Meteoritics & Planetary Science 39, Nr 2, 199–216 (2004)
Abstract available online at http://meteoritics.org

Observations at terrestrial impact structures: Their utility in constraining crater formation

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(Received 22 July 2003; revision accepted 15 December 2003)

Abstract–Hypervelocity impact involves the near instantaneous transfer of considerable energy from the impactor to a spatially limited near-surface volume of the target body. Local geology of the target area tends to be of secondary importance, and the net result is that impacts of similar size on a given planetary body produce similar results. This is the essence of the utility of observations at impact craters, particularly terrestrial craters, in constraining impact processes. Unfortunately, there are few well-documented results from systematic contemporaneous campaigns to characterize specific terrestrial impact structures with the full spectrum of geoscientific tools available at the time. Nevertheless, observations of the terrestrial impact record have contributed substantially to fundamental properties of impact. There is a beginning of convergence and mutual testing of observations at terrestrial impact structures and the results of modeling, in particular from recent hydrocode models. The terrestrial impact record provides few constraints on models of ejecta processes beyond a confirmation of the involvement of the local substrate in ejecta lithologies and shows that Z-models are, at best, first order approximations. Observational evidence to date suggests that the formation of interior rings is an extension of the structural uplift process that occurs at smaller complex impact structures. There are, however, major observational gaps and cases, e.g., Vredefort, where current observations and hydrocode models are apparently inconsistent. It is, perhaps, time that the impact community as a whole considers documenting the existing observational and modeling knowledge gaps that are required to be filled to make the intellectual breakthroughs equivalent to those of the 1970s and 1980s, which were fueled by observations at terrestrial impact structures. Filling these knowledge gaps would likely be centered on the later stages of formation of complex and ring structures and on ejecta.

INTRODUCTION

Natural impact craters are the result of the hypervelocity impact of an asteroid or comet with a planetary surface. Impact is now recognized as a ubiquitous geologic process affecting all the terrestrial planets. Impact involves the virtually instantaneous transfer of the considerable kinetic energy in the impacting body to a spatially limited, near-surface volume of a planet's surface, where it is partitioned into kinetic energy (leading to a craterform) and internal energy (leading to shock metamorphism) in the target area. As a consequence of the nature of the impact process, local geology of the target area is generally of secondary importance to the final results of the impact process. The net result is that impacts of similar size on a given planetary body tend to produce similar results. This property is at the center of the utility of observations at impact craters, particularly terrestrial impact craters, to provide information on cratering processes.

The Earth is the most geologically active of the terrestrial planets and, given that impact craters are surficial features, has retained the poorest sample of the record of hypervelocity impact throughout geologic time. The study of terrestrial impact craters also does not have a long-established tradition in the geosciences (the first terrestrial craterform ascribed to impact was by Barringer (1906), and most detailed studies of terrestrial impact craters have occurred since the mid 1960s). Until recently, the provenance of such studies has been a small number of workers with strong ties to planetary geoscience. Nevertheless, although the known sample of terrestrial impact structures is small, the terrestrial impact record has a major role in understanding and constraining cratering processes. A listing of known terrestrial impact structures (~170) and some of their salient

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characteristics can be found at http://www.unb.ca/passc/ImpactDatabase/index.html.

The results of hypervelocity impact cratering are not amenable to exact duplication by experiment because of problems of scale and experimental impact velocities that are slower than those in natural hypervelocity impacts. The terrestrial record is, therefore, the major source of ground-truth data on the geological and geophysical results of hypervelocity impact at a variety of scales from hundreds of m to hundreds of km. The terrestrial record is of particular utility in terms of the three-dimensional structural and lithological character of impact craters. These three-dimensional characteristics have been determined directly by observation through drilling or the characterization of similar-sized structures exposed to different erosional levels and indirectly by the interpretation of geophysical data.

The terrestrial impact record as a whole contains a number of biases, reflecting modification and obliteration of terrestrial impact craters by post-impact endogenic geologic processes. As terrestrial processes, such as erosion and/or sedimentation, have modified the form of terrestrial impact craters, some no longer correspond to the definition of a crater, i.e., a negative topographic feature. Therefore, we use the term terrestrial impact structure to describe terrestrial impact craters, which is more generic and has no specific inference as to current topographic appearance. The known record is biased toward larger and geologically younger impact structures occurring on geologically stable areas, such as cratons. The knowledge base at any given terrestrial impact structure is highly variable and, in most cases, contains observational gaps. With a number of exceptions (e.g., Masaitis et al. 1975; Grieve 1988), there have been few systematic campaigns designed to study specific impact structures with the full spectrum of geoscientific tools available at the time of the campaign. In most cases, the knowledge base has been acquired incrementally over time and reflects the current understanding of impact processes at the time of specific observations. While one would wish observations to be unbiased and objective, the reality is that both observations and the interpretations of these observations concentrate on what is recognizable and generally known. Consider, for example:

1. The recognition of shatter cones at Sudbury, Canada. They were recognized only after they were predicted to occur by Dietz (1962, 1964). This was in spite of the fact that they are abundant at Sudbury, which had been the site of intensive geological exploration for many decades, because of the world-class Cu-Ni ore deposits associated with the Sudbury Igneous Complex (SIC), now generally interpreted as a differentiated impact melt sheet (e.g., Therriault et al. 2002). In this case, the observation was not made because it was not looked for or recognized as being significant.
2. The recognition of a geochemical anomaly and unusual features in quartz grains at the Cretaceous/Tertiary (K/T) boundary in the 1970s (Christensen et al. 1973). It was almost a decade until more sophisticated geochemical analyses (Alvarez et al. 1980) linked this anomalous geochemistry to an impact origin. These more recent analyses, however, were performed in the context of geochemical anomalies occurring in impact lithologies. It was several more years until scanning electron microscope (SEM) images of planar deformation features (PDFs) in quartz from K/T boundary deposits were presented (Bohor et al. 1984). In this case, the interpretation of the significance of the initial observations of anomalous chemical and physical characteristics was not recognized due to a lack of context at the time.

With the above caveats as an overarching context, this work describes some of the salient observations at impact structures that have, historically and as the knowledge base has expanded, provided constraints on crater formation and models thereof. The features observed at impact structures reflect the result of the time integration of all stages of impact phenomena, with a direct observational bias to later times in cratering processes. Nevertheless, by inference, deduction, and comparison with experimental and model impacts, the observed record can produce constraints on earlier times in the cratering process. As the final form of impact structures is scale-dependent and becomes more complex with size, we develop what constraints have been provided from simple to complex impact structures (Dence 1968, 1972) to impact basins. The latter are defined here as a craterform with one or more topographic ring structures interior to the main topographic rim (Spudis 1993).

SIMPLE IMPACT STRUCTURES

Form: Not That Simple

Simple craters have the form of a bowl-shaped depression, with a structurally uplifted rim, which includes an overturned flap and ejecta. The classic example in the terrestrial environment is Meteor or Barringer Crater, USA (Fig. 1; Shoemaker 1963). At most terrestrial simple impact structures, however, the ejecta and overturned flap have been removed by erosion, and the structures associated with the uplifted rim are not as easily defined as in the detailed, horizontal, and preserved stratigraphy at Barringer. Historically, symmetric residual gravity lows over the apparent floor of terrestrial simple impact structures indicated reduced densities, suggesting brecciation beneath the apparent floor. This was confirmed by drilling at several simple craters, which recovered cores of breccia, e.g., Holleford, West Hawk, and Brent, Canada (Beals 1960; Halliday and Griffin 1967; Dence 1968) and Lonar, India (Fredriksson et al. 1973).

Fig. 1. Oblique aerial photograph of the classical example of a terrestrial simple impact structure, the 1.2 km-diameter Barringer or Meteor Crater, Arizona, USA.

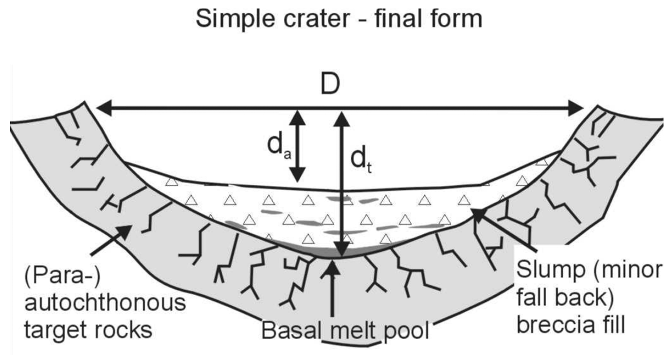

Fig. 2. Schematic cross-section of a terrestrial simple impact structure, indicating various lithological, morphometric, and structural attributes. D = rim diameter; d_a = apparent depth; d_t = true depth. (See text for morphometric relations.)

This crater-filling breccia is polymict and is not, in general, highly shocked, but it does contain examples and zones of shocked material, including impact melt glasses and rocks. All these observations are consistent with the breccia beneath the apparent floor of simple craters being allochthonous. Gravity modelling and extended and deeper drilling defined this breccia to be lensoid in cross-section and contained within shocked and fractured parautochthonous wall and floor target rocks. This structure defined by the parautochthonous target rocks is also roughly parabolic in cross-section and is known under the designation of a true crater. Thus, the concept of apparent and true craters (Fig. 2) arose out of observations at simple terrestrial impact structures. Early interpretations ascribed this breccia in-filling of the true crater to some form of fallback material from the "explosion" resulting from the impact (cf., Nordyke 1961).

Constraints on the cratering process provided by morphometric data from simple impact structures is generally good, as morphometric data are surface data and can be supplied from observations of impact structures on other terrestrial planets, particularly the Moon (e.g., Pike 1977). On Earth, reliable morphometric data, however, are limited to some seven simple impact structures (i.e., Barringer, Brent, Lonar, West Hawk, Aouelloul and Tenoumer, Mauritania, and Wolfe Creek, Australia). They define the empirical relationships:

$$d_a = 0.13\ D^{1.06} \text{ and } d_t = 0.28\ D^{1.02}$$

where d_a is the depth of the apparent crater, d_t is the depth of the true crater, D is the rim diameter of the structure, and the units, here and in all other relationships cited in the text, are in km (Fig. 2; Grieve and Pilkington 1996). Although there are relatively few data, there is no obvious difference in these relations as a function of target rock type, i.e., crystalline or sedimentary. The form of the apparent depth relationship is similar to that for the Moon, where $d_a = 0.196\ D^{1.01}$ (Pike 1977).

Formation: The Concept of a Transient Cavity

From the drilling campaigns at Brent and other simple impact structures, the depth to the base of the true crater was determined to be approximately one-third of the rim diameter or approximately twice the depth of the apparent crater (Fig. 2; Grieve and Garvin 1984). This three-dimensional picture was in contrast to that based on impact experiments into sand (e.g., Stöffler et al. 1975). Such experimental craters lacked an in-filling breccia lens and, thus, an apparent crater. These experimental craters, however, did duplicate the overturned flap in the rim area, the rim uplift, and also indicated that the true crater was formed partially by the ejection and partially by the displacement of target materials. These results, in turn, introduced concepts concerning the cratering "flow-field," i.e., the downward and outward movement of target materials set in motion by the shock and rarefaction waves (Maxwell 1977), and turned attention away from earlier concepts of "explosions" during crater formation.

While the observation that simple craters were partially filled by polymict, allochthonous breccia was duplicated at a number of simple impact structures, it was largely the drilling campaign at Brent that led Dence (1968) to introduce the concept of the "transient cavity." He reasoned that, if the interior breccia lens generally displayed little or no shock metamorphic features, it was not a fallback deposit. Therefore, it had to have been generated within the simple crater itself. That is, the cavity, formed through excavation and displacement induced by the cratering flow-field, was unstable at larger natural simple craters, compared to the experimental craters, which were orders of magnitude smaller. The fractured and weakened parautochthonous wall rocks collapsed inward in the later stages of the cratering event and resulted in the formation of the interior breccia lens (Fig. 2). This is consistent with the polymict character and general low or lack of shock in the interior breccia lens (the fewer, more highly shocked materials in the breccia lens represent material set in motion but that had failed to be ejected and were lining the cavity walls at the time they collapsed). Thus, the fundamental concept of the transient cavity, formed directly by the cratering flow-field and subsequently modified by transient cavity wall failure and collapse to produce the final simple craterform, was born.

The concept of transient cavity and its subsequent modification is a cornerstone in the knowledge and understanding of simple crater formation from an impact event. The variance or inconsistency between natural (km-scale) observations and observations at small-scale (cm) experimental craters characterized an observational gap that drove the advance in the conceptual understanding of simple crater formation. The concept of the formation of a transient cavity and its subsequent collapse by wall failure is now the generally accepted working hypothesis for simple impact structures. There is some additional observational evidence in support of the working hypothesis. For example, there is evidence of an over heightened transient cavity rim area created directly by the flow-field (thrusts) and its subsequent inward collapse (faults) at Barringer (Shoemaker 1974; Roddy et al. 1975; Pilon et al. 1991). There is also evidence that uplift, due to the flow-field, extends to a distance of approximately one crater radius outside the rim (Pilon et al. 1991).

Given its importance, however, it is surprising how little scrutiny the working hypothesis of the transient cavity and its collapse at simple impact structures has undergone. A geometric model for wall slumping was developed and combined with a Z-model (Maxwell 1977) to simulate the excavation and displacement flow-field within the transient cavity (Grieve and Garvin 1984). This analytical model was tested using drilling data from Barringer, Brent, Lonar, and West Hawk combined with 3D interpretations based on gravity models at Aouelloul, Tenoumer, and Wolfe Creek (Grieve et al. 1989). Although some graphic information from drilling is available from other structures, notably in the former Soviet Union (e.g., Masaitis et al. 1980), textual or tabular information on their dimensions is generally not available. Thus, testing of the validity of the concept of transient cavity wall collapse at simple craters is based on data and interpretations at only seven simple impact structures (approximately 15% of the currently known simple impact structures in the terrestrial record), and three of these are non-unique, although consistent, interpretations of depths from gravity data. There is a clear need for additional accurate three-dimensional data for simple impact structures in the terrestrial record.

Shock Attenuation: A Compressed Section

Shock metamorphic effects in the parautochthonous target rocks at terrestrial simple impact structures are confined to the base of the true crater (Fig. 2). To our knowledge, the systematic study of recorded shock attenuation with depth at simple impact structures is limited to Brent. Here, the highest shock pressure recorded in the parautochthonous target rocks is ~23 GPa, based on the orientations of PDFs in quartz (Robertson and Grieve 1977), although recently, Dence (2002) has argued that the samples measured for recorded shock pressure may be allochthonous and not parautochthonous. Some of the shock record close to the immediate area of the true crater floor has been lost due to recrystallization of quartz by thermal metamorphism from a small coherent impact melt body near the base of the in-filling breccia lens. Thus, ~23 GPa is a minimum for the recorded shock pressures at the base of the true crater, which is generally equated, in terms of depth, with the base of the transient cavity in simple impact structures (Grieve and Garvin 1984). The axial shock attenuation rate observed in the true crater floor of Brent follows the relationship $P \propto R^{-20}$, where P is recorded shock pressure, and R is radial distance normalized to the radius of the transient cavity (Robertson and Grieve 1977). This extremely rapid shock attenuation rate can be rationalized to lower and more reasonable rates, comparable to those in model calculations (e.g., Ahrens and O'Keefe 1977; Dence 2002), if the section of parautochthonous rocks of the true crater floor is expanded to its precompression thickness before displacement in transient cavity formation, based on observed net shortening of sections observed at experimental and nuclear craters, such as Piledriver (Dence et al. 1977). This reasoning is based on the transient cavity at simple impact structures being formed partially by ejection and partially by displacement, with these displacements being essentially frozen into the true crater floor of the final structure. It is, however, a somewhat circular argument, and the data are from only a single simple impact structure (Brent).

Ejecta: Only One Datum

The terrestrial record provides very little constraint on the depth of excavation at simple impact structures. Again, there is currently only a single data source. In this case, it is Barringer. The deepest lithology recorded in the exterior ejecta blanket of Barringer is Coconino sandstone, which occurs at depths of ~90–310 m in the target rocks (Shoemaker 1963). These depths are consistent with the general Z-model but poorly constrain the exact testing of the specifics of maximum ejection depth. Ejecta are known from a few other simple impact structures (e.g., Tabun-Kara-Obo, Mongolia; Masaitis 1999), but these examples provide no precise constraints of the depth of origin of ejecta.

COMPLEX IMPACT STRUCTURES

Form: Not All Have Central Peaks

Complex impact structures (Dence 1968) are highly modified craterforms (Fig. 3), where the transient cavity defined by simple impact structures has undergone considerably more modification in achieving the final crater form. While this statement may sound trite and self-evident today, this was not always the case. It was observations at terrestrial impact structures that again played a major role in developing the concepts of the formation processes of complex impact structures. It was also a lack of understanding of these processes that plagued the peer-reviewed literature with challenges to the impact origin of some complex impact structures (e.g., Bucher 1963) and has continued to do so until relatively recent times (e.g., Nicolaysen and Ferguson 1990; Reimold 1990; Bridges 1997).

Morphologically, complex impact structures consist of a structurally complex rim area, a topographically lower annular trough, and a central, relatively elevated, topographic peak (Fig. 4). These basic morphological elements are, generally, common to complex impact structures on the terrestrial planets. The rocks present at the center of terrestrial complex impact structures have been uplifted from depth and display increasing depth of origin and age as the center is approached (e.g., Gosses Bluff, Australia [Milton et al. 1972]; Red Wing, USA [Brenan et al. 1975]; Sierra Madera, USA [Wilshire and Howard 1968]). The observed amount of structural uplift (SU; Fig. 4) undergone by the deepest unit exposed at the center of terrestrial complex impact structures defines the empirical relationship:

$$SU = 0.06\ D^{1.1}$$

(Grieve et al. 1981). This was based on data from some 15 terrestrial complex impact structures and was later revised to:

$$SU = 0.086\ D^{1.03}$$

based on empirical data from a total of 24 structures (Fig. 5; Grieve and Pilkington 1996). An independent estimate by Ivanov et al. (1982) gives the amount of structural uplift as:

$$SU = 0.1\ D.$$

According to these estimates, a good working hypothesis is that the observed structural uplift is approximately one-tenth of the rim diameter at terrestrial complex impact structures. The greatest uncertainty in defining these relationships, because of erosion, is the estimate of the rim diameter. An additional complication is that different observational data sets (e.g., surface geology, drill core, and geophysics) can produce slightly different estimates of rim diameters at a given complex impact structure, due to variations in spatial resolution and the specific physical element being observed (e.g., Grieve 1988).

Compared to simple impact structures, there are even fewer data available to define the morphometry of terrestrial complex impact structures. Grieve and Pesonen (1992) derived the empirical relations:

$$d_a = 0.12\ D^{0.30} \text{ and } d_a = 0.15\ D^{0.43}$$

for sedimentary and crystalline targets, respectively. These have considerable uncertainty, as they are based on only five

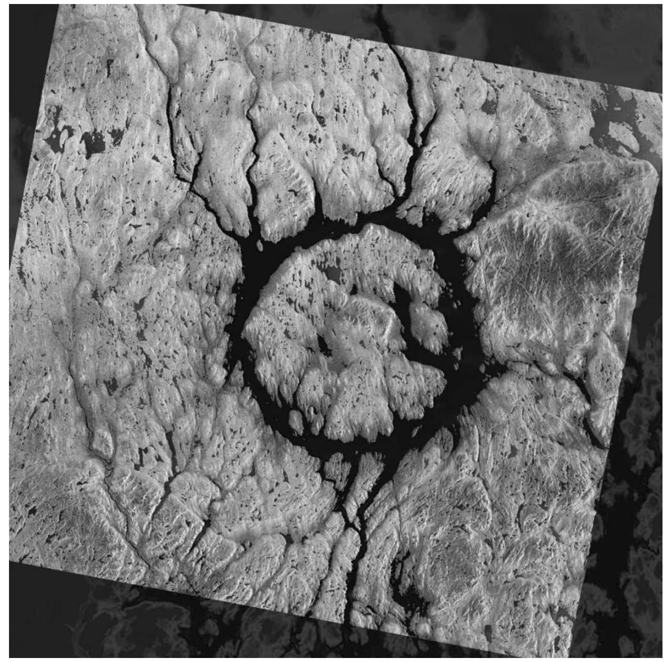

Fig. 3. Digital elevation model (DEM) of a complex impact structure, Manicouagan, Canada. The lake fills an annular valley and is, in part, man-made. The lake has a diameter of ~65 km. The original rim diameter of Manicouagan is estimated to be ~100 km. The lake is at an elevation of ~400 m, and the central peak (Mont de Babel) rises to ~1000 m.

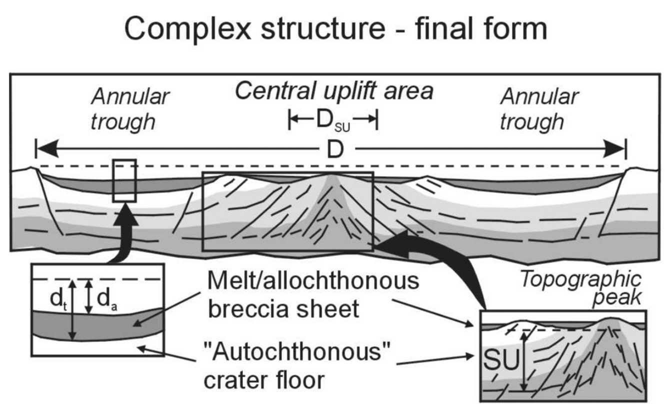

Fig. 4. Schematic cross-section of a terrestrial complex impact structure, indicating various lithological, morphometric, and structural attributes. D = rim diameter; d_a = apparent depth; d_t = true depth; D_{SU} = diameter stratigraphic uplift; SU = stratigraphic uplift. (See text for morphometric relations.)

impact structures in each target type. The general form of the relationships, however, is similar to that derived from lunar data, where $d_a \propto D^{0.3}$ was derived for complex impact structures (Pike 1977). Similar values have been derived for complex impact structures on Mercury and Venus (e.g., McKinnon et al. 1997), while the relationship on Mars is less clear, with considerable dispersion in depth estimates of "fresh" (Barlow 2000) martian complex impact structures (Whitehead et al. 2002).

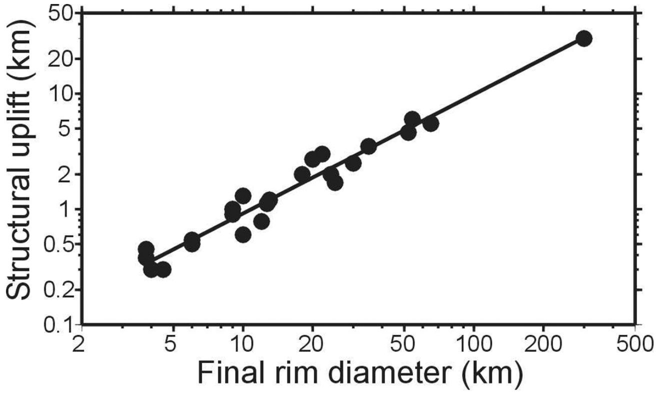

Fig. 5. Log-Log diagram of the amount of structural uplift (SU) versus final rim diameter (D) of 24 terrestrial impact structures (Grieve and Pilkington 1996). See text for definition and equation of regression line.

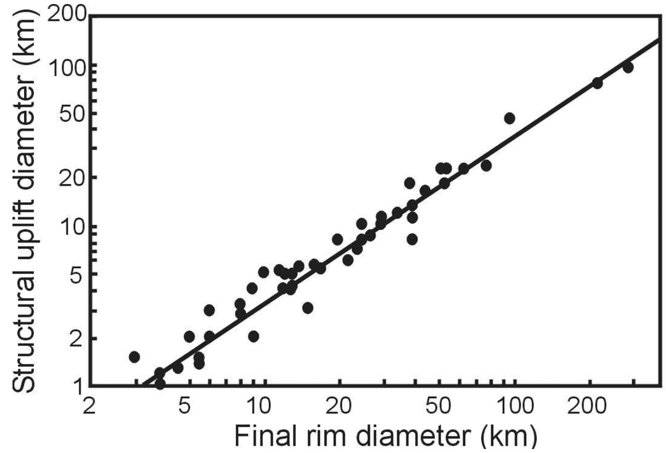

Fig. 6. Log-Log relationship between structural uplift diameter (D_{SU}) and final rim diameter (D) for 44 terrestrial impact structures (Therriault et al. 1997). See text for definition and equation of regression line.

Morphometric data on the diameter of the structural uplift (D_{SU}), as measured at its base within the stratigraphic column (Fig. 4), from 44 terrestrial complex impact structures allowed Therriault et al. (1997) to develop the empirical relation (Fig. 6):

$$D_{SU} = 0.31\ D^{1.02}.$$

Data on the physical height of the central topographic peak at terrestrial complex impact structures formed by structural uplift exhibit too much scatter, due to erosional effects, to be meaningful. This is not the case for lunar complex impact structures, where empirical relations exist for both physical height and diameter of the central peak, as expressed at the present ground surface (Pike 1977). Although genetically related, it is important to note that the morphological element of the diameter of the central peak in lunar craters is not equivalent to the morphological element of the diameter of the structural uplift at terrestrial complex impact structures.

Not all terrestrial complex impact structures have a central topographic peak. There are no central peaks at Haughton (Canada) (Fig. 7), Ries (Germany), or Zhamanshin (Kazakhstan). These are all relatively young (23 ± 1 Ma [Jessberger 1988], 14.5 ± 0.2 Ma [Schwarz and Lippolt 2002], and 1.0 ± 0.1 Ma [Dieno and Becker 1990], respectively), uneroded impact structures in the same size range (~10–25 km). The question is: Why do these complex impact structures not have central peaks? These structures were all formed in mixed targets with relatively thick sedimentary sections (1.7 km [Robertson and Sweeney 1983], ~470–820 m [Pohl et al. 1977], and ~300 m [Masaitis 1999], respectively) overlying crystalline basement. From observations of impact structures obviously formed by oblique impacts on the other terrestrial planets (e.g., Schultz 1992; Ekholm and Melosh 2001), it cannot be argued that the lack of a central peak is due to oblique impacts.

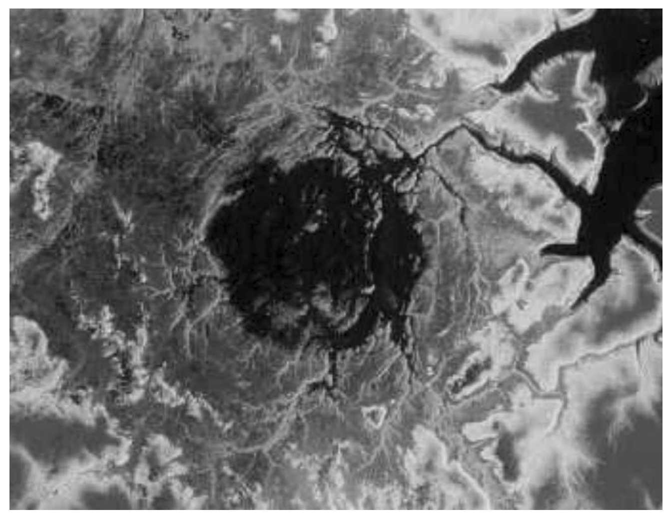

Fig. 7. DEM image of the complex impact structure Haughton (Canada) indicating no central peak. There are ~350 m of elevation difference within the impact structure.

It could be argued that these craterforms without central peaks represent the (near) pristine form of terrestrial complex impact structures and that central peaks are actually progressively exposed by erosion of crater-fill products in the terrestrial environment. The similar-sized (~25 km) Boltysh impact structure (Ukraine) has a central topographic peak, which is physically emergent from the impact lithologies within the structure. Boltysh (65.2 ± 0.6 Ma) is a buried structure and is relatively well-preserved along with some of its ejecta deposits (e.g., Masaitis 1999). Thus, erosion is not a serious contributing factor with respect to its present form. Boltysh, however, was formed in a crystalline target. By elimination, therefore, it would appear that the lack of central peaks at Haughton, Ries, and Zhamanshin is most likely an effect of target material. It is not, however, simply due to mixed versus crystalline targets.There are terrestrial complex impact structures in mixed targets that do have central topographic peaks, e.g., Puchezh-Katunki (Russia; Ivanov 1994), Obolon (Ukraine), and Logoisk (Belarus; Masaitis 1999). In these cases, however, the thickness of the sedimentary section is relatively less with respect to rim diameter (e.g., 2 km for the 80 km-diameter Puchezh-Katunki structure) than at Haughton, Ries, and Zhamanshin. It would appear, therefore, the lack of a central topographic peak is a more complex (but yet unknown) function of target and impact characteristics.

Formation: The Concept of Structural Uplift

While complex impact structures were being characterized from the 1930s through the 1970s (e.g., Boon and Albritton 1937; Stearns et al. 1968), there was little consensus as to their formation. Much of the literature was descriptive (e.g., Masaitis et al. 1980), but it did not deal in detail with formational processes. Even in the late 1970s, the formation of complex impact structures was being attributed to specific types of impacts, such as those of relatively low-density comets. In part, this can be traced to the morphological comparison of complex impact structures, such as Flynn Creek, USA, with the results of surface explosions of TNT on relatively unconsolidated media (e.g., Roddy 1977). Attempts to computationally model the formation of complex impact structures by the impact of low-density bodies generally failed but continued into the 1980s (e.g., O'Keefe and Ahrens 1980).

Most of the debate or uncertainty revolved around the depth of the original cavity formed by the cratering flow-field. Some researchers called for so-called proportional growth, i.e., a transient cavity conceptually similar in shape to that at simple impact structures but scaled up in size (Dence et al. 1977), while others called for so-called non-proportional growth, i.e., a transient cavity that was considerably shallower with respect to its diameter than that at simple impact structures (Head et al. 1975). The latter group was largely influenced again by morphology, in particular, by photogeologic interpretations of lunar impact structures. That is, their observations of impact structures were focused on the surface features and lacked geologic ground-truth information, particularly in the third dimension.

Observations at a number of terrestrial complex impact structures, particularly those in Canada (e.g., Dence et al. 1977) and in the former Soviet Union (e.g., see review by Masaitis [1999]) indicated that, in mixed targets of sedimentary cover rocks overlying crystalline basement, the sedimentary cover rocks were preserved in the annular troughs. In some cases, it could be demonstrated that these preserved cover rocks were also present in interior allochthonous breccia deposits, for example, at Puchezh-Katunki (Masaitis 1999). This observation indicated that the near-surface cover rocks were intimately involved in and affected by the cratering flow-field but only in the central portion of complex impact structures. Thus, by analogy with simple impact structures, the cratering flow-field was restricted to the central portion of complex impact structures. From a compilation of observations of the preservation of such cover rocks in annular troughs, Grieve et al. (1981) derived an empirical relationship that indicated that, at terrestrial complex impact structures, the near-surface target rocks directly affected by the cratering flow-field were limited to pre-impact locations within a diameter of <0.5–0.65 of the final rim diameter. Other empirical relations have been suggested, with, for example, 0.57 ± 0.03 as the factor (Lakomy 1990). (A number of other empirical relations for specific geologic features at terrestrial complex impact structures can be found in Stöffler et al. [1988].)

By analogy with simple impact structures, the equivalent of the transient cavity at complex impact structures was, thus, limited in horizontal extent with respect to the final rim diameter. This indicated that additional processes were occurring at complex impact structures compared to simple impact structures. It also did not necessitate the abandonment of the overarching concept of a transient cavity at complex impact structures. Observations at terrestrial complex impact structures favored deeper transient cavities not unlike those conceived for simple impact structures (e.g., Dence et al. 1977; Offield and Pohn 1977). This was based, in general, on observations of the third (depth) dimension of complex impact structures from stratigraphic reconstructions, reflection seismic data, and drilling results that indicated that progressively deeper strata have been removed by the cratering flow-field, as the center of the complex impact structure is approached, suggesting a deepening transient cavity.

The target rocks originally above the structural uplift in the center of complex impact structures have been physically removed, in the impact event, through excavation by the cratering flow-field. The target rocks in the structural uplift itself, however, were not excavated but were displaced by the cratering flow-field. Displacement was initially downward and outward during transient cavity formation and then upward and inward during transient cavity modification. In sedimentary targets, details of overthrusting and bed duplication during the formation of the structural uplift are evident (e.g., Red Wing [Brenan et al. 1975]; Gosses Bluff [Milton et al. 1996]; Cloud Creek, USA [Stone and Therriault 2003]), as the displaced target rocks are forced into a smaller volume during uplift. Convergent inward motion is also manifested through recently recognized so-called radial transpression ridges (Kenkmann and von Dalwigk 2000). Presumably, this volume problem is the reason that the structural uplift is generally manifested at the surface by a topographic central peak.

As noted earlier, from the empirical relationship, this structurally uplifted material came from a maximum depth of approximately one-tenth of the final rim diameter at complex impact structures, while the transient cavity diameter is ~0.5–

0.65 of the modified final rim diameter. This places the deepest, non-excavated material of the structural uplift at an original depth of 1/5 to 1/6 of the estimated diameter of the transient cavity. This depth is similar to the depth of origin of the non-excavated but displaced target rocks of the transient cavity at simple craters. Thus, there is observational support, at least to a first order, for the hypothesis of proportional growth and the conclusion that transient cavity at complex impact structures is similar in geometry to that at simple impact structures.

Other observations based on stratigraphic reconstructions in sedimentary target rocks support the conclusions regarding transient cavities at complex terrestrial impact structures. For example, the interior allochthonous deposits at Haughton contain shocked clasts of crystalline basement (Metzler et al. 1988). As noted earlier, Haughton was formed in a mixed target with crystalline rock occurring at depths of approximately 1.7 km (Robertson and Sweeney 1983). The crystalline clasts are in their present surface location due to mobilization by the cratering flow-field within the transient cavity. Based on reflection seismic data, Scott and Hajnal (1988) estimated a diameter of excavation of ~10 km at Haughton. This would give a depth-diameter ratio of 1/6 for the portion of the transient cavity that was due to excavation. This ratio is similar to that estimated for simple impact structures, based on Z-model type calculations (Maxwell 1977; Grieve et al. 1981).

Structural uplift cannot be directly measured at terrestrial complex impact structures in crystalline targets. Shock metamorphic effects recorded in complex terrestrial impact structures in crystalline targets, however, are spatially confined to the central, structurally uplifted area, and the level of recorded shock decreases radially outward (e.g., Robertson 1975) and downward (e.g., Stöffler et al. 1988). The peak recorded pressures are on the order of 25–30 GPa for smaller (~20 km) complex impact structures, increasing to ~45 GPa for larger complex impact structures (Grieve and Cintala 1992). As the rate of shock wave attenuation is not a function of the size of the impact event, the similar values of the peak recorded shock pressures at the smaller complex impact structures and at the bases of the true crater at simple impact structures (for which there are two data points: ~25 GPa for Rotmistrovka, Ukraine [Grieve and Cintala 1992] and 23 GPa for Brent [Robertson and Grieve 1977]) are again consistent with similar transient cavity geometries. Basilevsky et al. (1983) estimated the depth of origin of the structural uplift for Kara (Russia) at 5.7 km, based on recorded shock metamorphic effects. Kara is 65 km in diameter (Masaitis 1999), which gives a structural uplift relation (and transient cavity depth constraint) similar to that in sedimentary targets. Some additional depth estimates for the operation of cratering flow-field at complex terrestrial impact structures in mixed targets can be found in Grieve et al. (1981).

Hydrocode models of transient cavity collapse in the formation of complex impact structures require that target rock strength be reduced (e.g., Melosh and Ivanov 1999). Suggested mechanisms include acoustic fluidization (Melosh 1979) and weakening of the target rocks by shock heating (O'Keefe and Ahrens 1993). Drill cores at terrestrial complex impact structures in sedimentary targets contain evidence of thrusting and faulting in the structural uplift (e.g., Brenan et al. 1975; Offield and Pohn 1977), indicating that the target material was behaving as blocks, at least during uplift. As noted earlier, mapping of the structural uplift at terrestrial complex impact structures in sedimentary targets indicates a general bull's-eye pattern with target rocks decreasing in age radially outward and maintaining, in general, their pre-impact stratigraphic relations. In areas of excellent outcrop, such as in arid or desert environments, detailed mapping indicates that, although this general pre-impact stratigraphic integrity is maintained, lithological relations are complicated, in detail, by fault boundaries (e.g., Gosses Bluff [Fig. 8] and Sierra Madera and Upheaval Dome, USA). Reflection seismic data over structural uplifts indicate a loss of coherent reflections and a reduction in seismic velocity (e.g., Scott and Hajnal 1988; Milton et al. 1996). This signature is attributed to the occurrence of discrete blocks in the structural uplift. If it can be inferred that movement was also as discrete blocks, this characteristic of the structural uplifts at terrestrial complex impact structures in sedimentary targets can be considered generally consistent with models of acoustic fluidization to account for reduced strength during modification.

The perception that the structural uplift behaved as some form of relatively coherent mass or series of large blocks comes mostly from terrestrial complex impact structures in crystalline targets, particularly from less eroded examples with relatively poor rock exposure over the structural uplift (e.g., Manicouagan, Canada). At more deeply eroded structures, where the structural uplift is better observed, and, particularly, in areas of good exposure (e.g., the coastline of Slate Islands, Canada), breccia dikes separating discrete blocks are ubiquitous in the structural uplift (Grieve and Robertson 1976; Dressler et al. 1998). Similarly, in drill cores, where sampling of the structural uplift is (close to) 100%, such as at Puchezh-Katunki (Ivanov et al. 1996), Manicouagan, and West Clearwater (Canada; Geological Survey of Canada, unpublished data), blocks in the tens to hundreds of m size-range are apparent. Veining by pseudotachylite is also apparent within structural uplifts (e.g., Vredefort [South Africa] is the type of site for pseudotachylite; Shand 1916; Reimold and Colliston 1994).

Shock Metamorphism: An Extended Section

Data presented by Ivanov (1994) for the axial variation in recorded shock effects with depth in the structural uplift of the 80 km-diameter Puchezh-Katunki structure correspond to $P \propto R^{-1.7}$. Using two data points of shock pressure estimates from PDF orientations, Whitehead et al. (2003) calculated an axial

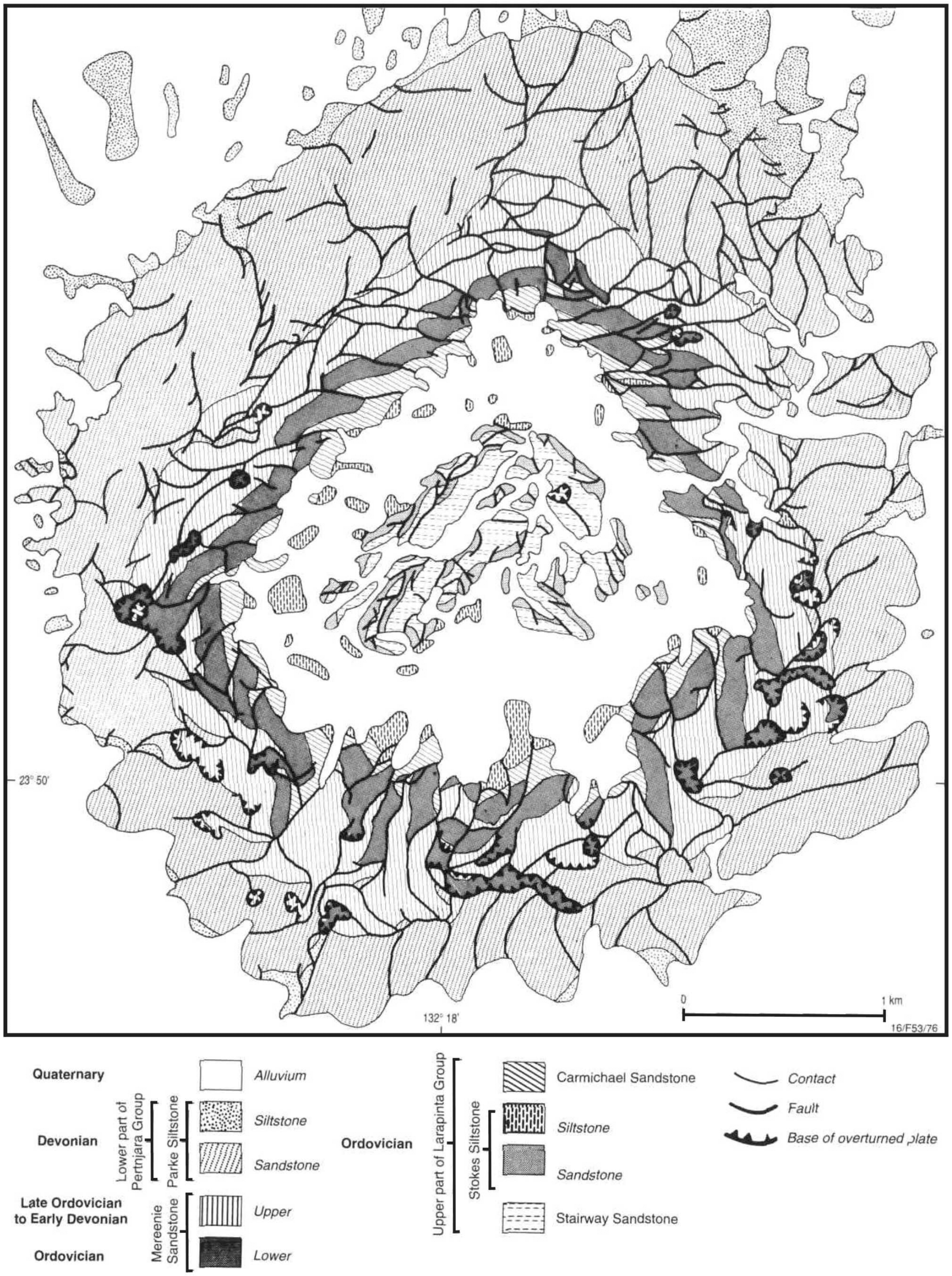

Fig. 8. Simplified geologic map of the central ring of Gosses Bluff (Australia) showing intricate fault boundaries within the structural uplift (Milton et al. 1996).

decay rate with depth of $P \propto R^{-2.2}$ from a drill core from Woodleigh, Australia. In contrast, Reimold et al. (2003) did not detect a systematic decrease in recorded shock pressure, based on PDF orientations, over ~125 m of a drill core from Woodleigh. They, however, used a non-standard method for obtaining an estimate of the recorded shock pressure. Attempts to obtain an attenuation rate from a drill core at West Clearwater also failed to define a systematic decay because of reversals in the value of recorded pressure due to overthrusting within blocks in the structural uplift (Geological Survey of

Canada, unpublished data). Basilevsky et al. (1983) obtained a value of $P \propto R^{-0.2}$ from observations of PDF orientations in quartz in core from the structural uplift of Kara.

Attempts to systematically map the radial attenuation of the surface recorded shock at terrestrial complex impact structures are equally limited. Robertson and Grieve (1977) presented data for Slate Islands and Charlevoix, Canada. They then attempted to reconstruct the transient cavity and the original depth position of the structurally uplifted rocks now at the surface. From the reconstruction, they derived an axial decay rate with depth of $P \propto R^{-4.5}$. There is, however, a computational error in the least squares regression fit to the data, and the rate is actually closer to $P \propto R^{-0.4}$ (J. Whitehead 2003, personal communication). Dressler et al. (1998) have also challenged the conclusions with respect to the Slate Islands, arguing that there is little evidence for systematic radial attenuation at Slate Islands. They, however, did not use the same methodology as Robertson and Grieve (1977), and the results may not be directly comparable.

The net result is a somewhat confusing variation in the apparent rate of axial shock attenuation at terrestrial complex impact structures. While some of the variation may be due to erosional effects and different levels of exposure, the common factor is that the attenuation rates appear to be low compared to model calculations (e.g., Ahrens and O'Keefe 1977). This can be explained by the fact that these rates reflect observations made on rocks from the structural uplift in their present position, that is, after they have been first driven down by the cratering flow-field and then uplifted in transient cavity modification. Given that observations at terrestrial complex impact structures in sedimentary targets directly demonstrate strata duplication (e.g., repetitions of Mississippian beds at Red Wing has led to a 820 m oil column in the structural uplift compared to a 30 m oil column in the surrounding area; Brenan et al. 1975), the section over which these rates have been calculated is a thickened or expanded section in terms of original length or depth.

Ejecta: Only One Good Example

As with simple structures, erosion has decimated the record of ejection processes at terrestrial complex impact structures. There are a number of cases of preserved ejecta at complex impact structures in the former Soviet Union (e.g., Boltysh and El'gygytgyn [Russia] and Kärdla [Estonia]). In general, however, surface information is limited, or the ejecta have only been sampled by relatively few drill cores (Masaitis 1999). By far the greatest emphasis, with respect to ejecta at terrestrial complex impact structures, has been on the Ries structure, where ejecta deposits are preserved and have been studied for decades. The ejecta deposits at Ries are subdivided into a lower unit, Bunte breccia, and an upper unit, suevite. Ries is the type site for the occurrence and definition of both Bunte breccia and suevite.

Detailed studies of the Bunte breccia at Ries by Hörz et al. (1983) indicated that its primary components are almost exclusively derived from the upper-most section of the target, i.e., 600–700 m of Tertiary, Jurassic, and Triassic sediments. There is also a secondary component that consists of local material from the substrate on landing (Hörz et al. 1983). This secondary local component increases in volume fraction with increasing radial distance and, thus, correlates with the increasing impact or landing velocity of the primary components of the ejecta with distance. This observation is consistent with the qualitative model of Oberbeck (1975), which investigates the mobilization and incorporation of local substrate materials due to secondary kinetic energy from the impact of primary ejecta and correlates the degree of mobilization and incorporation with the impact velocity of the primary ejecta with distance. This process has become a basic tenet of the interpretation of lunar samples believed to be related to specific impact structures, including major basins (e.g., Spudis 1993), as well as of interpretations of remote sensing data from the Moon (e.g., Heiken et al. 1991).

The suevite breccia has also generally been interpreted as ejecta derived from deeper in the stratigraphic column at Ries. This interpretation was based largely on the occurrence of shocked crystalline clasts (e.g., Engelhardt 1990) and impact melt glass clasts (e.g., Engelhardt 1972) that had compositions equivalent to that of the crystalline basement rocks at Ries. In addition, it was generally held that shocked sedimentary material was not present in suevite (e.g., Pohl et al. 1977). Thus, the Ries ejecta display the inverted stratigraphy observed at simple impact structures, as exemplified by the Barringer crater and as predicted by cratering flow-field models.

There is, however, an apparent observational inconsistency in this conceptual model. The contact between the suevite and Bunte breccias is relatively sharp (e.g., Engelhardt et al. 1995). If the kinetic energy contained in the primary Bunte breccia materials resulted in the incorporation of the local substrate on landing, why did the same physical processes not apply to suevite ejecta, thereby, resulting in the incorporation of local Bunte breccia in the suevite? This contradiction has led to modified ejection scenarios involving a major role for the atmosphere (Engelhardt 1990) or the formation of suevite as some form of an ejecta "cloud" and ignimbritic flows (Newsom et al. 1990).

Recent research on the Ries suevite ejecta external to the structure has indicated that, in addition to impact melt glass clasts derived from the crystalline basement, it also contains carbonate impact melts derived from the overlying sedimentary succession (Graup 1999). Most recently, Osinski et al. (Forthcoming) have noted that other impact melts derived from other components of the sedimentary succession (e.g., Triassic sandstones) are present in the Ries suevite, both as discrete clasts and in its groundmass. This discovery has led to the reinterpretation of the Ries suevite as being more akin to an impact melt breccia (Osinski et al. Forthcoming) as opposed to its previous descriptive definition as a clastic matrix breccia, with impact melt and other shocked lithic clasts (Stöffler and Grieve 1994). In this respect, it may be more similar to the interpreted impact melt deposits believed to lie exterior to lunar and venusian complex impact structures (Cintala and Grieve 1998).

The genesis of such exterior melt deposits has been generally unclear (e.g., Howard and Wilshire 1975; Hawke and Head 1977). Recent hydrocode modelling calculations, however, suggest that impact melt deposits could be shed off over heightened central peaks during the intermediate stages of the formation of complex impact structures (Ivanov and Artemieva 2002; Ivanov and Melosh 2003). This is a case where the increasing sophistication of modelling and its ability to extend out to relatively late times in the impact process is resulting in potential convergence and providing mutual constraints on both interpretations of observations and the results of modelling.

There have been few cases where the Z-model (Maxwell 1977) has been tested against observations at terrestrial complex impact structures. Where it has been attempted (e.g., Ries, Haughton), the results have not been completely successful (Hörz et al. 1983; Redeker and Stöffler 1988). In general, it has been difficult to match, in detail, appropriate maximum depths of ejection, ejection volumes, and estimates of transient cavity diameter, as constrained by observational information. The models, however, all had a constant Z (= 2.7), and there are uncertainties as to the depth point for the origin of the Z-model flow-field.

IMPACT BASINS

Form: Not Yet Well-Defined

As noted earlier, impact basins are defined as a larger craterform with one or more topographic ring structures interior to the main topographic rim (Spudis 1993). This definition is relatively straightforward in terms of its application on the Moon (although more interpretative for the oldest, heavily degraded basins; Spudis 1993). It has been, however, highly interpretive in the terrestrial environment. For example, Pike (1985) listed 17 terrestrial impact structures, ranging in diameter from 10 to over 200 km, which were interpreted to be impact basins. The problem is that the identified "rings" in these structures do not necessarily have morphological equivalence or primary structural significance with respect to the original craterform. For example, the annulus of hills at Gosses Bluff was considered a ring. It is a topographic ring (Fig. 9), but it is an erosional remnant of the original structural uplift (Milton et al. 1996). Similarly, the annular lake at Manicouagan (Fig. 3) is listed as a ring but represents an over deepened annular valley produced by glaciation and formed at the competency boundary between the coherent impact melt and the parautochthonous fractured target rocks (Grieve and Head 1983). In other cases, e.g., West Clearwater and Deep Bay (Canada), the "ring" is exterior to the rim of the structure and corresponds to the annulus of increased target rock fracturing that occurs around complex impact structures.

Fig. 9. Oblique aerial photograph of the highly degraded Gosses Bluff complex impact structure showing the physical ring of hills as an erosional remnant of the original structural uplift. (See text for more details.)

When limited to the largest known terrestrial impact structures: Chicxulub (Mexico), Sudbury, and Vredefort, there is observational evidence of ring forms, but again, the genetic relationship and equivalence to what is observed at impact basins on the Moon (e.g., Spudis 1993) is not clear. At Chicxulub, interpretations of offshore reflection seismic data indicate an interior topographic peak ring (Fig. 10; Morgan and Warner 1999). At Sudbury, there appear to be rings of increased pseudotachylite development, which have been equated with the traces of super-faults related to rim collapse and modification (Spray et al. 2004). At Vredefort, there are a series of concentric anticlines and synclines, reflecting structures in the sub-floor of the impact structure. In terms of erosion, these structures represent increasing erosion and depth of exposure in the order of Chicxulub, Sudbury, Vredefort, i.e., crater fill and rim preserved, some crater fill but no rim preserved and some crater floor exposed, and no crater fill or rim but crater floor exposed, respectively. How the observed "rings" physically or genetically relate to each other is not clear at present (Grieve and Therriault 2000). At present, Chicxulub is, in fact, the only terrestrial impact structure with clear indications of an inner topographic ring that is not some erosional artefact or other target effect.

Formation: An Extension of Structural Uplift

Lunar observations indicate that the exposed volume of central peaks increases with increasing rim diameter (Fig. 11). There is, however, a change in the amount of relative increase in volume above rim diameters of ~80 km (Fig. 11; Hale and Grieve 1982). This corresponds to a reduction in the rate of increase in relative height of the central peak, which reaches a maximum of ~2 km (Pike 1977; Hale and Grieve 1982), suggesting some maximum equilibrium physical height. This

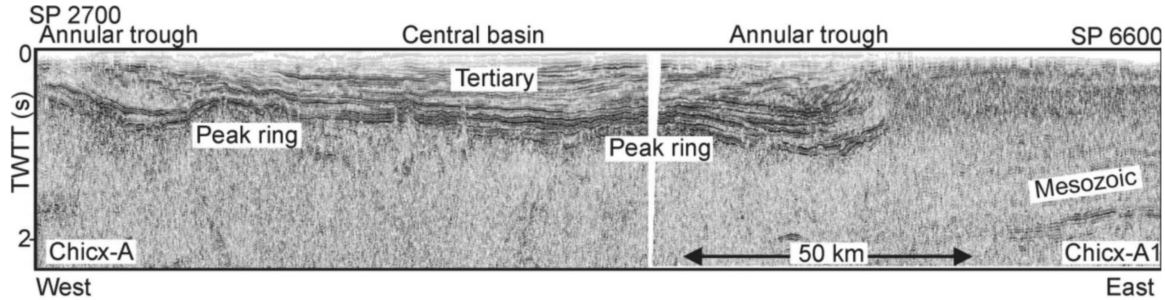

Fig.10. Reflection seismic cross-section of Chicxulub along Chicx-A and -A1 (Bell et al. Forthcoming). The post-impact Tertiary sediments are clearly identifiable as high-frequency reflections from 0 to ~1 sec two-way travel time (TWTT). A topographic peak ring, with draped sediments, is identifiable on the floor of Chicxulub and separates the central basin from a surrounding annular trough.

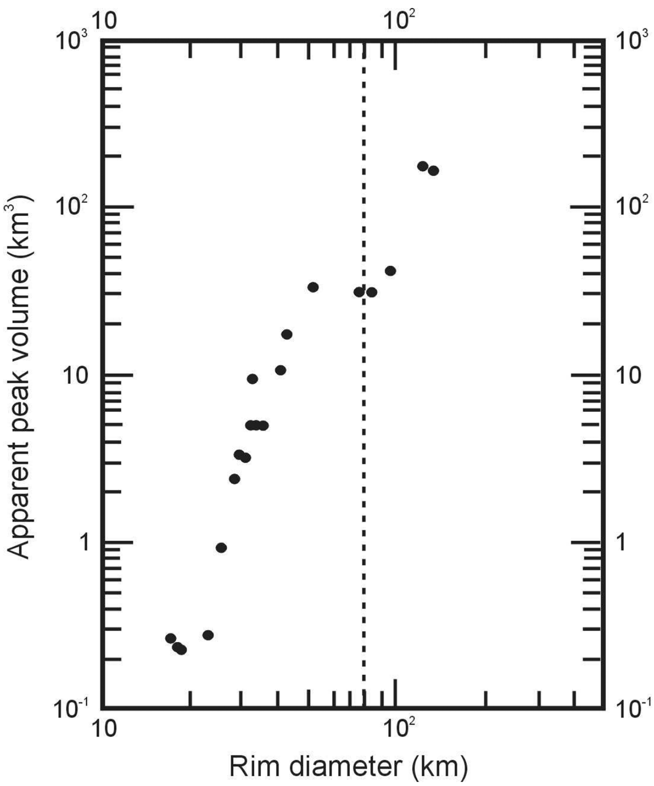

Fig. 11. Apparent peak volume plotted against crater rim diameter for 20 lunar craters. The dashed line represents a rim diameter of 80 km. (based on Hale and Grieve 1982.)

reduction in the rate of increase of central peak volumes coincides with the appearance of rings of high amplitude floor roughening between the central peak and the rim. This is generally consistent with ring formation being an extension of the process of structural uplift in complex impact structures with the dynamic collapse of an initially over heightened central peak. The volume of uplifted material, in excess of what can be accommodated by a near-equilibrium central peak, is manifested as topographically emergent interior rings (Melosh 1989).

Some terrestrial observational data also favor this extension of the structural uplift process in the formation of rings. Assuming that Sudbury and Vredefort had the original form of impact basins, they display the sequence of a central structural uplift with lithologies progressively decreasing in age outward and representing successively nearer surface lithologies in the original target (e.g., Hart et al. 1990; Boerner et al. 1999). The expression of the structural uplift at Sudbury has been complicated by post-impact tectonic movements, particularly in the south, deforming a large portion of the structure (Riller et al. 1998). The structural relations at Vredefort are less complicated but, like Sudbury, are a reflection of not only impact processes but of pre- and post-impact tectonic processes (e.g., Henkel and Reimold 1998). A recent detailed study, however, has indicated discrete block rotations in the uplifted center of Vredefort, with the rotations decreasing toward the center (Lana et al. 2003). At the erosional level of Vredefort, this rotation is consistent with the attenuation of block movements in the formation of the central structural uplift. Lana et al. (2003) also suggest that lubrication by the pervasive network of pseudotachylite veins may have provided the necessary strength degradation during modification to allow differential rotation and slip during uplift, with high strain rate deformation distributed as discrete shear in the pseudotachylite network.

What is observed at Sudbury, in terms of pseudotachylite development and block movements, has been interpreted in terms of collapse faulting exterior to the transient cavity during the formation of the final rim of the basin. The apparent relative increase and decrease of pseudotachylite development has been compared to the apparent $\sqrt{2}$ spacing of topographic rings in lunar impact basins (Spudis 1993; Spray and Thompson 1995). However, given the limited exposures, the apparent distribution of the pseudotachylite could be a function of the quantity and quality of rock exposures (L. Thompson 2003, personal communication).

More recent hydrocode models (e.g., Collins et al. 2002; Ivanov and Artemieva 2002) indicate the development of an over heightened central peak and its subsequent collapse. In this case, the models have achieved a degree of sophistication with respect to large-scale natural impact events such that their results can be tested with field observations. In the case of

Chicxulub, field testing would be related to the stratigraphic relations in the peak ring, which the models suggest would display inverted stratigraphy (e.g., Collins et al. 2002). As this peak ring will consist of crystalline material at Chicxulub, such a test of the models would depend on shock metamorphic zoning. Given that Chicxulub is buried by ~1 km of post-impact sediments, such a test can only be administered through drilling. The geographical location of the topographic peak ring has only been determined with a degree of certainty by interpretations of reflection seismic profiles offshore in the Gulf of Mexico (Morgan and Warner 1999). It would, therefore, seem that further constraints on the formation of rings in impact basins by ground-truth studies depend critically on such enterprises as the Integrated Ocean Drilling Program (IODP).

Shock Attenuation: Little Known

Attempts to determine the systematic variation of recorded shock pressures in the parautochthonous target rocks at Vredefort have had very limited success. This is due to extensive recrystallization of quartz by a thermal overprint (Grieve et al. 1990), which has resulted in the annealing of almost all PDFs with the exception of basal PDFs, which are Brazil twins (e.g., Goltrant et al. 1991). There is, however, some indication of increasing recorded shock pressure as the center of the structural uplift is approached (Grieve et al. 1990).

Similar annealing effects are present at the contact between the SIC and parautochthonous target rocks at Sudbury (Dressler 1984), in drill cores that penetrate the floor of the structure beneath the SIC (Geological Survey of Canada, unpublished data), and in the overlying Onaping Formation (Joreau et al. 1996). There is, however, evidence of decreasing recorded shock pressure, as determined by PDFs in quartz, away from the SIC in the North Range (Dressler 1984). Similar evidence is lacking in the South Range due to a Penokean orogenic overprint.

Thus, in general, current information on shock metamorphism at potential terrestrial impact basins results in no significant constraints on cratering processes beyond the fact that recorded shock zoning in the structural uplift appears to be consistent with that observed at smaller terrestrial complex impact structures. It would be interesting, however, to have information on the highest recorded shock pressure in the parautochthonous rocks of the structural uplift to determine the potential role of impact melting in the creation of ring structures, as opposed to central peaks. For example, Grieve and Cintala (1992) have suggested that relatively deeper impact melting with respect to transient cavity dimensions, due to differential scaling at larger impact structures, could lead to progressive weakening and, finally, melting of the center of the transient cavity floor prior to uplift, leading to a ring form on uplift.

Ejecta: Again, Only One Example

Chicxulub is the only putative terrestrial impact basin with preserved ejecta. Information on the ejecta is available from drill cores and outcrops. Close to the structure, equivalents to Bunte and suevite ejecta are observed and are several hundred m in thickness (Urrutia-Fucugauchi et al. 1996). The thickness of the ejecta attenuates with distance, with complex relations due to tsunami influences in some sections, and ultimately becomes a global mm-thin ejecta layer defining the Cretaceous/Tertiary (K/T) boundary (e.g., Smit 1999). While there is a high level of interest in these ejecta in relation to the deleterious effects of the impact on the global environment and related biosphere extinctions (e.g., Pope et al. 1997; Pierazzo et al. 1998), they provide few constraints, at this time, on such parameters as maximum depth of excavation.

CONCLUDING REMARKS

The value of geological and geophysical observations at terrestrial impact structures, as intellectual collateral for understanding impact processes, is clear. In some cases, it has been several years after a particular observation has been made that its significance becomes apparent. For example, the observation that the impact melt volumes at a small number of terrestrial impact structures appeared to increase with crater size (Grieve et al. 1977) has stood the test of time as observations have been made at more structures. It was not, however, until theoretical and modelling considerations introduced into the literature the concept that crater volumes and melt volumes scale independently (e.g., Melosh 1989) that the significance of the observations became apparent and they could be used to validate analytical models of differential scaling (e.g., Grieve and Cintala 1992) and more recent hydrocode models (e.g., Pierazzo et al. 1997).

Interpretations of observations have resulted in an evolution in the understanding of crater processes, e.g., the concept of the transient cavity. They continue to do so. For example, the polymict allochthonous breccia sheet that is contained within the Haughton structure was recognized as the spatial equivalent of coherent impact melt sheets observed at other terrestrial complex impact structures in crystalline targets (Grieve 1988). Its origin, however, was considered analogous to a suevite deposit (Redeker and Stöffler 1988), which would imply that it consisted of shocked clasts set in a clastic matrix. While the impression of a clastic matrix is evident in outcrop, in part because of the physical (highly friable) appearance of the lithology due to the actions of frost heave, more recent examination of the matrix at the 10–100 micron scale of the SEM indicates that the matrix contains both Si-Al-Mg-rich glass and microcrystalline carbonate containing a few weight percent Si and Al. Globular textures of calcite within the silicate glass and quench crystals of pyroxenes point to the matrix phases being originally molten and rapidly cooled (Osinski and Spray 2001). With these new observations on hand, the breccia deposit has been interpreted not only as the spatial equivalent but as the genetic equivalent of the coherent melt sheets found at complex impact structures in crystalline targets (Osinski and Spray 2001). This begins to close a loop in theoretical considerations and analytical models that indicated that sediments undergo impact melting at shock pressures equivalent to or lower than those for crystalline rocks but that coherent impact melt rocks of sedimentary origin were apparently absent in the observed record, apparently being replaced by clastic "suevite" breccia (Kieffer and Simonds 1980).

There is at least one case where interpretations of observations are at variance with hydrocode models. Based on observational data, Therriault et al. (1997) and Henkel and Reimold (1998) call for estimated original rim diameters of ~300 km and 280 km, respectively, for Vredefort. Conversely, model calculations by Turtle and Pierazzo (1998) and Turtle et al. (2003) indicate a smaller estimated diameter of 120–200 km as the consequence of an 80–100 km-diameter transient cavity. Turtle and Pierazzo (1998) suggest that the variance is due to a misinterpretation of the geometry of the post-modification isobars by Therriault et al. (1997). The larger diameter estimate, however, is based on more observations than post-modification shock geometry. It is supported by the semi-independent estimate of Henkel and Reimold (1998) based on geophysics, and such a diameter is consistent with estimated values of the amount and diameter of structural uplift and preservation of outliers of cover rocks in the target (Therriault et al. 1997). In this case, the weight of evidence would seem to suggest there might be some misestimate in the initial parameters of the model calculations.

One should not, however, automatically favor interpretations based on observations over results based on models. The "observational" record includes considerable uncertainties. For example, the Chesapeake impact structure, USA is cited as being 90 km in diameter (Poag 1996). This is a buried impact structure and is the likely source structure of the North American microtektites in Ocean Drilling Program (ODP) cores from the North Atlantic, on the basis of age and isotopic geochemistry (e.g., Whitehead et al. 2000). The cited diameter is based on reflection seismic data, which has been truncated at deeper levels for commercial proprietary reasons. The radial variation in thickness of the North American microtektites in ODP cores is, however, more consistent with a smaller source impact structure (Glass et al. 1998), on the basis of ejecta scaling relations (Stöffler et al 1975). A diameter closer to 40 km is more consistent with the residual gravity anomaly over Chesapeake. There are other examples where the interpretations of observational data to estimate such basic parameters as the diameters of buried terrestrial impact structures are contentious, e.g., Woodleigh (Mory et al. 2000; Renne et al. 2002) and Morokweng, South Africa (Andreoli et al. 1999; Reimold et al. 2002). In such cases, observational data sets are limited and incomplete.

As supplementing these data sets is costly, it behooves both the observational and modelling communities to agree on a set of critical impact structures and the observations that are required to continue the advance in the understanding of impact processes in concert with advances in modelling. Considerable data, cores, etc. already exist from terrestrial impact structures but are dispersed throughout the community and among nations. Some of these data and cores are not readily available for examination that would further understanding of impact processes. This is particularly true of the vast amounts of information potentially available from impact structures in the former Soviet Union (e.g., Masaitis et al. 1980; Masaitis 1999). With a number of terrestrial impact structures being the subject of drilling proposals at various stages of maturity (e.g., Bosumtwi in Ghana, Chesapeake, Chicxulub, El'gygytgyn, Sudbury) to the International Continental Drilling Program (ICDP) or IODP (e.g., Chicxulub, Mjølnir in the Barents Sea), it is appropriate for the community to discuss and consider the creation of a centralized repository for new and existing cores and data from terrestrial impact structures, perhaps under the auspices of an international body such as ICDP. Development of such a repository would certainly be more cost effective than redrilling or reinvestigating specific terrestrial impact structures. It would also create a critical mass of information, leading to potential synergies, significant advances in understanding, and identification of critical gaps in information from the terrestrial impact record as a whole.

Acknowledgments–We would like to acknowledge reviews by M. Pilkington, B. French, and E. Turtle. This is Geological Survey of Canada Contribution 2003115.

Editorial Handling—Dr. E. Pierazzo and Dr. R. Herrick

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