström 2000). The 40-km-diameter Mjølnir submarine impact structure in the Barents Sea, for example, consists of a central region of deep excavation surrounded by a shallow excavated shelf, without a raised crater rim (Tsikalas et al. 1998, 1999). This morphology is also observed at the 13.5-km-diameter Lockne impact structure, Sweden (Lindström et al. 1996).

Attempts to define morphometric relations, particularly depth-diameter relations, for terrestrial impact structures have had limited success, because of the effects of erosion and, to a lesser degree, post-impact sedimentation. Unlike depth, the variation of stratigraphic uplift (*SU*, Fig. 4) with diameter at complex impact structures is fairly

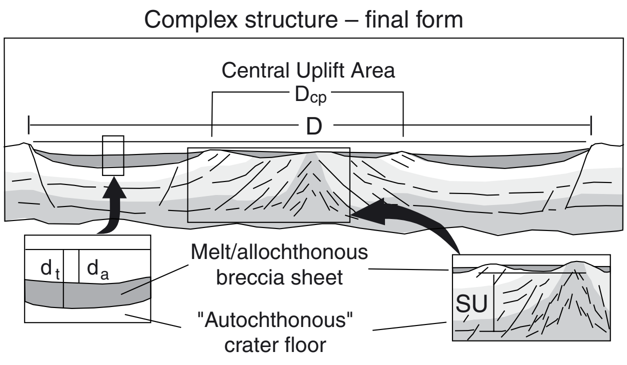


Figure 4. Schematic cross-section of complex impact structure. Notation as in Figure 2, with SU corresponding to structural uplift and D_{cp} to the diameter of the central uplift. Note preservation of beds in outer annular trough of the structure, with excavation limited to the central area. See text for details.

well constrained with $SU = 0.86D^{1.03}$ (n = 24), where n is the number of data points (Grieve and Pilkington 1996). Similarly, the diameter of the central uplift area (D_{cp}, Fig. 4), at its maximum radial expression, is constrained by $D_{cp} = 0.31\ D^{1.02}$ (n = 44) (Therriault et al. 1997).

Geophysics of impact structures

Geophysical anomalies over terrestrial impact structures vary in their character and, in isolation, do not provide definitive evidence for an impact origin. About 30 per cent of known terrestrial impact structures are buried by post-impact sediments. Geophysical methods resulted in their initial discovery and subsequent drilling provided geologic samples, which confirmed their impact origin. Interpretation of a single geophysical data set over a suspected impact structure can be ambiguous (for example, Hildebrand et al. 1998, Sharpton et al. 1993). When combined, however, with complementary geophysical methods and the existing database over other known impact structures, a more definite assessment can be made (e.g. Ormö et al. 1999).

Since potential-field data are available over large areas, with almost continuous coverage, gravity and magnetic observations have been the primary geophysical indicators used for evaluating the occurrence of possible terrestrial impact structures. Seismic data, although providing much better spatial resolution of subsurface structure, is used less, because it is less generally available. Electrical methods have been used even less (e.g. Henkel 1992). Given space limitations and some lack of specificity of the geophysical attributes of terrestrial impact craters, they are generally discussed here and the reader is referred to the most recent synthesis in Grieve and Pilkington (1996).

Gravity signature

The most notable geophysical signature associated with terrestrial impact structures is a negative gravity anomaly. These gravity lows are generally circular, extending to, or slightly beyond, the crater rim, and are due to lithological and physical changes associated with the impact process. In well-preserved impact structures, low-density sedimentary infill of the topographic depression of the crater contributes to the gravity low. In complex impact structures, relatively lower density impact-melt sheets also contribute to the negative gravity effect. However, such lithological effects are minor compared to density contrasts induced by fracturing and brecciation of the target rocks.

In general, the amplitude of the maximum negative gravity anomaly associated with impact structures increases with the final crater diameter (Dabizha and Fedynsky 1975, Dabizha and Feldman 1982). Over simple craters, a circular bowl-shaped negative anomaly is observed; whereas most of larger complex impact structures, greater than 30 km in diameter, tend to exhibit a central gravity high. Based on data from 58 terrestrial impact structures, Pilkington and Grieve (1992) showed that erosional level has only a secondary effect on gravity anomaly size.

It is important to note that due to differences in target lithologies, large variations in gravity signature are observed between structures of similar sizes. In general, structures formed in sedimentary lithologies produce smaller anomalies than similar sized ones formed in crystalline rocks. Structures formed in unconsolidated sediments in continental shelf areas may not produce detectable negative gravity anomalies but are marked only by a central gravity high.

Magnetic signature

In general, magnetic anomalies associated with terrestrial impact structures are more complex than gravity anomalies. This observation reflects the greater variation possible in the magnetic properties of rocks. The dominant effect over impact structures is a magnetic low or subdued zone ranging in amplitude from tens to a few hundred nanotesla that is commonly manifested as a truncation of the regional magnetic fabric (Dabizha and Fedynsky 1975, Clark 1983). Magnetic lows are best defined over simple and some small complex craters, where the anomaly is smooth and simple; whereas at larger impact structures, the magnetic low can be modified by the presence of shorter-wavelength, large-amplitude, localized anomalies that usually occur at or near the centre of the structure.

No correspondence exists between the magnetic anomaly character and crater morphology of impact structures. Moreover, the presence of a central gravity high does not imply the existence of a central magnetic anomaly. There are several structures with no obvious magnetic signature.

Shock effects, thermal effects or chemical effects may cause magnetic anomalies related to impact. Shock effects in impact structures can serve to increase or decrease magnetization levels. Thermal effects may result in the production of non-magnetic impact glasses (Pohl 1971) or in resetting magnetic minerals through thermoremanent magnetization in the direction of the Earth's magnetic field at the time of impact. Chemical effects may result in the production of new magnetic phases, through elevated residual temperatures and hydrothermal alteration, leading to the acquisition of a chemical remanent magnetization in the direction of the ambient field.

Seismic signature

Reflection seismic surveys allow for detailed imaging of impact structure morphology and delineating zones of incoherent reflections that are characteristic of brecciation and fracturing. The disturbance of coherent subsurface reflectors is most prominent in the central uplift of complex structures and decreases outward and downward from this zone (Brenan et al. 1975). Reflection data can provide estimates of such morphological parameters as the dimensions of the central uplift, annular trough and faulted blocks at the structural rim of complex structures (e.g. Morgan et al. 2002). The depth to horizontal reflectors that exist below the crater floor can be used to determine the amount of structural uplift.

Electrical signature

The presence of fluids in impact-induced fractures and pore spaces leads to decreased resistivity levels that can be mapped effectively by various electrical methods. The conductivity of rocks is heavily dependent on their water content: < 1% change in water content can produce more than an order of magnitude change in conductivity. The degree of fragmentation determines the amount and distribution of fluids within the rock and hence, its electrical properties.

Where a distinct contrast exists between the allochthonous breccia deposits and the underlying autochthonous target rocks, electrical profiling using resistivity sounding can map the structure of the true crater floor (e.g. Vishnevsky and Lagutenko 1986). In order to determine the deeper electrical structure associated with impact, magnetotelluric surveys have been carried out (e.g. Zhang et al. 1988, Campos-Enriquez et al. 1997).

Geology of impact structures

Although an anomalous circular topographic, structural, or geological feature may indicate the presence of an impact structure, there are other endogenic geological processes that can produce similar features in the terrestrial environment. An obvious craterform is an excellent indicator of a possible impact origin; particularly, if it has the appropriate morphometry, but as noted, such features are rare and short-lived in the terrestrial environment. The burden of proof for an impact origin generally lies with the documentation of the occurrence of shock-metamorphic effects.

Few structures preserve physical evidence of the impacting body. Such structures are limited to small, young, simple structures, where the impacting body (or, more commonly, fragments of it) has been slowed by atmospheric deceleration and impacts at less than cosmic velocity. These are restricted generally to the impact of iron or stony-iron meteorites. Stony meteorites are weaker than their iron-bearing counterparts and small stones are generally crushed as a result of atmospheric interaction (Melosh 1981). Larger impacting bodies (>100–150 m in diameter) survive atmospheric passage with undiminished impact velocity. Consequently, the peak shock pressures upon impact are sufficient, in most cases, to result in the melting and vaporisation of the impacting body, destroying it as a physical entity.

On impact, the bulk of the impacting body's kinetic energy is transferred to the target by means of a shock wave. This shock wave imparts kinetic energy to the target materials, which leads to the formation of a crater. It also increases the internal energy of the target materials, which leads to the formation of so-called shock-metamorphic effects. The details of the physics of impact and shockwave behavior can also be found in Melosh (1989), and references therein.

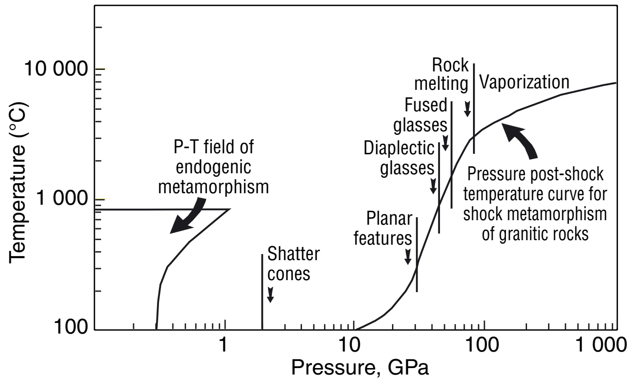


Figure 5. Temperature and pressure range of shock metamorphic effects compared to that of endogenic metamorphism. Planar features include planar deformation features (PDFs) and planar fractures (PFs). Scale is log-log. See text for details.

Shock metamorphism is the progressive breakdown in the structural order of minerals and rocks due to the passage of a high-pressure shock wave and requires pressures and temperatures well above the pressure-temperature field of endogenic terrestrial metamorphism (Fig. 5). The dependence on high pressures for the formation of shock-metamorphic effects has been shown by their duplication in nuclear and chemical explosion craters, and in laboratory shock recovery experiments (e.g. Hörz 1968, Müller and Hornemann 1969, Borg 1972). Minimum shock pressures required for the production of diagnostic shock-metamorphic effects are 5–10 GPa for most silicate minerals. Strain rates produced by impact cratering process are of the order of 10^6 s^{-1} to 10^9 s^{-1} (Stöffler and Langenhorst 1994), many orders of magnitude higher than typical tectonic strain rates (10^{-12} s^{-1} to 10^{-15} s^{-1}; e.g. Twiss and Moores 1992), and shock-pressure duration is measured in seconds, or less, in even the largest impact events (Melosh 1989). These physical conditions are not reproduced by endogenic geologic processes. They are unique to impact and, unlike endogenic terrestrial metamorphism, disequilibrium and metastability are common phenomena in shock metamorphism.

The extreme pressures and high strain rates of shock deformation are fundamental differences from normal endogenic causes of compression (Ashworth and Schneider 1985, Goltrant et al. 1991, 1992, Langenhorst 1994). A shock wave passing through a heterogeneous rock mass undergoes numerous modifications, as it interacts with grain boundaries, fractures, foliations, and different mineral species with different shock impedances within the

Figure 6. Outcrop (~ 80 m high) of coherent impact melt rock at the Mistastin complex impact structure, Canada.

rock. There is, thus, local variations in shock pressure. Petrographic study indicates that shock pressures may vary by a factor of two or more over distances ranging from millimeters to meters in outcrops (Grady 1977). Hence, each individual mineral grain experiences its own particular shock history based upon its physical properties and its relationship to both the adjacent grains and the overall structural character of the rock. A maximum shock effect in grains of a particular mineral species in a hand specimen may, thus, be a means of measuring relative deformation intensities throughout an impact structure. For example, shock pressures of at least ~ 5 GPa are required to produce PFs in quartz and greater than 10 GPa to produce PDFs in quartz or feldspars. This variation of shock deformation of important rock-forming minerals of the target rocks with increasing shock pressures have been used to delineate zones of shock metamorphism in the floor of a number of impact structures, e.g. Charlevoix, Canada (Robertson 1968), Brent, Canada (Dence 1968), Ries, Germany (von Engelhardt and Stöffler 1968), and Manicouagan, Canada (Dressler 1990), with the intensity of deformation decreasing from the center outwards.

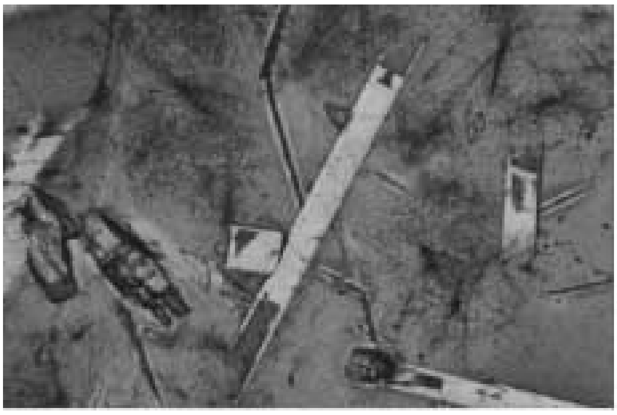
The exact physical conditions on impact are a function of the specific impact parameters. The density of the impacting body and the target, and the impact velocity determine the peak pressure on impact. The shock wave attenuates with distance from the impact point with the kinetic energy of the impact event determining the absolute radial distance in the target at which a specific shock pressure is achieved and, thus, which specific shock-metamorphic effects occur. Shock-metamorphic effects are well described in papers by Chao (1967), Bunch (1968), Stöffler (1971, 1972, 1974), Stöffler and Langenhorst (1994), Grieve et al. (1996), French (1998), Langenhorst and Deutsch (1998), and Langenhorst (this volume). They are discussed here only in general terms.

Impact melting

During compression, considerable pressure-volume work is done and the pressure release occurs adiabatically. Heating of the target rocks, thus, occurs as not all this pressure-volume work is recovered upon pressure release and results in irreversible waste heat. Above 60 GPa, the waste heat is sufficient to cause whole-rock melting and, and at higher pressures, vaporisation of a certain volume of target rocks (Melosh 1989). This volume is a function of the impact velocity, physical properties of the impacting body and target, and, most importantly, the size of the impacting body (Grieve and Cintala 1992).

Impact melt lithologies may occur as glass bombs in crater ejecta (von Engelhardt 1990), as dykes within the crater floor and walls, as glassy to crystalline pools and lenses within the breccia lenses of simple craters, or as coherent annular sheets (Fig. 6) lining the floor of complex craters and stratigraphically located immediately above breccias and/or brecciated basement rocks and overlain by breccias.

When crystallized, impact-melt sheets have igneous textures, but tend to be heavily charged with clastic debris



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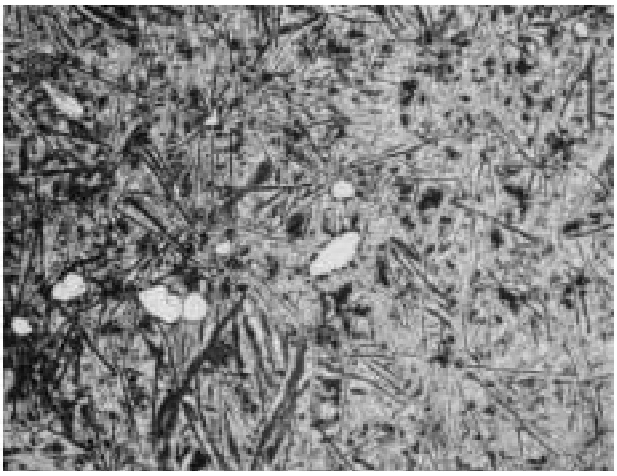

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Figure 7. Photomicrographs of far-from-equilibrium textures examples in impact melts: (a) plagioclase crystals with swallow-tail texture, Boltysh impact melt sheet, Ukraine, plane light, field-of-view = 2.28 mm; (b) pyroxene-plagioclase spherulitic texture, Vredefort Granophyre impact melt dyke, South Africa, plane light, field-of-view = 5 mm.

towards their lower and upper contacts. They may, therefore, have a textural resemblance to endogenic igneous rocks. Impact melts are superheated, reaching thousands of degrees Kelvin. Temperature differences with host rocks may result in rapid cooling of the melt leading to far-from-equilibrium textures (Fig. 7). Grain-size in thick impact-melt sheets increases inwards from the contacts, but, in general, impact-melt rocks are usually fine-grained to glassy. An important textural property of impact-melt rocks is the presence of mineral and rock fragments, which have undergone shock metamorphism of different degrees, and have been variously reworked by the melt. The size of such fragments ranges from millimeters to several hundreds of meters, and gradational changes in inclusion content are observed in thick melt sheets, varying from one to several tens of percent (e.g. von Engelhardt 1984), with highest concentrations towards their lower and upper contacts.

Impact-melt rocks can have an unusual chemistry compared with endogenic volcanic rocks, as their composition depends on the wholesale melting of a mix of target rocks, as opposed to partial melting and/or fractional crystallization relationships for endogenous igneous rocks. The composition of impact-melt rocks is characteristic of the target rocks and may be reproduced by a mixture of the various country rock types in their appropriate geological proportions. Such parameters as $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ ratios may also reflect the pre-existing target rocks within the impact-melt rocks composition (Jahn et al. 1978, Faggart et al. 1985). In general, unlike endogenous magmatic rock masses of comparable size (up to a few hundred meters thick), even relatively thick impact-melt sheets are chemically homogeneous over distances of millimeters to kilometers. In cases where the target rocks are not homogeneously distributed, this observation may not hold true, such as for Manicouagan, Canada (Grieve and Floran 1978), Chicxulub (Kettrup et al. 2000) and Popigai (Kettrup et al. 2002). Differentiation is not a characteristic of relatively thick coherent impact-melt sheets (with the exception of the extremely thick, ~ 2.5 km, Sudbury Igneous Complex, Sudbury Structure, Canada; Ostermann 1996, Ariskin et al. 1999, Therriault et al. 2002).

Enrichments above target rock levels in siderophile elements and Cr have been identified in some impact-melt rocks. These are due to an admixture of up to a few percent of meteoritic material from the impacting body. In some melt rocks, the relative abundances of the various siderophiles have constrained the composition of the impacting body to the level of meteorite class, (e.g. East Clearwater, Canada, was formed by a C1 chondrite, Palme et al. 1979). In other melt rocks, no siderophile anomaly has been identified. This may be due to the inhomogeneous distribution of meteoritic material within the impact-melt rocks and sampling variations (Palme et al. 1981) or to differentiated, and, therefore, relatively non-siderophile-enriched impacting bodies, such as basaltic achondrites. More recently, high precision osmium-isotopic analyses have been used to detect a meteoritic signature at terrestrial impact structures (e.g. Koeberl et al. 1994). Unfortunately, Re-Os systematics are, in themselves, not an effective discriminator between meteorite classes.

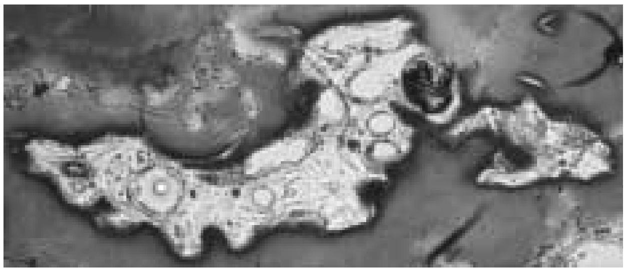

Figure 8. Photomicrograph of fused glass (lechatelierite), Ries, Germany, plane light, field-of-view = 2.5 mm.

Fused glasses and diaplectic glasses

In general, shock fused minerals are characterized morphologically by flow structures and vesiculation (Fig. 8). Peak pressures required for shock melting of single crystals are in the order of 40 to 60 GPa (Stöffler 1972, 1974), for which postshock temperatures (> 1000 °C) exceed the melting points of typical rock-forming minerals (Fig. 5). At these conditions, the minerals in the rock will melt immediately and independently after the passage of the shock wave. This melt has approximately the same composition as the original mineral before any flow or mixing takes place, and the melt regions are initially distributed through the rock in the same manner as the original mineral grains (French 1998). Melting is mineral selective, producing unusual textures in which one or more minerals show typical melting features; whereas, others, even juxtaposed ones, do not. One of the most common fused glasses observed at terrestrial impact structures is that of quartz, i.e. lechatelierite (e.g. Fig. 8).

Conversion of minerals to an isotropic, dense, glassy phase at peak pressures of 30 to 50 GPa (Fig. 5) and temperatures well below their normal melting point is a shock metamorphic effect unique to framework silicates. These phases are called diaplectic (from the Greek “destroyed by striking”) glasses, which are produced by breakdown of long-range order of the crystal lattice without fusion. Although diaplectic forms may occur as the direct result of compression by the shock wave, they are probably more commonly produced by inversion from a high-pressure crystalline phase, which is unstable in the postshock P-T environment (Robertson 1973). Based on shock recovery experiments, the formation of diaplectic glass occurs between 30 and 45 GPa for feldspar and 35 to 50 GPa for quartz (e.g. Stöffler and Hornemann 1972). The morphology of the diaplectic glass is the same as the original mineral crystal and shows no evidence of fluid textures (e.g. Grieve et al. 1996). Diaplectic glasses have densities low-

Figure 9. Photomicrograph of partial conversion to maskelynite of plagioclase feldspar crystals, Manicouagan, Canada, cross-polarized light, field-of-view = 5 mm.

er than the crystalline form from which they are derived, but higher than thermally melted glasses of equivalent composition (e.g. Stöffler and Hornemann 1972, Langenhorst and Deutsch 1994). With increasing pressure, the bulk density of diaplectic glass decreases. This decrease is due in part to progressively greater portions of the mineral having been converted to low density, disordered phases, but also to the fact that diaplectic phases exist in a sequence of intermediate structural states, whose refractive index and density decrease with increasing pressure and temperature (Stöffler and Hornemann 1972). The refractive index of diaplectic glasses is also generally higher than for synthetic, or thermally melted, glasses of equivalent composition (e.g. Robertson 1973, Grieve et al. 1996). However, in the case of K-feldspar, its diaplectic glass has a slightly lower refractive index than the fused feldspar glass (Stöffler and Hornemann 1972). Maskelynite, the diaplectic form of plagioclase (Fig. 9), is the most common example from terrestrial rocks; diaplectic glasses of quartz (Chao 1967) and of alkali feldspar (Bunch 1968) are also reported but in lesser abundance. Diaplectic glasses of different minerals can exist adjacent to one another without mixing (e.g. Robertson 1973).

High-pressure polymorphs

Shock can result in the formation of metastable polymorphs, such as stishovite and coesite from quartz (Chao et al. 1962, Langenhorst this volume) and diamond and lonsdaleite from graphite (Grieve and Masaitis 1996, Masaitis 1998, Langenhorst this volume). Coesite and diamond are also products of endogenic terrestrial geological processes, including high-grade metamorphism, but the paragenesis and, more importantly, the geological setting are completely different from that in impact events.

Under high pressure, the mineral lattice is unstable and is converted to a more stable configuration. Such transformation begins at ~ 11.5 GPa for K-feldspars (Robertson 1973) and at ~ 12 GPa for quartz (De Carli and Milton 1965). With increasing pressure, a greater proportion of the mineral is converted to a high-pressure polymorph until complete transformation is achieved at ~ 30 GPa for feldspars (Ahrens et al. 1969) and ~ 35 GPa for quartz (Stöffler and Langenhorst 1994). Neither the high-pressure phase of K-feldspar, thought to be the dense hollandite-type structure with Al and Si in octahedral co-ordination, nor an equivalent plagioclase polymorph have been recovered from shock experiments or identified in non-impact terrestrial rocks (Robertson 1973). It would appear that these phases are very unstable in postshock environments and, more likely, invert to more disordered, metastable phases. The high-pressure polymorphs of quartz (i.e. stishovite and coesite) have only rarely been produced by laboratory shock recovery experiments (cf. Stöffler and Langenhorst 1994). Contrary to what is expected from equilibrium phase diagram, stishovite is formed at lower pressures (12–30 GPa) than coesite (30–50 GPa; Stöffler and Langenhorst 1994) in impact events. This is mainly due to the fact that stishovite is formed during shock compression, whereas, coesite crystallizes during pressure release. In terrestrial impact structures, these polymorphs occur in small or trace amounts as very fine-grained aggregates and are formed by partial transformation of the host quartz. In crystalline or dense rocks, coesite is found in quartz with planar deformation features (PDFs) and strongly lowered refractive index and, more commonly, in diaplectic glass; whereas, in porous sandstone, coesite co-exists with > 80% of quartz displaying planar fractures (PFs) and diaplectic quartz glass (Grieve et al. 1996). Stishovite occurs most commonly in quartz with PDFs and less frequently in diaplectic glass (Stöffler 1971). For details on the characteristics of coesite and stishovite, the reader is referred to Stöffler and Langenhorst (1994) and references therein.

Planar microstructures

The most common documented shock-metamorphic effect is the occurrence of planar microstructures in tectosilicates, particularly quartz (Hörz 1968). The utility of planar microstructures in quartz reflects the ubiquitous nature of the mineral and its stability, including the stability of the microstructures themselves, in the terrestrial environment, and the relative ease with which they can be documented. For details, the reader is referred to the accompanying paper by Langenhorst. Recent reviews of the nature of the shock metamorphism of quartz, with an emphasis on the nature and origin or planar microstructures in experimental and natural impacts, can be found in Stöffler and Langenhorst (1994) and Grieve et al. (1996).

Planar deformation features (PDFs) in minerals are produced under pressures of ~ 10 to ~ 35 GPa (Fig. 5). Planar fractures (PFs) form under shock pressures ranging from ~ 5 GPa up to ~ 35 GPa (Stöffler 1972, Stöffler and Langenhorst 1994).

Shatter cones

The only known diagnostic shock effect that is megascopic in scale is the occurrence of shatter cones (Dietz 1968). Shatter cones are unusual, striated, and horse-tailed conical fractures ranging from millimeters to meters in length produced in rocks by the passage of a shock wave (e.g. Sagy et al. 2002). The striated surfaces of shatter cones are positive/negative features and the striations are directional, i.e., they appear to branch and radiate along the surface of the cone. The acute angle of this distinctive pattern points toward the apex of the cone and the shatter cones themselves generally point upward with their axes lying at any angle to the original bedding. Once the host rocks are graphically restored to their original impact position, shatter cones indicate the point of impact.

Shatter cones are initiated most frequently in rocks that experienced moderately low shock pressures, 2–6 GPa (Fig. 5), but have been observed in rocks that experienced ~25 GPa (Milton 1977). These conical striated fracture surfaces are best developed in fine-grained, structurally isotropic lithologies, such as carbonates and quartzites. They do occur in coarse-grained crystalline rocks but are less common and poorly developed. They are generally found as individual or composite groups of partial to complete cones (Fig. 10) in place in the rocks below the crater floor, especially in the central uplifts of complex impact structures, and rarely in isolated rock fragments in breccia units. Shatter cones are used as a diagnostic field criterion to identify impact structures (e.g. Dietz 1947, Milton 1977).

Figure 10. Complete shatter cone in limestone, Cap de la Corneille, Charlevoix, Canada.

Conclusion

The detailed study of impact events on Earth is a relatively recent addition to the spectrum of studies engaged in by the geological sciences. More than anything, it was preparations for and, ultimately, the results of the lunar and the planetary exploration program that provided the initial impetus and rationale for their study. Some recent discoveries have resulted from the occurrence or re-examination of unusual lithologies, rather than an obvious circular geological or topographic feature. For example, unusual breccias at Gardnos, Norway and Lockne, Sweden had been known for some time, but their shock-metamorphic effects were documented only recently, and they are now associated with the remnants of impact structures (French et al. 1997, Lindström and Sturkell 1992).

The level of knowledge concerning individual terrestrial impact structures is highly variable. In some cases, it is limited to the original discovery publication. In terms of understanding the terrestrial record, this is compensated, to some degree, by the fact that impact structures with similar dimensions and target rocks have the same major characteristics. Nevertheless, there is still much to be learned about impact processes from terrestrial impact structures, particularly with respect to details of the third dimension. This is the property that is unobtainable from impact structures on other bodies in the solar system, where it must be studied by remote-sensing methodologies.

Apart from increasing our understanding of impact processes, the study of terrestrial impact structures has influenced the siting of significant economic deposits (Grieve and Masaitis 1994, Donofrio 1997, Grieve 1997). In addition, the documentation of the terrestrial impact record provides a direct measure of the cratering rate on Earth and, thus, a constraint on the hazard that impact presents to human civilization (Gehrels 1994). The K/T impact may have resulted in the demise of the dinosaurs as the dominant land-life form and, thus, permitted the ascendancy of mammals and, ultimately, humans. It is, however, inevitable that human civilization, if it persists long enough, will be subjected to an impact-induced environmental crisis of potentially immense proportions.

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