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Chicxulub – The K-T Boundary Impact Crater: A Review of the Evidence, and an Introduction to Impact Crater Studies

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20 Text-Figures, 3 Tables and 1 Plate (in pocket)

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Yukatán
K-T boundary
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Chicxulub – der Impaktkrater an der Kreide/Tertiär-Grenze

Zusammenfassung

Während der letzten 15 Jahre hat eine intensive Diskussion der Ereignisse, die zum Massensterben am Ende der Kreidezeit vor 65 Millionen Jahren geführt haben, stattgefunden. Bisher wurde in den Geowissenschaften die Bedeutung von Impaktereignissen auf die geologische und biologische Entwicklung der Erde unterschätzt bzw. ignoriert, trotz der Tatsache, daß Impaktprozesse deutlich sichtbare Narben an den Oberflächen der Planeten und Monde des Sonnensystems hinterlassen haben. In detaillierten Untersuchungen, hauptsächlich in den letzten 40 Jahren, wurden auf der Erdoberfläche bisher etwa 150 Impaktkrater nachgewiesen.

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Seit 1980 hat man überzeugende Anzeichen dafür, daß vor 65 Millionen Jahren ein gigantisches Impaktereignis auf der Erde stattgefunden hat, zeitgleich mit dem Massensterben am Ende der Kreidezeit. Die Entdeckung eines zugehörigen Impaktkraters hat allerdings bis etwa 1991 gedauert, als der Chicxulub-Krater auf der Halbinsel Yucatán (Mexico) als Impaktstruktur erkannt wurde. Die Entdeckung hat so lange gedauert, da die Kraterstruktur von tertiären Sedimenten mit einer Dicke von etwa 1 km bedeckt ist und an der Oberfläche nicht zu erkennen ist. Die seit 1991 folgenden Untersuchungen haben gezeigt, daß es sich bei dieser Struktur, die einen Durchmesser von etwa 200–300 km aufweist, um einen der größten Impaktkrater auf der Erde handelt. Die Anzeichen für eine Impaktentstehung der Struktur umfassen: Schwerkraft-, Magnetik- und Seismik-Anomalien; Entdeckung von Mineralen mit klarer Stoßwellenmetamorphose (dies ist ein eindeutiger Beweis für eine Impaktentstehung) in Bohrkernproben; Nachweis einer meteoritischen Komponente in den Impaktschmelzen im Krater; isotopengeochemische Daten, die eine klare Verbindung mit den Gesteinen der Kreide-Tertiär-(K-T-)Grenze herstellen; und Altersbestimmungen an den Impaktschmelzen, die eine Zeitgleichheit mit der K-T-Grenze nachweisen.

Der Einschlag, der zur Bildung des riesigen Chicxulub-Kraters führte, war also verantwortlich für eine der traumatischsten Ereignisse in der jüngeren Erdgeschichte und führte zum Aussterben von mehr als der Hälfte aller Tier- und Planzenarten auf der Erde vor 65 Millionen Jahren.

Abstract

Over the last 15 years, renewed interest in the events that led to the extinction of the majority of all life on earth at the end of the Cretaceous period, 65 Ma ago, implicated a large-scale asteroid or comet impact as the cause of this catastrophe.

In the past, impact cratering as a geological process has been rather unappreciated by the general geological community, despite the fact that on all other planets and satellites with a solid surface, impact cratering is the most important process that alters the surface at the present time, and during most of the history of the solar system. Detailed studies, mainly since the 1960s, have led to the recognition of about 150 impact structures on earth. Although the mass extinction at the Cretaceous–Tertiary (K-T) boundary was well known to geologists, no single cause had been identified.

However, in 1980, the first compelling evidence for an asteroid or comet impact at that time was published. During the 1980s, additional evidence in support of this proposal mounted, but the proponents of the “impact hypothesis” had to explain the absence of a large impact crater of the proper age. This deficiency was remedied in the early 1990s, when a large concealed structure centered at the northwestern tip of the Yucatan peninsula in Mexico was recognized as an impact structure, the Chicxulub crater. In the present paper, the evidence leading to the recognition of the Chicxulub structure is discussed. To put these studies in a proper framework, some fundamental principles of impacts and how to recognize impact craters are also reviewed. The formation conditions of impact craters lead to pressure and temperature conditions in the target rocks that are significantly different from those reached during any internal terrestrial processes. Among the most characteristic changes induced by the impact-generated shock waves are irreversible changes in the crystal structure of rock-forming minerals such as quartz and feldspar. These shock metamorphic effects are characteristic of impact and do not occur in natural materials formed by any other process. It was such mineralogical evidence, together with several independent chemical, isotopic, dating, and geophysical studies, that provided abundant testimony for an impact origin of the Chicxulub crater, and the enormous environmental consequences of such a large-scale, short-time event that ultimately caused the demise of a significant fraction of the biomass on earth at the time of the K-T boundary.

1. Introduction: Impact Cratering in Geology and the K-T Boundary

The debate regarding the cause of the mass extinction that marks the end of the Cretaceous period, at the Cretaceous–Tertiary (K-T) boundary, has been one of the liveliest exchanges in the geological (and related) sciences during the 1980s and the early 1990s. Renewed interest in the events at the K-T boundary was kindled by the publication of a paper by ALVAREZ et al. (1980). These authors found that the concentrations of the rare platinum group elements (PGEs; Ru, Rh, Pb, Os, Ir, and Pt) and other siderophile elements (e.g., Co, Ni) in the thin clay layer that marks the K-T boundary are considerably enriched compared to those found in normal crustal rocks. These significant enrichments (up to 4 orders of magnitude) and the characteristic interelement ratios were interpreted by ALVAREZ et al. (1980) to be the result of a large asteroid or comet impact, which also caused extreme environmental stress. Ever since the K-T boundary impact theory was proposed, alternative hypotheses were proposed. However, the evidence continued to mount in favor of the impact theory, as documented by papers in, e.g., the proceedings of the so-called Snowbird I, II, and III conferences (SILVER & SCHULTZ, 1982; SHARPTON & WARD, 1990; RYDER et al., 1996; see also DRESSLER et al., 1994). Before discussing in more details some of the arguments in favor of the impact theory, and the discovery of the associated impact crater, it may be useful to briefly review our basic knowledge of terrestrial impact craters and shock metamorphism. The first part of the following discussion is based in part on KOEBERL (1994) and KOEBERL & ANDERSON (1996b).

Researchers not directly involved in this debate may not be familiar with the subject and the concept of impact cratering. This is because the importance of impact cratering on earth has traditionally not been accentuated in classical geological studies. In traditional geology, as established by, e.g., James HUTTON (1726–1797) and Charles LYELL (1797–1875), it is widely accepted that slow, endogenic processes lead to gradual changes in our geological record. In this conception, which is called uniformitarianism, it is preferred to call upon these internal forces, before resorting to seemingly more exotic processes to explain geological phenomena that often give the impression of occurring over very long periods of time. In this view, geological processes occur gradually at almost constant rates. In contrast, impact appears as an exogenic, relatively rare, violent, and unpredictable event, which seems to violate every tenet of uniformitarianism. However, this is not necessarily so, as the main question regards the time scale involved. Geological uniformitarianism includes integrating various individual catastrophes, such as earthquakes, volcanic eruptions, floods, landslides, etc., over a long period of time. It seems that, in some applications of uniformitarianism to geology, it has been defined only as such processes that can be witnessed during human lifetimes, and the formation frequency of at least larger impact craters is clearly outside the reach of recent history. Thus, the explanation of craters on the moon or on earth as being of impact origin has been opposed by traditional geologists over much of this century; this is amplified quite pointedly by a little-known study of WEGENER (1922), who concluded (partly from experiments that he conducted himself) that the craters on the moon are of meteorite impact origin.

The development of the study and acceptance of impact craters and processes over this century is somewhat similar to the history of accepting the process of plate tectonics (see, e.g., MARK, 1987; HOYT, 1987; MARVIN, 1990, for a history of impact crater studies).

The reservations towards accepting impact cratering as a wide-spread geological process seems to have been mainly cultivated only by certain parts of the geoscience community, while other branches of the natural sciences have much more readily embraced the importance of impact cratering. The debates focussing on the research related to the cause of the mass extinctions at K-T boundary have already led to a historical and sociological evaluation (GLEN, 1994).

In this survey, GLEN (1994, p. 51/52) found that

> *" ... those scientists who were very narrowly focused in their studies ... were less likely to embrace any part of the impact hypothesis ... "*

and that

> *" ... resistance to the [impact] hypothesis seemed inverse to familiarity with impacting studies."*

Planetary exploration and extensive lunar research in the second half of our century led astronomers and planetary scientists to recognize that essentially all craters visible on the moon are of impact origin. This conclusion implies that, over its history, the earth was subjected to an even larger number of impact events compared to the moon, because of its larger gravitational cross-section. It is not completely clear why many geologists found this a rather exotic conclusion.

Part of the reason may be that terrestrial processes (weathering, plate tectonics, etc.) effectively work to obliterate the surface expression of these structures on earth. Through studies of their orbits, astronomers have a relatively good understanding of the rate with which asteroids and comets strike the earth (e.g., SHOEMAKER et al., 1990; WEISSMAN, 1990). For example, minor objects in the solar system with diameters $\leq$1 km (mainly asteroids), collide with the earth at a frequency of about 4.3 impacts per million years (SHOEMAKER et al., 1990), and each such impact forms a crater about $\geq$10 km in diameter. Impactors about 2 km in diameter collide with the earth about every 1–2.10^6 years.

Impacts of earth-orbit crossing asteroids (such as the one shown in Text-Fig. 1) dominate the formation of craters on earth that are smaller than about 30 km in diameter, while comet impacts probably form the majority of craters that are larger than about 50 km in diameter (SHOEMAKER et al., 1990).

On all planets and satellites of the solar system that have solid surfaces, e.g., Mercury, Venus, Mars, and the satellites of all planets, impact cratering is either the most important, or one of the most important processes that affects the shaping of their surfaces (see, e.g., TAYLOR [1992] for a discussion and further references). Thus, planetary scientists, astronomers, and meteoriticists have grown accustomed to view

> *" ... large-body impact as a normal geological phenomenon – something to be expected throughout earth history – but another group, the paleontologists, is confounded by what appears to be an ad hoc theory about a nonexistent phenomenon ... "* (RAUP in GLEN, 1994, p. 147).

It seems that one scientist's uniformitarianism is another scientist's deus ex machina.

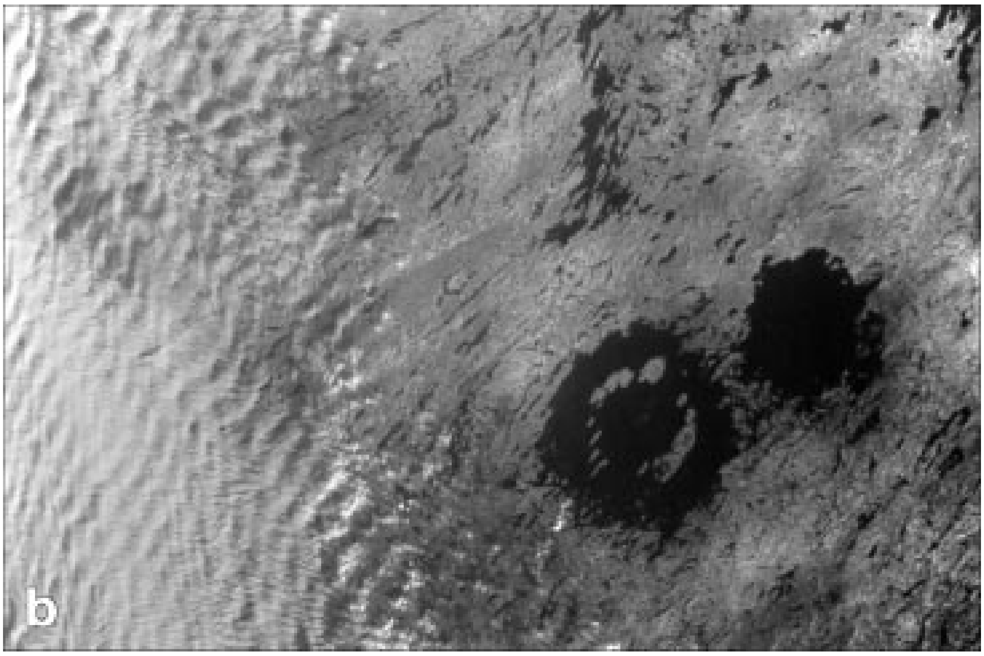
Text-Fig. 1.
Asteroid Ida, as photographed by the Galileo spacecraft.
This asteroid has the typical appearance of the objects that form large craters on earth when in an earth-crossing orbit.

An aspect of impact cratering that may be underestimated is the influence of impacts on the geological and biological evolution of our own planet. Many earlier interpretations of rare geological features were confined to the preconceived limitations of internal processes that were used to explain the evolution of the earth. However, we finally begin to appreciate that impacts played a larger role in the evolution of the earth than was realized before. The studies related to the events of the K-T boundary provide a specific example in the discussion about the causes for major biological extinctions. While many researchers now accept that a large impact event has played a major role in the K-T boundary mass extinctions, the acceptance rate was slow because until recently no associated impact crater was known. As discussed in greater detail later in this paper, the Chicxulub crater in Mexico has now been recognized to fit this requirement perfectly. However, despite intensive efforts of several researchers, its discovery took a long time, because it is covered by about 1 km of Tertiary sediments and has no surface expression. Considering the past history of this research field, it is not surprising that there are still some researchers who accept neither the overwhelming evidence for an impact 65 Ma ago, nor any of the consequences of such an impact. However, as noted by GLEN (1994), more familiarity with impact studies should help to provide the basis of understanding this important geological phenomenon.

Numerous impact craters of various sizes that are covered by later deposits unrelated to the impact event are certain to exist on earth. Most of them have yet to be discovered. How the events that led to their formation affected evolution as severely as Chicxulub, has to be the subject of future research. Impacts similar to those that occurred in the past will happen again. Even the impact of relatively small asteroids or comets can have disastrous consequences for our civilization. There is a 1 in 10,000 chance that a large asteroid or comet 2 km in diameter (corresponding to a crater of about 25–50 km in diameter) may collide with the earth during the next century, severely disrupting the ecosphere and destroying a large percentage of the earth's population (CHAPMAN & MORRISON, 1994). However, even the impact of much smaller bodies can be devastating. The understanding of impact structures and the processes by which they form should, thus, be of interest not only to earth scientists, but also to society in general.

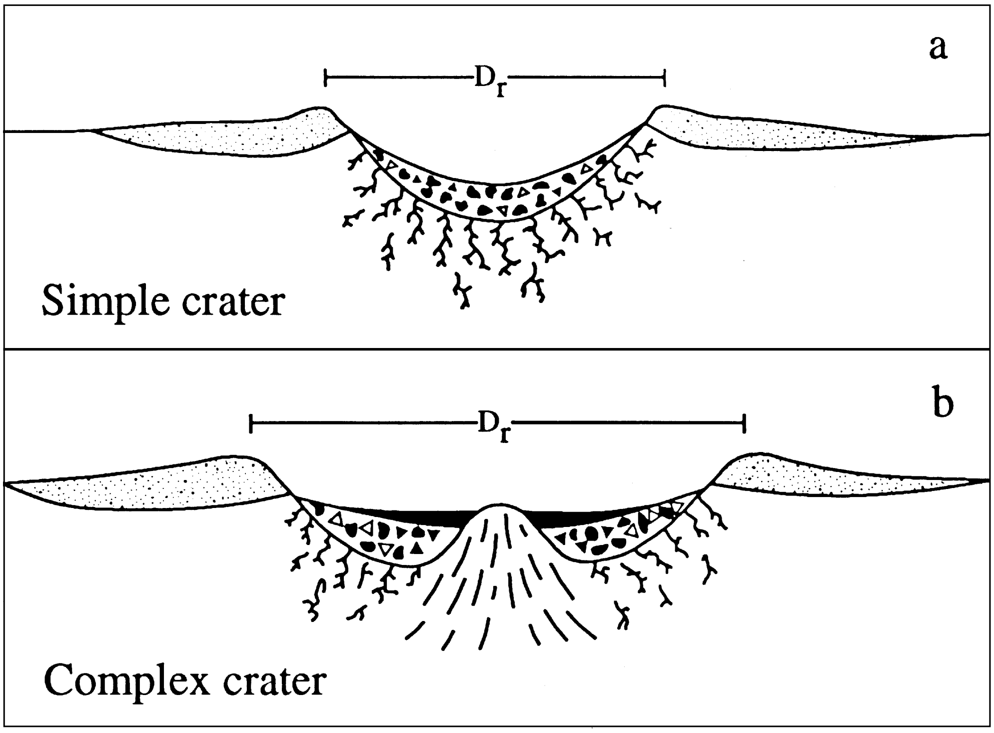
2. Crater Morphology

Whereas craters on earth can be either eroded or hidden from the view, impact craters on the moon or on other planets have, in general, been identified only by their morphology. The study of craters on earth allows to obtain ground truth for planetary impact processes. Impact changes the geological structure of the target area in



Text-Fig. 2.
Typical appearances of simple and complex impact craters on the earth.
a) Simple crater: Roter Kamm crater, Namibia; 2.5 km diameter. Space Shuttle image 61C-40-001.
b) Complex crater: East (right) and West (left) Clearwater twin impact crater, showing the structural uplift as a central peak-ring. Space Shuttle image 61A-201-088.

Text-Fig. 3.
Schematic cross sections through simple (top) and a complex (bottom) impact craters.
After KOEBERL & SHARPTON (1992) and KOEBERL (1994).



a characteristic way. One way to help distinguish volcanic craters or plutonic pipes from meteor impact craters is to study the deep basement of the feature. Meteorite impact craters are a circular surfical feature without any deeper roots, while in volcanic structures, the disturbances continue to (or, rather, emerge from) great depth. Impact craters are practically always very circular. Non-circular structures are rare, resulting either from highly oblique impacts (see, e.g., SCHULTZ et al., 1994), or post-impact tectonic processes (such as the Sudbury structure in Canada; see, e.g., STÖFFLER et al., 1994). It may, thus, be useful to distinguish between "crater" (the pre-erosional depression) and "structure" (the geological remnant after erosion, deformation, etc.).

Impact craters show, in general, two distinctly different morphological appearances, depending on their size. Small craters with a diameter of ≤4 km in crystalline rocks (≤2 km in sedimentary rocks) have a bowl-shaped depression and an upraised rim and are called simple craters, while features 4 km in diameter have a central uplift, are shallower, and are called complex craters. The diameters mentioned above depend on the surface gravity (and the target rock type) and are valid for the earth; on the moon or other planets, the transition from simple to complex craters occurs at a slightly different diameter. All craters have an outer rim and some crater infill (e.g., brecciated and/or fractured rocks, impact melt rocks). The central structural uplift in complex craters consists of a central peak or of one or more peak ring(s), and exposes rocks that are uplifted from considerable depth. Examples for simple and complex craters are shown in Text-Fig. 2. Text-Fig. 2a shows the 2.5 km diameter Roter Kamm crater in Namibia, a simple crater, while Text-Fig. 2b shows a pair of complex craters, the East and West Clearwater craters in Canada (23 and 36 km in diameter). Cross-sections through typical simple and complex craters are shown in Text-Fig. 3, indicating the difference in depth to diameter ratio and the interior structure.

3. Distribution, Size, and Age of Impact Craters on Earth

So far, no large impact event has been observed by humans over the last several thousand years (which is, of course, not a geologically long period of time). Thus, for understanding impact processes, we are restricted to inferences drawn from impact experiments (see below), and the detailed study of impact craters on earth. Unfortunately, impact structures are not easily recognized on earth, as many factors conspire to rapidly obliterate or cover most impact structures formed on earth. In contrast to the moon or most other planets, the surface of the earth is continuously reshaped by erosion, sedimentation, volcanism, and tectonics (rifting, subduction, faulting, etc.), leading to a geologically rapid eradication of much of the impact record on the surface of the earth.

A more complete understanding of the criteria for the recognition of impact structures than earlier in this century, and the dedication of a few researchers, brought a recent increase in the number of impact structures known on earth. While in 1972 only about 50 confirmed impact craters were listed, by 1991 the number had increased to more than 130 (GRIEVE, 1991), and stands now (1996) close to 150 (e.g., GRIEVE & SHOEMAKER, 1994). Currently, about three to five new craters are discovered every year. However, it is unfortunate that our knowledge of many of the known impact structures is very restricted. While it is not surprising that many of the 15 impact structures known in Africa are not well studied (see KOEBERL [1994] for a review), it is more of an embarrassment that almost two thirds of the confirmed or probable impact craters in the USA have only been studied superficially (see KOEBERL & ANDERSON, 1996b).

As far as we know today, celestial mechanics (of the encounter of the orbits of comets and asteroids with the cross-section of the earth) requires a random distribution of impact sites on the earth (as well as on other planets). However, the spatial distribution of terrestrial impact craters known to date is not random. Almost all of them are on land, with just very few exceptions (see below), and all are on continental crust. In addition, the craters on land are not randomly distributed, but are concentrated in North America (especially Canada), Australia, parts of the former Soviet Union, and northern Europe. These nonrandom distributions result from the facts that these regions have been studied better than other areas on earth, and that practically all of these locations are on (old) cratons

(GRIEVE & SHOEMAKER, 1994). Other cratonic areas, e.g., in Africa, have not been studied as thoroughly, but recent studies indicate that many additional craters will be discovered in those areas as well (KOEBERL, 1994).

The diameters of impact craters, in contrast to the spatial distribution of the known craters on earth, show a more arbitrary variation, which is, however, the result of biased processes. Astronomical studies indicate that small craters (<20 km diameter) should be more common than larger structures, with craters on the order of 1 km diameter being most abundant. However, the terrestrial processes mentioned before create a severe bias against the preservation, and, thus, discovery of craters of any size, especially the smaller craters. The erosional processes that obliterate small craters after a few million years create a severe deficit of smaller (<20 km diameter) craters, compared to the number that is expected from the number of larger craters, and astronomical observation (GRIEVE & SHOEMAKER, 1994). This also indicates that most small craters have to be young. Older craters of larger initial diameter also suffer during erosion, which may lead to the destruction of the original topographical expression, or to the burial of the structures under later sediments. Thus, the currently observed size distribution of impact craters on earth has been severely affected by terrestrial processes, and appears random rather than dominated by smaller craters, as is observed on other solar system bodies.

Terrestrial craters are, so far, the only ones in the solar system (with the exception of a few lunar craters where samples were brought back), for which ages can be measured directly. The methods most commonly used for determining crater ages include isotopic age determinations (e.g., using K-Ar, ^{40}Ar-^{39}Ar, fission track, Rb-Sr, Sm-Nd, or U-Th-Pb dating) and biostratigraphic or stratigraphic ages (by studying, for example, post-impact lake sediments). Unfortunately, accurate ages are, so far, available only for only less than half of the known terrestrial craters (GRIEVE & SHOEMAKER, 1994). In addition, not all of these ages are known with the same accuracy, leading to large differences in the accuracy of the quoted dates.

A problem often encountered in dating impact craters is posed by the absence (or inaccessibility) of rocks that can be dated properly. The selection or isolation of suitable impact-derived rocks, that are reasonably fresh and unaltered, is often more difficult than the age measurement itself. For practically all radiometric age determinations, it is important to obtain lithologies that have been completely reset by the thermal event associated with the impact. Some of the most suitable material for such dating efforts are totally melted and quenched rocks, forming impact glasses, or minerals that have recrystallized from the melt. However, impact glasses are often inhomogeneous and/or devitrified, yielding fine-grained impact melt rocks or breccias that are often clast-rich and difficult to date by the K-Ar method, because of inherited Ar in the clasts. Dating impact craters is complicated and tedious and, if not done with utmost care, can easily lead to erroneous results (see, e.g., BOTTOMLEY et al. [1990] and DEUTSCH & SCHÄRER [1994] for reviews of impact crater dating).

Recently, considering a possible connection with a possible correlation between large-scale impact events and mass extinctions, the question of a periodicity in the ages of impact craters has been raised. However, if known crater ages and the errors associated with these age determinations are being assessed, no statistically significant periodicity in the ages of impact craters on earth was found (GRIEVE & SHOEMAKER, 1994).

4. Formation of Impact Craters

The formation of a crater by hypervelocity impact is a very rapid process that is typically divided into three stages:

1) compression stage,
b) excavation stage, and
3) post-impact crater modification stage.

Cratering mechanics has been studied for military and scientific reasons for many decades. Most of especially the initial phases of crater formation are relatively well understood from theoretical and experimental considerations; however, due to space limitations, the reader is referred to the literature (see, e.g., GAULT et al., 1968; RODDY et al., 1977; MELOSH, 1989; and references therein) for a detailed discussion of the physical principles of impact crater formation. Here, only a few key concepts should be mentioned.

The large amount of kinetic energy ($1/2\ mv^2$) that is released upon the impact of a large meteorite, asteroid, or comet, was largely underestimated earlier in this century, because the velocities with which such bodies hit the earth have not been properly estimated. It is now know that any body that is not slowed down by the atmosphere will hit the earth with a velocity between about 11 and 72 km/s. An iron or stony meteorite 250 m in diameter has a kinetic energy equivalent to about 1000 megatons of TNT. The impact of such a body would produce a crater about 5 km in diameter. The relatively small Meteor (or Barringer) crater in Arizona (1.2 km diameter) was produced by an iron meteorite of about 30–50 m in diameter. Many of the characteristics of an impact crater are the consequence of the enormous kinetic energy that is released almost instantaneously during the impact. The energy released during a typical meteorite impact can be compared to that of “normal” terrestrial processes, such as volcanic eruptions or earthquakes. During a small impact event, which may lead to craters of 5–10 km in diameter, about $10^{24–25}$ ergs ($10^{17–18}$ J) are released, while during formation of larger craters (50–200 km diameter) about $10^{28–30}$ ergs ($10^{21–23}$ J) are liberated (e.g., FRENCH, 1968; KRING, 1993). These data can be compared to the about $6 \cdot 10^{23}$ ergs ($6 \cdot 10^{16}$ J) released over several months during the 1980 eruption of Mount St. Helens, or 10^{24} ergs (10^{17} J) of the big San Francisco earthquake in 1906. It may also be surprising that the total annual energy release from the earth (including heat flow, which is by far the largest component, volcanism, and earthquakes) is about $1.3 \cdot 10^{28}$ ergs ($1.3 \cdot 10^{21}$ J/y) (FRENCH, 1968; SCLATER et al., 1980; MORGAN, 1989). The latter amount of energy is comparable to the energy that is released almost instantaneously during large impact events (however, it has to be considered that in an impact this huge amount of energy is released at a very small spot on the earth’s surface).

The most important changes in the target rocks occur during the compression stage, while the morphology of a crater is defined in the second and third stage. These processes are well described in the literature (e.g., GRIEVE 1987, 1991; MELOSH, 1989; and references therein). The most important phenomenon, which is characteristic of impact, is the generation of a supersonic shock wave that is propagated into the target rock. Matter is being accelerated very rapidly and, as a consequence of the decrease of compressibility with increasing pressure, the resulting stress wave will become a shock wave moving at supersonic speed. Material is moving behind the shock front at somewhat lower velocity. In nature, a shock front has a finite extent and is trailed by a rarefaction wave, which gradually overtakes the shock front and causes a decrease in pressure with increasing distance of propagation. The shock wave leads to compression of the target rocks at pressures far above a material property called the Hugoniot elastic limit. The Hugoniot elastic limit (HEL) can generally be described as the maximum stress that a material can be subjected to, while above this limit plastic, or irreversible, distortions occur in the solid medium through which the compressive wave travels (see, e.g., compilations by RODDY et al., 1977; MELOSH, 1989; and references therein). The value of the HEL is about 5–10 GPa for most minerals and whole rocks. The only known process that produces shock pressures exceeding the HELs of crustal rocks and minerals in nature is impact cratering. In addition to structural changes, phase changes occur as well.

During the excavation phase of crater formation, a deep cavity, called the "transient crater" is formed. The depth of this cavity is the sum of the excavation depth, which is about one-third of the depth of the transient crater (and equal to about one-tenth of the transient crater diameter), plus the amount of downward displacement of the target rocks (e.g., GAULT et al., 1968; RODDY et al., 1977; GRIEVE, 1987; MELOSH, 1989; and references therein). This cavity is unstable, leading to a collapse of the crater walls. Small bowl shaped craters roughly preserve this form, but are partially filled with various types of allochthonous and autochthonous impact breccias. In large craters the cavity floor is unstable and rises rapidly to form a central uplift, followed by slumping of the rim during the post-impact crater modification stage that leads to terracing and a lower depth to diameter ratio compared to simple craters.

5. Recognition of Impact Structures

Several criteria for the recognition and confirmation of impact structures were developed over the past decades. The most important of these characteristics are:

a) evidence for shock metamorphism;
b) crater morphology;
c) geophysical anomalies; and
d) presence of meteorites or geochemical discovery of traces of the meteoritic projectile.

Of the criteria mentioned above, only the presence of diagnostic shock metamorphic effects (and, in some cases, the discovery of meteorites, or traces thereof) provides unambiguous evidence for the impact origin of a certain structure.

However, morphological and geophysical observations are important in providing supplementary – but not confirming – evidence. The morphology that is typical for simple and complex impact structures has been briefly mentioned above (Text-Fig. 3). It should be mentioned, though, that in complex craters the central structural uplift usually contains severely shocked material and is often more resistant to erosion than the rest of the crater. Also, the central uplift usually exposes, as the name suggests, rocks at or near the surface that are normally at greater stratigraphic depths in the area. In old eroded structures the central uplift may be the only remnant of the crater that can be identified.

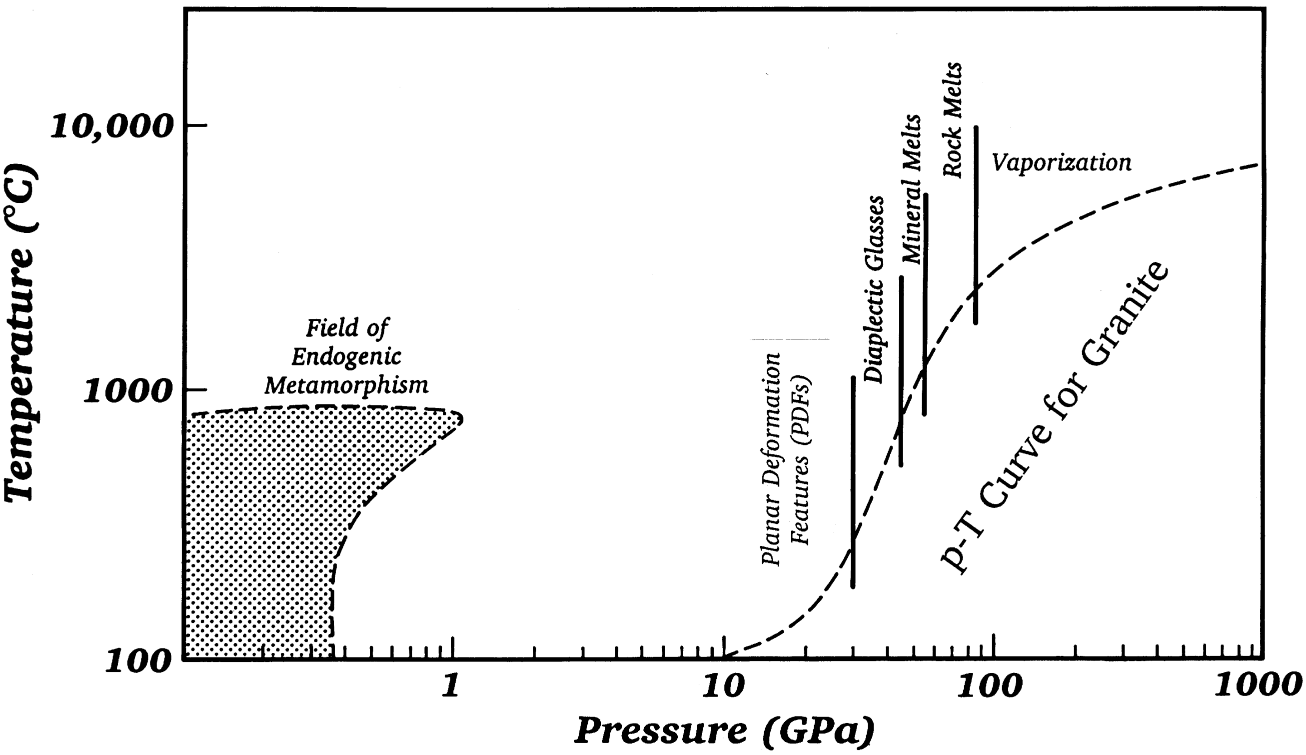
5.1. Geophysical Anomalies

Geophysical studies have been important in the initial discovery of anomalous subsurface structural features. These studies gain importance for deeply eroded craters or for those that are covered by later sediments. The latter category contains several craters in, for example, the United States (e.g., Ames, Avak, Manson, Newporte, Red Wing Creek), Mexico (Chicxulub), and some underwater structures (Montagnais, off-shore Nova Scotia, Canada; Chesapeake Bay, Virginia, USA). Geophysical characteristics of impact craters that have been investigated include gravity, magnetic properties, reflection and refraction seismics, electrical resistivity, and others (see PILKINGTON & GRIEVE, 1992, for a review).

The gravity signature of an impact crater is often rather straightforward, if the structure is not too deeply eroded. Simple craters usually have negative gravity anomalies, as the breccia lens and fractured bedrock have a lower density than unaffected target rocks. Complex craters show more intricate gravity profiles. The central uplift is often associated with a gravity high that is surrounded by an annular gravity low over the breccia zone in the annular trough. Seismic studies, especially reflection seismic surveys, provide important details on the subsurface structure of craters. The discovery of the large 85–90 km diameter underwater Chesapeake Bay crater (POAG et al., 1994), for which an impact origin was recently confirmed (KOEBERL et al., 1995b, 1996), was greatly aided by seismic studies. Magnetic anomalies associated with impact structures are often rather complicated and varied. Large structures may show high-amplitude anomalies due to remanently magnetized target rocks. Recently, ground penetrating radar has been used to study the subsurface distribution of ejecta in or around smaller impact craters (e.g., GRANT & SCHULTZ, 1993). However, geophysical studies alone can not provide confirming evidence for an impact origin.

5.2. Shock Metamorphism

A large meteorite impact produces shock pressures and temperatures of up to many 100 GPa and several 1000°C. This is in contrast to conditions for endogenic metamorphism of crustal rocks, with maximum temperatures of 1200°C and pressures of 2 GPa (Text-Fig. 4); also, strain rates differ by several orders of magnitude. During impact, material can also be subjected to superheating (without being vaporized). Compared to many other natural processes that can be described by thermodynamics, shock compression is not a thermodynamically reversible process and the Hugoniot equations conserve mass, momentum, and energy, but not entropy (see, e.g., review by MELOSH, 1989). Most of the structural and phase changes in mineral crystals and rocks are uniquely characteristic of the high pressures (5–>50 GPa) and extreme strain rates (10^6–10^8 s^{-1}) associated with impact. Static compression, and volcanic or tectonic processes, yield different products because of lower peak pressures and strain rates that are different by more than 11 orders of magnitude. Numerous shock recovery experiments (i.e., controlled shock wave experiments, which allow the collection of the shocked samples for further studies), using various techniques, have been performed in the last three decades, leading to a good understanding of the conditions for formation of shock metamorphic products and a pressure-temperature calibration of the effects of shock pressures up to about 100 GPa (see, e.g., HÖRZ, 1968;



Text-Fig. 4.
Pressure-temperature regime of endogenic metamorphism compared to shock metamorphism.
Also indicated are the onset pressures of various irreversible structural changes in the rocks due to shock metamorphism. The dashed curve in the right part of the diagram shows the relation between pressure and post-shock temperature for shock metamorphism of granitic rocks.
After GRIEVE (1987).

FRENCH & SHORT, 1968; STÖFFLER, 1972, 1974; GRATZ et al., 1992a,b; HUFFMAN et al., 1993; STÖFFLER & LANGENHORST, 1994; and references therein).

Among planetary scientists and impact researchers, it is well established that the presence of rocks and minerals exhibiting evidence for shock metamorphism is an unambiguous indication for the high pressures uniquely associated with impact cratering. However, the literature on the K-T boundary debate in the last decade or so has shown that there is still some incomplete and inadequate understanding in the geological community of the precise nature of diagnostic shock effects (for a discussion, see, e.g., FRENCH, 1990; SHARPTON & GRIEVE, 1990; STÖFFLER & LANGENHORST, 1994). Even the otherwise balanced discussion of the K-T debate by GLEN (1994) largely avoids the discussion of shock metamorphism. It should be reaffirmed that the study of the response of materials to shock is not a recent development, but has been the subject of thorough investigations with a variety of methods over several decades, in part stimulated by military research (see, e.g., FRENCH & SHORT, 1968; STÖFFLER, 1972, 1974; STÖFFLER & LANGENHORST, 1994). As mentioned above, the effects of shock metamorphism are a consequence of the extremely high pressures and strain rates (and, to a lesser extent, temperatures) that the minerals and rocks experienced during an impact event. In contrast to some assertions (e.g., LYONS et al., 1993), the existence of shock metamorphic features in volcanic rocks has never been substantiated (see, e.g., DE SILVA et al., 1990; GRATZ et al., 1992). Table 1 lists the most characteristic products of shock metamorphism, as well as the associated diagnostic features.

A wide variety of macroscopic and microscopic shock metamorphic effects has been recognized, depending upon the peak shock pressure experienced. Observations of naturally and experimentally shocked rocks have enabled calibration of the pressure ranges for the occurrence of the different shock features. A good macroscopic indicator of shock effects is the occurrence of shatter cones (e.g., DIETZ, 1968; MILTON, 1977). Such cones have also been formed in explosion crater experiments. Their formation is dependent on the type of target rock and has estimated to take place at pressures in the range of 2 to 30 GPa. In general, shatter cones are cones with regular thin grooves that radiate from the top. They can range in size from less than one centimeter to more than one meter (Text-Fig. 5a,b). Unfortunately, no definitive criteria for the recognition of "true" shatter cones have yet been defined. If they are strongly eroded, it is possible to confuse concussion features, pressure-solution features (cone-in-cone structure) or abraded or otherwise striated features with shatter cones. It would be important to arrive at some generally accepted criteria for the correct identification of shatter cones, as some impact craters have been identified almost exclusively by the occurrence of shatter cones. However, shatter cones are good indicators for structures that need more research.

The best and most generally accepted indicators for shock metamorphism are features that are only visible at the microscopic level. Various shock effects, such as planar microstructures, optical mosaicism, changes in refractive index, birefringence, and optical axis angle, isotropization, and phase changes, can be discerned by studying thin sections using the polarizing optical microscope. Several features (mostly microscopic) that are diagnostic for shock are described in Table 1. Shock effects in the low pressure regime lead to crystalline and partly amorphous states, such as fracturing, mosaicism, planar fractures, and planar deformation features. For example, mosaicism is characterized as an irregular mottled optical

Text-Fig. 5.
a) Assemblage of shatter cones, with sizes between about 10 and 60 cm, exposed in limestone, at the Kentland crater in Indiana, USA.
b) Large meter-sized complex shatter cone, cut at the quarry at the Kentland crater, Indiana, USA, with the author as scale bar.

extinction pattern, which is distinctly different from undulatory extinction that occurs in tectonically deformed quartz. Mosaicism can be measured in the optical microscope, or, preferably, by X-ray diffraction. Many of these effects occur in most or all rock forming, and some accessory, minerals. However, the most commonly and thoroughly studied mineral in respect to shock effects is quartz (STÖFFLER & LANGENHORST, 1994).

Planar deformation features (PDFs) in rock forming minerals (quartz, feldspar, or olivine) are generally accepted to be diagnostic evidence for shock (see, e.g., FRENCH & SHORT, 1968; STÖFFLER, 1972, 1974; ALEXOPOULOS et al., 1988; SHARPTON & GRIEVE, 1990; STÖFFLER & LANGENHORST, 1994). PDFs are parallel zones with a thickness of about $<$1–3 μm that are spaced about 2–10 μm apart (see, e.g., Text-Figs. 6a–d).

It was demonstrated in TEM studies (see GOLTRANT et al., 1991, and STÖFFLER & LANGENHORST, 1994, for details) that PDFs consist of amorphous silica, which is, however, structurally slightly different from regular silica glass. The glass state of PDFs allows them to be preferentially etched by, e.g., HF, amplifying the structure (see Text-Fig. 6c,d).

Rarely PDFs can be curved as a result of post-impact mineral deformation. PDFs, together with the somewhat less definitive planar fractures (PFs), are well developed in quartz (STÖFFLER & LANGENHORST, 1994). They occur in planes corresponding to specific crystallographic orientations, with the (0001) or c (basal), $\{10\bar{1}3\}$ or ω, and $\{10\bar{1}2\}$ or π orientations being the most common in quartz. PDFs practically always occur in more than one crystallographic orientation per grain.

They become more closely spaced and more homogeneously distributed with increasing shock pressure. Depending on the peak pressure, PDFs are observed in 2 to10 (maximum 18) orientations per grain. The crystallographic orientation of PDFs is studied using either a universal or a spindle stage (REINHARD, 1931; EMMONS, 1943), or by transmission electron microscopy (TEM; see, e.g., GOLTRANT et al., 1991; GRATZ et al., 1992a; LEROUX et al., 1994). The optical and TEM properties of PFs and PDFs in quartz are summarized in Table 2. In addition, there is an inverse relationship between the refractive index of a shocked grain with PDFs and the shock pressure in the 25 to 35 GPa range (see STÖFFLER & LANGENHORST, 1994).

The degree of planarity and the crystallographic orientations of the individual sets of PDFs are important parameters for the correct identification of bona fide PDFs, and allow their distinction from planar features produced at a low strain rate, e.g., tectonically deformed quartz.

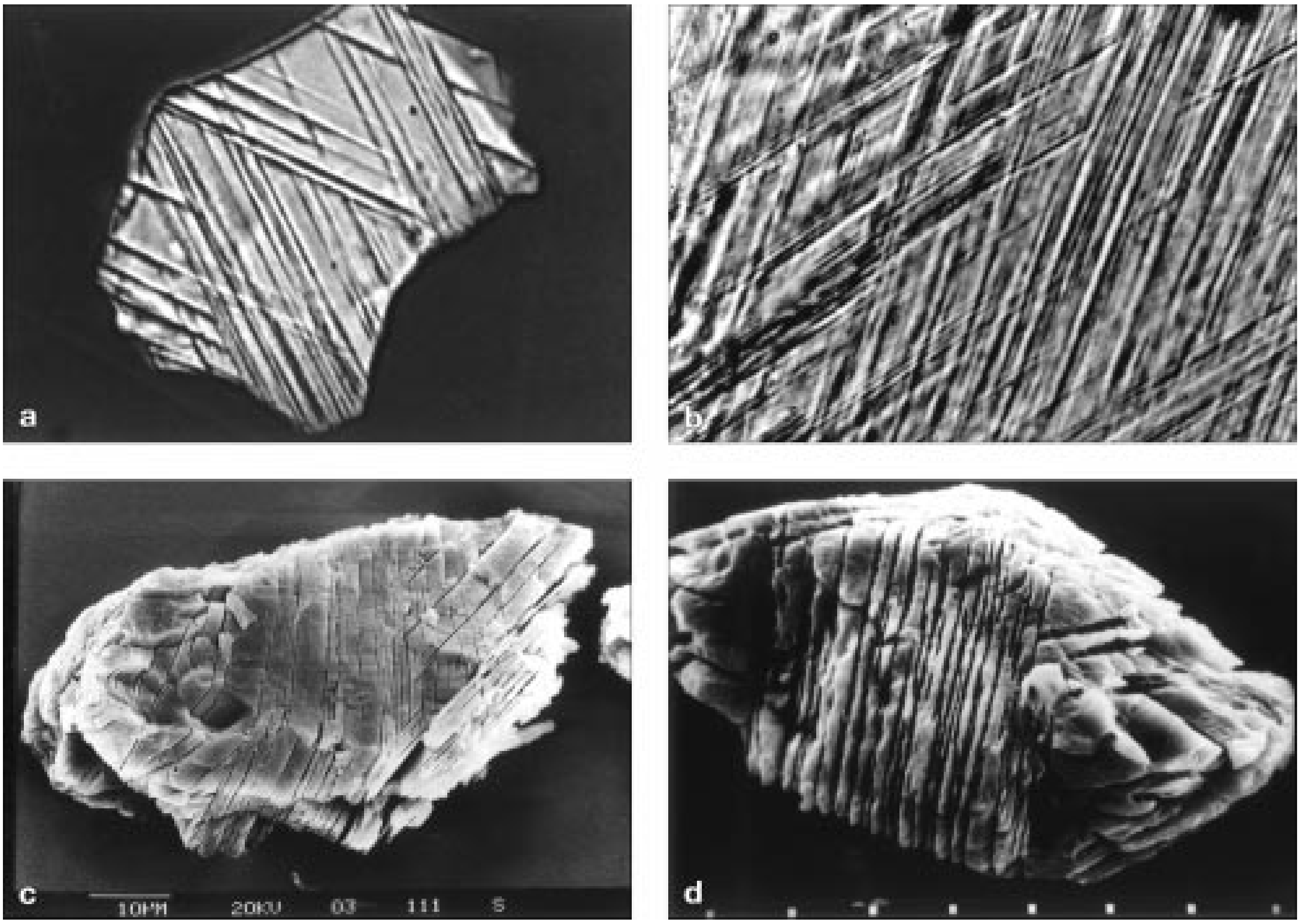
The crystallographic orientations of PDFs and the related shock pressures and optical characteristics are given in Table 2. The relative frequencies of the crystallographic orientations can be used to calibrate shock pressure regimes, as listed in Table 3 (see, e.g., ROBERTSON et al., 1968; HÖRZ, 1968; STÖFFLER & LANGENHORST, 1994). For example, at 5 to 10 GPa, PDFs with (0001) and $\{10\bar{1}1\}$ orientations are formed, while PDFs with $\{10\bar{1}3\}$ orientations start to form between about 10 and 12 GPa. Brazil twins in quartz are structures that can be best studied with TEM techniques. They are always parallel to the (0001) orientation and form either as the result of hydrothermal growth or in shock processes at pressures of about 8 GPa

Table 1.
Shock metamorphic features and their characteristics.
Data from: ALEXOPOULOS et al. (1988), FRENCH & SHORT (1968), SHARPTON & GRIEVE (1990), STÖFFLER (1972, 1974), KOEBERL et al. (1995a).
After KOEBERL (1994).

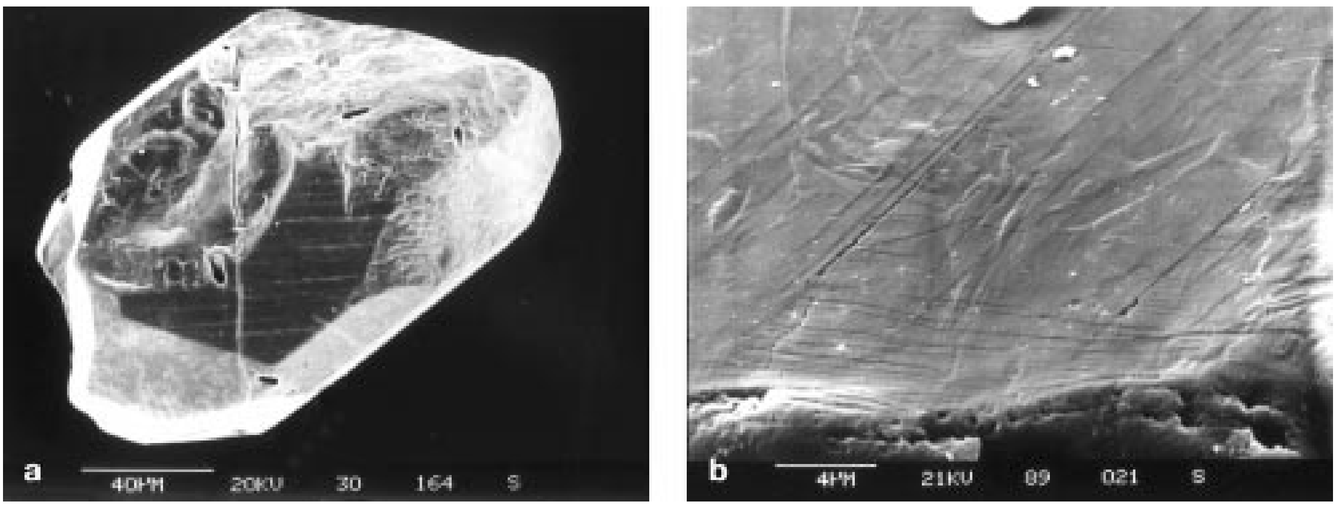
Pressure Range (GPa)	Features	Target Characteristics	Feature Characteristics
2-30	Shatter cones	Best developed in homogeneous, fine-grained, massive rocks	Conical fracture surfaces with subordinate striations radiating from a focal point.
5-45	Planar fractures and Planar deformation features (PDFs)	Highest abundance in crystalline rocks; Found in many rock-forming minerals; e.g., quartz, feldspar, olivine, and zircon.	Sets of extremely straight, sharply defined parallel lamellae; occur often in multiple sets with specific crystallographic orientations.
30-40	Diaplectic glass	Most important in quartz and feldspar (e.g., maskelynite from plagioclase).	Isotropization through solid-state transformation under preservation of crystal habit as well as primary defects and sometimes planar features. Index of refraction lower than in crystal but higher than in fusion glass.
15-50	High-pressure polymorphs	Quartz polymorphs most common: Coesite, Stishovite; but also Ringwoodite from olivine, and others.	Recognizable by its crystal parameters, confirmed usually with XRD or NMR; abundance influenced by post-shock temperature and shock duration; Stishovite is temperature-labile.
>35	Impact diamond	From carbon (graphite) present in target rocks; rare.	Cubic and hexagonal form; usually very small but occasionally up to mm-size; inherit graphite crystal shape.
45->70	Mineral melts	Rock-forming minerals (e.g., lechatelierite from quartz)	Contrary to diaplectic glass, complete transformation of a mineral into glass.
>60	Rock melt	Best developed in massive silicate rocks Occur as individual melt bodies (mm to m size) or as coherent melt sheets, up to 1000 km^3.	Impact melts are either glassy (fusion glasses) or crystalline; of macroscopically homogeneous, but microscopically often heterogeneous composition.

Table 2.
Microscopic characteristics of planar fractures and planar deformation features in quartz.
Data after STÖFFLER & LANGENHORST (1994).

Nomenclature	1. Planar fractures (PF) 2. Planar deformation features (PDF) 2.1. Non-decorated PDFs 2.2. Decorated PDFs
Crystallographic orientation	1. PFs: usually parallel to (0001) and $\{10\bar{1}1\}$ 2. PDFs: usually p to $\{10\bar{1}3\}$, $\{10\bar{1}2\}$, $\{10\bar{1}1\}$, (001), $\{11\bar{2}2\}$, $\{11\bar{2}1\}$, $\{10\bar{1}0\}$, $\{11\bar{2}0\}$, $\{21\bar{3}1\}$, $\{51\bar{6}1\}$, etc.
Optical microscope properties	Multiple sets of PFs or PDFs (up to 15 orientations) per grain Thickness of PDFs: <2 - 3 µm Spacing: >15 µm (PFs), 2 - 10 µm (PDFs)
TEM properties (PDFs)	Two types of primary lamellae are observed: 1. amorphous lamellae with a thickness of ca. 30 nm (at pressures <25 GPa) and ca. 200 nm (at pressures >25 GPa) 2. Brazil twin lamellae parallel to (0001)



Text-Fig. 6.
Shocked quartz.
a) Shocked quartz grain with two sets of PDFs, from the K-T boundary layer at Teapot Dome, Wyoming; longest dimension of grain 0.07 mm, crossed polars.
b) Close- up of quartz grain from the K-T boundary at Starkville South, Colorado, showing two sets of PDFs and the closely spaced nature of the lamellae; width of image 100 µm, crossed polars.
c) SEM image of quartz grain from K-T boundary layer at DSDP Site 596 (Southwest Pacific), after brief etching with HF, showing at least three different sets of PDFs.
d) SEM image of shocked quartz grain from the K-T boundary layer at Brownie Butte, Montana, after brief etching with HF, showing three different sets of PDFs; distance between scale bars = 10 µm.
Courtesy B. Bohor (U.S. Geological Survey).



Text-Fig. 7.
SEM images of etched shocked zircons from the Berwind Canyon (Raton Basin, Colorado, USA) K-T boundary sections.
a) whole grain, displaying PDFs in two different orientations.
b) close-up of surface, showing two intersecting sets of PDFs.
Courtesy B. Bohor (U.S. Geological Survey).

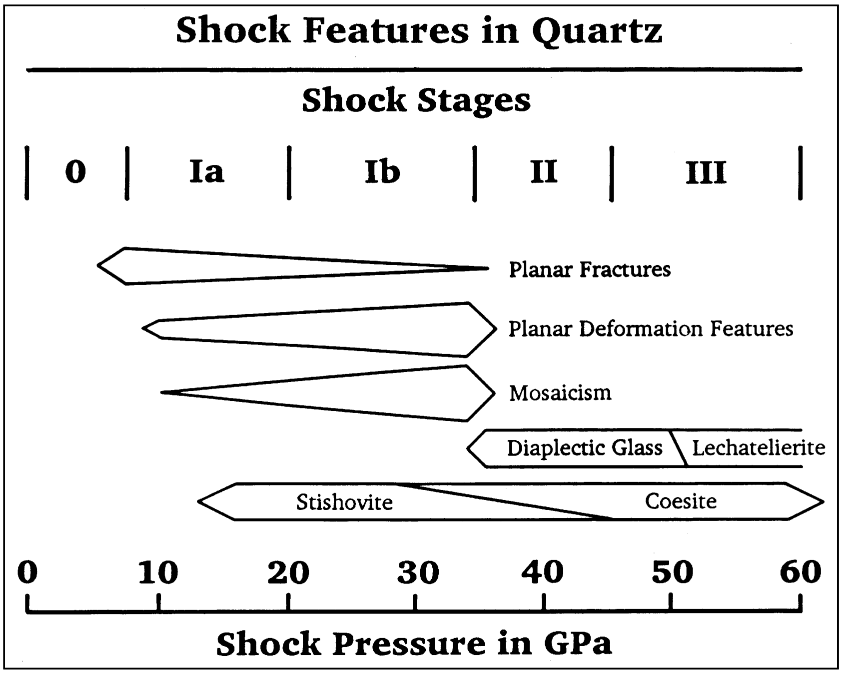
Table 3.
Relation between shock stage and crystallographic orientation (indices) of planar microstructures in quartz.
PF = planar fractures; PDF = planar deformation features.
After STÖFFLER & LANGENHORST (1994).

Shock stage	*Main orientations*	*Additional orientations*	*Optical properties*
1. Very weakly shocked	PFs: (0001)	PFs: rarely $\{10\bar{1}1\}$ PDFs: none	normal
2. Weakly shocked	PDFs: $\{10\bar{1}3\}$	PFs: $\{10\bar{1}1\}$, (0001) PDFs: rare	normal
3. Moderately shocked	PDFs: $\{10\bar{1}3\}$	PFs: $\{10\bar{1}1\}$, (0001) rare PDFs: $\{11\bar{2}2\}$, $\{11\bar{2}1\}$, (0001), $\{10\bar{1}0\}+\{11\bar{2}1\}$, $\{10\bar{1}1\}$, $6\{21\bar{3}1\}$, $\{51\bar{6}1\}$	normal or slightly reduced refractive indices
4. Strongly shocked	PDFs: $\{10\bar{1}2\}$ $\{10\bar{1}3\}$	PFs: rare or absent PDFs: $\{11\bar{2}2\}$, $\{11\bar{2}1\}$, (0001),$\{10\bar{1}0\}+\{11\bar{2}1\}$, $\{10\bar{1}1\}$, $\{21\bar{3}1\}$, $\{51\bar{6}1\}$	reduced refractive indices 1.546-1.48
5. Very strongly shocked	PDFs: $\{10\bar{1}2\}$ $\{10\bar{1}3\}$	none	reduced refractive indices (<1.48)

– two mechanisms that are easily distinguishable. Most rock forming minerals, as well as accessory minerals, such as zircon (Text-Fig. 7), develop PDFs. The occurrence of diagnostic shock features is by far the most important criterion for evaluating the impact origin of a crater, particularly when several of the features that are typical of progressive shock metamorphism, as listed in Table 1, have been found.

At shock pressures in excess of about 35 GPa, diaplectic glass is formed. This is an amorphous, isotropic phase that preserves the crystal habit and, in some cases, planar features. It forms without melting and is a stage intermediate between crystalline and normal glassy phases (STÖFFLER & HORNEMANN, 1972). Diaplectic quartz glass has been found at numerous impact craters and has a refractive index that is slightly lower, and a density that is slightly higher, than that of synthetic quartz glass. Minerals other than quartz show the same behavior. For example, feldspar is being converted to the diaplectic glass maskeylinite.

At pressures that exceed about 50 GPa, a lechatelerite, a mineral melt, forms by fusion of quartz. Other minerals also undergo melting at similar pressures. This complete melting is not the same process that results in the formation of diaplectic glass. The differences between diaplectic glass and lechatelierite (both after quartz) are described by STÖFFLER & HORNEMANN (1972) and STÖFFLER &



Text-Fig. 8.
Pressure dependency of various characteristic shock indicators in quartz, and relation to shock stages.
After STÖFFLER & LANGENHORST (1994).

LANGENHORST (1994). Another form of shock effect is the formation of high-pressure polymorphs of minerals, e.g., stishovite and coesite from quartz. In contrast to expectations from the equilibrium phase diagram of quartz, stishovite forms at lower pressures than coesite, probably because stishovite forms directly during shock compression, while coesite crystallizes during pressure release. The formation probabilities and conditions for these phases are strongly dependent of the porosity of the target rocks. In general, the occurrence of various types of shock effects can be used to estimate the maximum shock pressure for a certain rock, as shown in Text-Fig. 8.

At pressures in excess of about 60 GPa rocks undergo melting to form whole-rock impact melts (mineral melts can form at slightly lower pressures – see Table 1). The melts can attain very high temperatures as a result of the passage of shock waves that generate temperatures far beyond those commonly encountered in normal crustal processes or in volcanic eruptions. This is shown by the presence of inclusions of high-temperature minerals, such as lechatelierite, which forms from pure quartz at temperatures >1700°C (see also above), or baddeleyite, which is the thermal decomposition product of zircon, forming at a temperature of about 1900°C. Impact melts may also undergo a phase of superheating (i.e., staying liquid even though the vaporization temperature has been exceeded) at temperatures of 10,000°C or higher (e.g., JAKES et al., 1992). Depending on the initial temperature, the location within the crater, the composition of the melt, and the speed of cooling, impact melts either form impact glasses (if they cool fast enough), or, more commonly, very fine-grained impact melt rocks (if they cool slower). Glasses are metastable supercooled liquids, and, thus, impact glasses slowly recrystallize (if dissolution is not acting faster), at a rate that depends on the composition of the glass and post-impact depositional conditions. This explains why impact glasses are usually found at young impact craters, but are relatively rare in old structures. Impact glasses have chemical and isotopic compositions that are very similar to those of individual target rocks or mixtures thereof. For example, it is possible to use the rare earth element (REE) distribution patterns, or the Rb-Sr isotopic composition, which are identical to those of the (often sedimentary or metasedimentary) target rocks, to distinguish them from any intrusive or volcanic rocks.

Impact glasses are characterized by low water contents (only if the glass is not altered; about 0.001–0.05 wt.-%), inhomogeneous chemical compositions (at the 0.1 mm–cm scale), the preservation of (sometimes shocked) minerals from the target rocks, the presence of high-temperature decomposition phases, such as baddeleyite, and (sometimes) an indication for the admixture of a small meteoritic component. Detailed descriptions of impact melts and glasses and their characteristics and compositions are discussed by, for example, EL GORESY et al. (1968), DENCE (1971), STÖFFLER (1984), KOEBERL (1986, 1992a,b), and references therein. Impact glasses and, more commonly, impact melts (i.e., recrystallized material) are also found in breccias in the form of melt clasts, e.g., in suevitic breccias (see below). Impact melts (or impact melt breccias) are much more common than impact glasses, because the glass is not stable over long periods of time.

In the crater, shocked minerals, impact melts, and impact glasses are usually found in various impact-derived breccias. In outcrops or drill cores, the presence of a layer of fragmental breccia as crater fill or overlying a possibly raised, partially brecciated, and up- or overturned rim is good structural evidence for impact. Ejecta at the crater rim may display a stratigraphic sequence that is inverted compared to the normal stratigraphy in the area. The youngest target rock at the top of the target sequence are ejected and deposited first, followed by older target rocks. The impact process leads to the formation of various breccia types, which are found within and around the resulting crater (see also STÖFFLER & GRIEVE, 1994): monomict or polymict breccias consisting of

1) cataclastic (fragmental),
2) suevitic (fragmental with a melt fragment component), or
3) impact melt (melt breccia – i.e., melt in the matrix – with a clastic component) breccias.

The breccias can be allochthonous or autochthonous. In addition, dikes of fragmental breccia or pseudotachylite (which contains evidence of melting) can be found in the basement rocks. The schematic distribution of breccias, melt, and breccia dikes at simple and complex craters is shown in Text-Fig. 3. Whether all these breccia types are actually present at an impact crater depends on factors including the size of the crater, the composition and porosity of the target area (e.g., KIEFFER & SIMONDS, 1980), and the level of erosion (for more details see, e.g., RODDY et al., 1977; HÖRZ, 1982; HÖRZ et al., 1983; GRIEVE, 1987; and references therein).

5.3. Remnants of the Meteoritic Projectile

At most meteorite impact craters, no meteorites are found. At first this observation may seem surprising, but it follows as a logical consequence of the physics of an impact event. A shock wave, similar to the one that penetrates through the target, also passes through the meteoritic impactor and, within fractions of a second, vaporizes most or all of the projectile. Only during the impact of small objects (less than about 40 m in diameter, depending on impact angle and velocity), as a result of spallation during entry into the atmosphere or due to lower impact velocity resulting from atmospheric drag on even smaller objects (<10 m diameter), a small fraction of the initial mass of the projectile material survives. The cut-off diameter of impact craters at which meteorite fragments may be preserved is about 1–1.5 km, but even at such small craters (e.g., Meteor Crater [Barringer], Arizona; Odessa, Texas; Wabar, Saudi Arabia; Wolf Creek, Australia) only a few percent of the original projectile mass have survived. The preservation of meteorites is limited as a result of the low resistivity of meteoritic matter to erosion. Under normal terrestrial climatic conditions, stony meteorites last only a few thousand years, while iron meteorites may resist about ten times longer. Thus, even under optimistic conditions, meteoritic fragments are only preserved at very young and small craters. The absence of meteorite fragments can, therefore, not be used as evidence against an impact origin of a crater structure.

A potentially more powerful impact-diagnostic method is the detection of traces of the meteoritic projectile in target rocks, which allows to establish the impact origin for a crater structure. During the first phases of the impact crater formation, the meteoritic projectile undergoes vaporization; a small amount of the meteoritic vapor is incorporated with the much larger quantity of target rock vapor and melt, which later forms impact melt rocks, melt breccias, or glass. In most cases, the contribution of meteoritic matter to these impactite lithologies is ≪1 wt.-%. Thus, any chemical changes resulting from the admixture of extraterrestrial material are barely noticeable, and only elements that have high abundances in meteorites, but low abundances in terrestrial crustal rocks, can be used to detect such a meteoritic component. During the last two decades, studies of the abundances and interelement ratios of the siderophile elements, such as Cr, Co, Ni, and, especially, the platinum group elements (PGEs) have been used for these investigations (see, e.g., MORGAN et al., 1975; PALME, 1982; EVANS et al., 1993; and references therein). However, the use of elemental abundances does not necessarily lead to unambiguous results. The incorporation of mantle-derived ultramafic rocks or ore minerals that are present among the target rocks may result in elevated PGE abundances and near-chondritic interelement ratios, mimicking an extraterrestrial component.

The use of the Re-Os isotopic system, in contrast to elemental abundances of the PGEs, has numerous advantages. It is superior in detection limit and selectivity as discussed by KOEBERL & SHIREY (1993) and KOEBERL et al. (1994a). In principle, the abundances of Re and Os and the ^{188}Os/^{187}Os isotopic ratios, which are measured by very sensitive mass spectrometric techniques, allow to distinguish the isotopic signatures of meteoritic and terrestrial Os. Meteorites (and the terrestrial mantle) have much higher (by factors of 10^4–10^5) PGE contents than terrestrial crustal rocks. In addition, meteorites have relatively low Re and high Os abundances, resulting in Re/Os ratios less or equal to 0.1, while the Re/Os ratio of terrestrial crustal rocks is usually no less than 10. Even more important, the ^{188}Os/^{187}Os isotopic ratios for meteorites and terrestrial crustal rocks are significantly different.

Due to the high Os abundances in meteorites, adding even a minute meteoritical contribution to crustal target rocks leads to an almost complete resetting of the Os isotopic signature of the resulting impact melt or breccia because 99% or more of the Os in the breccia originates from the meteorite. More details of this method were discussed by KOEBERL & SHIREY (1993; 1996a,b) and KOEBERL et al. (1994a,b). Measurement of the Re-Os isotopic characteristics of both target rocks and impact melt rocks or breccias, together with supplementary chemical and mineralogical information on the samples, provides an excellent indication for impact, which may, under certain conditions, be as distinct as the discovery of shock metamorphic features.

6. The K-T Boundary Impact Event

The short review of the various characteristics of impact craters and impact products allows a perceptive discussion of the nature and origin of the rocks that mark the K-T boundary, and the link with the Chicxulub structure in Mexico. Despite some claims to the contrary, the K-T boundary provides several independent lines of evidence in favor of the conclusion that an impact event was responsible for the end of the Cretaceous. In the following paragraphs I am summarizing the most important arguments in favor of an impact origin; however, space limitations prevent a detailed discussion of these arguments (see, e.g., papers in SILVER & SCHULTZ [1982] and SHARPTON & WARD [1990]).

The following observations are indicative of an impact event 65 Ma ago.

6.1. PGE Enrichments

The first physical evidence pointing to a contribution of extraterrestrial material that was discovered was the presence of anomalously high PGE abundances in K-T boundary clay in Italy (ALVAREZ et al., 1980) and other locations around the world (e.g., SMIT & HERTOGEN, 1980; GANAPATHY, 1980; KYTE et al., 1980). Iridium and other PGEs were found to be enriched in these K-T boundary clay layers by up to four orders of magnitude compared to average terrestrial crustal abundances. While some suggestions were made to explain the PGE enrichments from terrestrial sources, e.g., volcanic emissions or concentration in biological organisms, such models fail to explain the overall amount of PGEs found worldwide at the K-T boundary (see, e.g., KRING [1993] for a more detailed discussion), as well as the interelement ratios (see next paragraph). Text-Fig. 9 shows an Ir abundance profile across a typical K-T boundary section, indicating the significant Ir enrichment at the boundary.

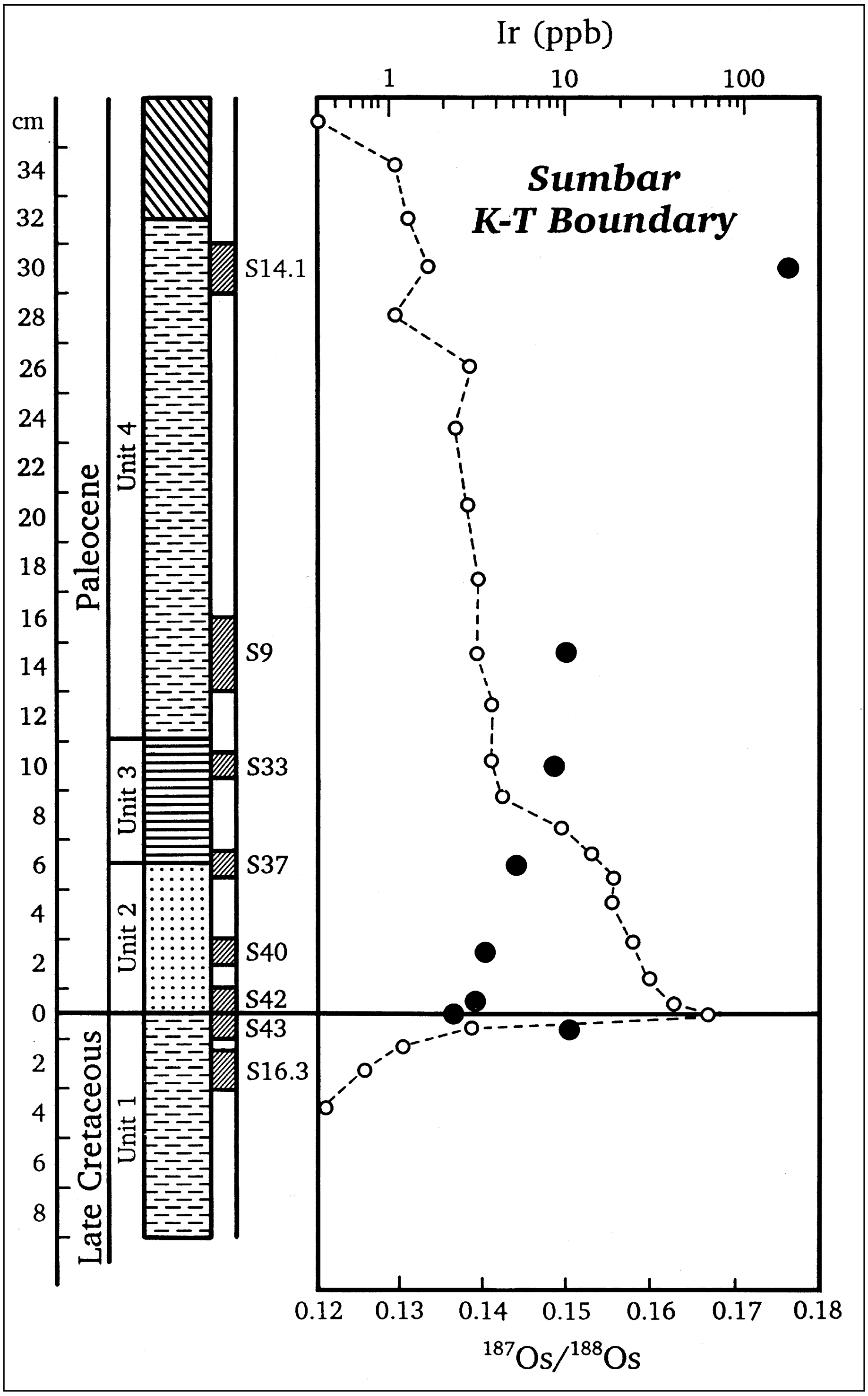
6.2. Near-Chondritic Interelement Ratios of PGEs

It was found that the interelement ratios of the PGEs in the K-T boundary clays are very similar to the values observed in chondritic meteorites (e.g., ALVAREZ et al., 1980; GANAPATHY, 1980; KYTE et al., 1980; PALME, 1982; TREDOUX et al., 1989). Terrestrial sources do not easily explain cosmic interelement ratios.

6.3. Meteoritic Os-Isotopic Signature

TUREKIAN (1982) proposed to measure the Os isotopic ratio in the K-T boundary clays, following the reasoning discussed above. LUCK & TUREKIAN (1983) subsequently analyzed marine manganese nodules and two K-T boundary clay samples. The manganese nodules had ^{187}Os/^{188}Os ratios of about 0.7 to 1, showing a clear continental crustal signature, while the K-T boundary samples from Stevns Klint (Denmark) and Starkville South (Colorado, USA) yielded values of 0.200 and 0.155, respectively (LUCK & TUREKIAN, 1983; see also KOEBERL & SHIREY, 1996b). These results indicate clearly that the PGE signature at the K-T boundary is not the result of the concentration of PGEs from a crustal source by terrestrial processes. Subsequent analyses by LICHTE et al. (1986) showed a ^{187}Os/^{188}Os ratio of 0.135 in clay for the Woodside Creek (New Zealand) K-T boundary; and KRÄHENBÜHL et al. (1988), ESSER & TUREKIAN (1989), SCHMITT (1990), and MEISEL et al. (1993) found comparable values, ranging from 0.137 to 0.212, for K-T boundary samples from Starkville South, Madrid and Berwind Canyon (Colorado, USA), Raton (New Mexico, USA), Shatsky Rise (DSDP 577), Stevns Klint, and Sumbar (Turkmenistan). PEUCKER-EHRENBRINK et al. (1994) found that the ^{187}Os/^{188}Os ratio of seawater at the K-T boundary shows a sharp decrease, which they attributed to leaching of meteorite debris from the impact event. Recently, MEISEL et al. (1995) measured the variation of the ^{187}Os/^{188}Os ratio across a K-T boundary section (at Sumbar, Turkmenistan) and found a significant and sudden decrease of the ^{187}Os/^{188}Os ratio from the end-Cretaceous rock layers to the actual K-T boundary clay that correlates with the maximum Ir (and Os) concentration (Text-Fig. 9); in the early Tertiary rocks, the ^{187}Os/^{188}Os ratio returns to higher values. These results provide clear evidence for an extraterrestrial component at the K-T boundary. Unfortunately, despite the potential

Text-Fig. 9.
Profile of the distribution of Ir and the variation of the Os-isotopic composition across a K-T boundary profile in Sumbar, Turkmenistan.
After MEISEL et al. (1993, 1995).



of these analyses, no other K-T boundary sections, or any other boundary sections, have yet been studied.

6.4. Soot (Carbon Black)

WOLBACH et al. (1985) found evidence for global wildfires in the form of a charcoal and soot layer at numerous K-T boundaries around the world, coinciding with the Ir-rich layer. The insoluble carbon fraction after acid dissolution is dominated by kerogen and elemental carbon. Kerogen is enriched about 15 times, and nitrogen is enriched 20 times, compared to the maximum abundances in the uppermost Cretaceous limestones (GILMOUR et al., 1990). Both also show a marked change in their isotopic composition across the K-T boundary. No comparable soot enrichments of local or global distribution occur in the Late Cretaceous or in a wide range of other marine sediments (WOLBACH et al., 1990). The presence of the hydrocarbon retene in the soot layer is diagnostic of wood fires in which resinous (coniferous) plants and trees were burning, indicating that most or all of the fuel was biomass (WOLBACH et al., 1990). The isotopic composition of the carbon in the soot layer (average $\delta^{13}C = -25.8$ ‰) resembles that of natural charcoal and atmospheric carbon particles originating from biomass fires (WOLBACH et al., 1990). The total amount of soot in the atmosphere due to the global wildfires at the end of the Cretaceous has been estimated at $7 \cdot 10^{16}$ g, which must have had a large influence on the environment.

6.5. Evidence of Shock Metamorphism

The first clear evidence of shock metamorphism at the K-T boundary was found by BOHOR et al. (1984) in the form of shocked quartz grains in sediments from the Brownie Butte location. This landmark discovery confirmed the hypothesis of ALVAREZ et al. (1980) that a large-scale impact event occurred at the end of the Cretaceous.

Shocked quartz (Text-Fig. 6) and feldspar, and other shocked minerals (Text-Fig. 7) were later found at practically all K-T boundary sites around the world (see, e.g., BOHOR et al., 1987; BOHOR, 1990). The shocked quartz grains show multiple intersecting sets of PDFs with shock-characteristic crystallographic orientations.

As discussed above, such shocked minerals are associated with pressures far beyond those of any endogenic processes and are uniquely characteristic of hypervelocity impact. Furthermore, modal abundance of shocked minerals, as well as the composition of shocked feldspars at the K-T boundary, are indicative of a continental crustal source, and are not compatible with material derived from oceanic crust (SHARPTON et al., 1990).

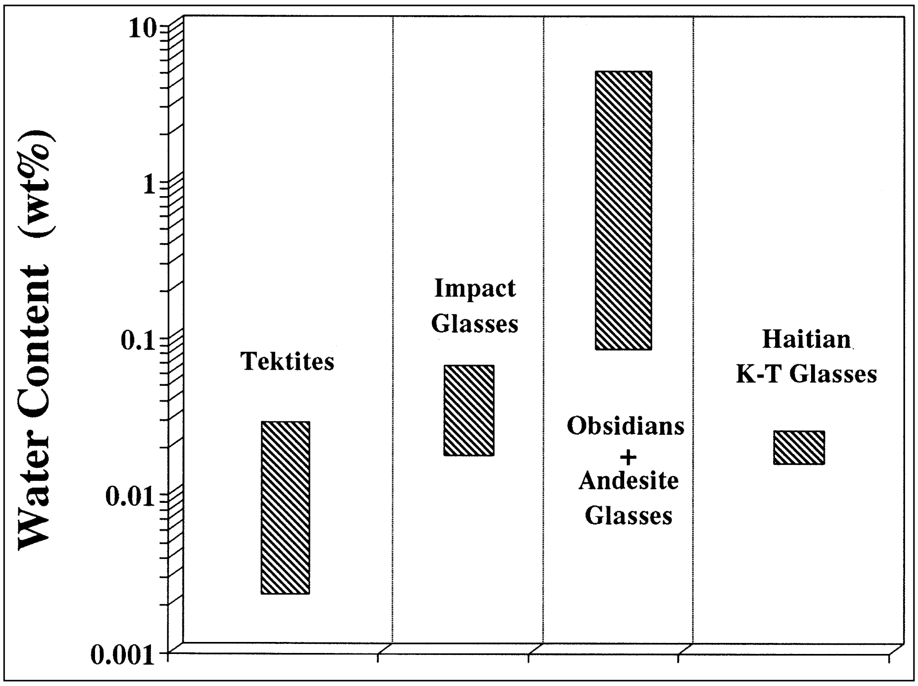
In addition to the finding of shocked minerals, the high-pressure quartz polymorph stishovite was also found in K-T boundary sediments (MCHONE et al., 1989).

6.6. Impact Glass

SIGURDSSON et al. (1991a,b), IZETT (1991), and KRING & BOYNTON (1991) have described the presence of relict glass within alteration spherules (that are otherwise very common at numerous K-T boundary locations) from the K-T boundary at Beloc, Haiti, and interpreted the material as impact glass. SIGURDSSON et al. (1992) have shown, from comparison with experimental glasses, that the Haitian glasses have been quenched from temperatures much higher than common for volcanic processes. A detailed geochemical study by KOEBERL & SIGURDSSON (1992) provided not only detailed geochemical arguments for the impact origin of these glasses, but also demonstrated the existence of rare inhomogeneous glasses with lechatelierite and other mineral inclusions, which are typical for an origin by impact. BLUM & CHAMBERLAIN (1992) have obtained oxygen isotope data on Haitian glasses that specifically rule out a volcanic origin of these glasses. BLUM et al. (1993) have confirmed this result with Rb-Sr and Sm-Nd isotopic data, showing that the Haitian glasses are mixtures of silicate rocks of upper crustal composition with a high CaO-endmember (e.g., limestone). CHAUSSIDON et al. (1994) have shown that the sulfur in the yellow glasses occurs in the form of sulfate, which is not compatible with a volcanic source. KOEBERL (1992b) measured the water content in glasses from Haiti and found a range of 0.013 to 0.021 wt.-% H_2O, which is further evidence for an origin by impact, as impact glasses are extremely dry (see Text-Fig. 10). KOEBERL et al. (1994c) have used Re-Os isotope systematics to find evidence for the presence of a small meteoritic component in the Haitian glasses. Furthermore, high-precision age determinations on the Haitian glasses have shown that the materials have an age indistinguishable from that of the K-T boundary, at 65 Ma (e.g., IZETT et al., 1991; SWISHER et al., 1992). Glasses with similar properties have later been recovered from some K-T boundary locations in Mexico (e.g., Mimbral) as well. All available evidence supports an impact, and not a volcanic, origin for these glasses.

6.7. Impact-Derived Diamonds

Small, nanometer sized diamonds were first reported from K-T boundary sediments in Alberta, Canada (CARLISLE & BRAMAN, 1991). More recently, larger diamonds have also been found at other K-T boundary locations (GILMOUR et al., 1992), including some in Mexico (HOUGH at al., 1995). These diamonds, which have a unique C and N isotopic signature, are clearly connected to the impact process and are somewhat similar to diamonds found at known impact craters (e.g., KOEBERL et al., 1995a).



Text-Fig. 10.
Water content of tektites, impact glasses, volcanic glasses, and glasses from the Haitian K-T boundary at Beloc, indicating the similarity in water content between the Haitian glasses and impact glasses.
The low water content of the Haitian glasses is very good evidence for their impact origin.
After KOEBERL (1992b).

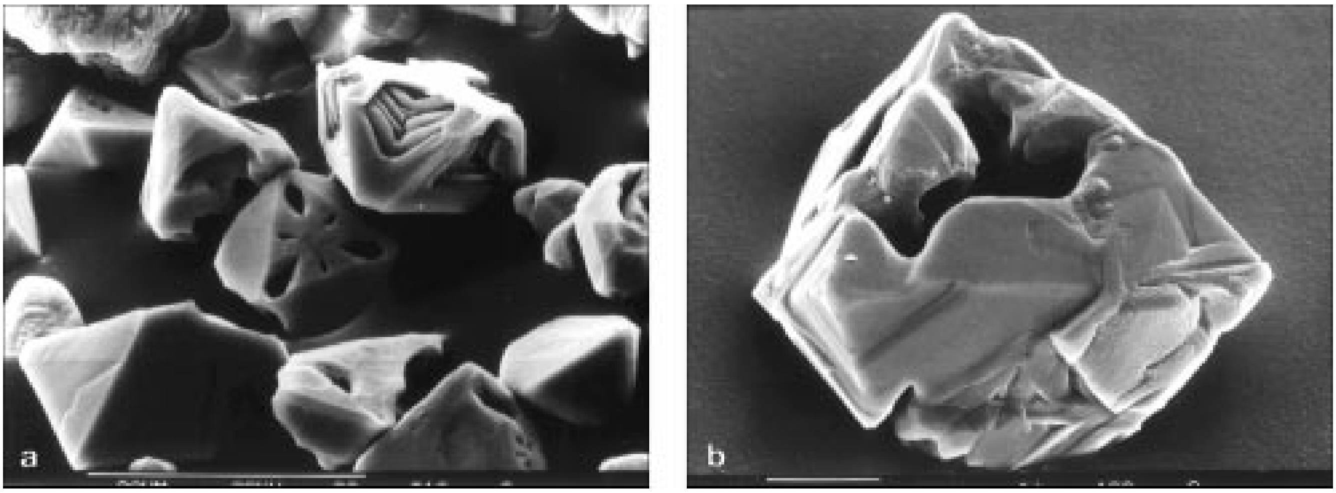
6.8. Occurrence of Spinel

Several varieties of spinel (magnesioferrite) at the K-T boundary (from magnetic spherules at the Petriccio, Italy, section) were first reported by MONTANARI et al. (1983). Spinels at the K-T boundary can be used as event markers, with abundance peaks similar to those observed for the PGEs. These spinels occur in a variety for morphologies (Text-Fig. 11), are all highly oxidized (high Fe^{3+} content), and have high Ni, Co, and Ir contents, but low Cr and Ti abundances. Numerous studies have been performed on the spinels, leading to their interpretation as condensation products from meteoritic vapor or, less likely, direct ablation products from the meteoritic bolide (see, e.g., KYTE & BOSTWICK [1995] and references therein).

The list of arguments supporting an impact event at the K-T boundary is not restricted to the eight points listed above, although they probably represent the most important evidence. It is also important to realize that, while some unusual and extremely rare “endogenic” scenarios might be constructed to explain some of these individual arguments, all these points together provide rather straightforward evidence for an impact event.

7. Search for an Impact Crater

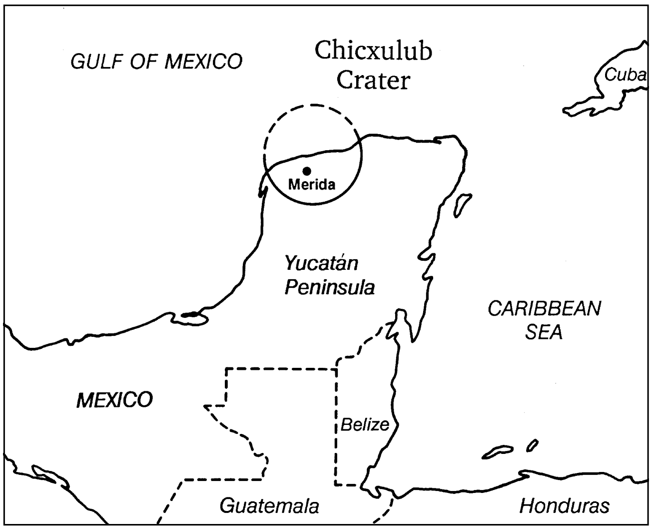
In the late 1980s, the evidence for a K-T boundary impact was already substantial, but the impact hypothesis was still handicapped by lack of a large 65 Ma old impact crater. Crater sizes, as derived from estimates of total meteoritic material present at the K-T boundary worldwide (e.g., ALVAREZ et al., 1980), were required to be no less than about 100–200 km. Several craters were proposed, but a link with the K-T boundary events remained tenuous. Numerous rather eccentric proposals were made (see, e.g., KRING [1993] for a discussion). The more serious suggestions included the Kara crater in Russia (estimated at about 65–80 km diameter), with an age that was at first



Text-Fig. 11.
SEM images of spinel (magnesioferrite) at the K-T boundary, which are interpreted as condensation products from meteoritic vapor or direct ablation products from the meteoritic bolide.
a = Assemblage of magnesioferrite grains from the Caravaca (Spain) K-T boundary; scale bar on bottom: 20 µm; b = large single magnesioferrite grain from the SDP Site 596 (SW Pacific) K-T boundary; grain diameter = 45 µm.
Courtesy B. BOHOR (U.S. Geological Survey).

estimated at 65 Ma (NAZAROV et al., 1989), but later shown to be about 72 Ma old (KOEBERL et al., 1990).

Another very likely candidate for a crater of K-T boundary age was the Manson impact structure, which is a well-preserved 36 km diameter complex impact structure (HARTUNG & ANDERSON, 1988). Early attempts to determine the age of the structure led to the belief that the impact occurred at 65.7 Ma (KUNK et al., 1989). A 1991–1992 research core drilling program led to the recovery of 1200 m core from 12 locations (see papers in KOEBERL & ANDERSON, 1996a). Research was directed at obtaining a new and more accurate age for the formation of the Manson structure. $^{40}Ar/^{39}Ar$ age spectrum analysis of sanidine feldspar (recrystallized from impact melt) by IZETT et al. (1993) demonstrated that the Manson structure was formed about 74 Ma, removing Manson from the (short) list of possible K-T boundary craters. Additionally, IZETT et al. (1993) discovered impact materials in the similar-aged Crow Creek Member of the Cretaceous Pierre Shale in South Dakota, which they interpreted as distal ejecta from Manson.



Text-Fig. 12.
Geographical location of the Chicxulub impact structure on the NE part of the Yucatan (Mexico) peninsula.
After KOEBERL (1993).

Detailed studies of impact-derived deposits at K-T boundary locations around the world were used to constrain the location of the possible source crater. For example, taking the abundance and size variation of shocked quartz grains into account, it soon became clear that the source crater had to be somewhere on or near the North American continent (see, e.g., KRING [1993] for a discussion). However, it was not until 1991 that HILDEBRAND et al. (1991) proposed that a large buried structure in NE Yucatan (Mexico; Text-Fig. 12), which had been suggested by PENFIELD & CAMARGO (1981) to be an impact structure, might be the elusive K-T boundary crater.

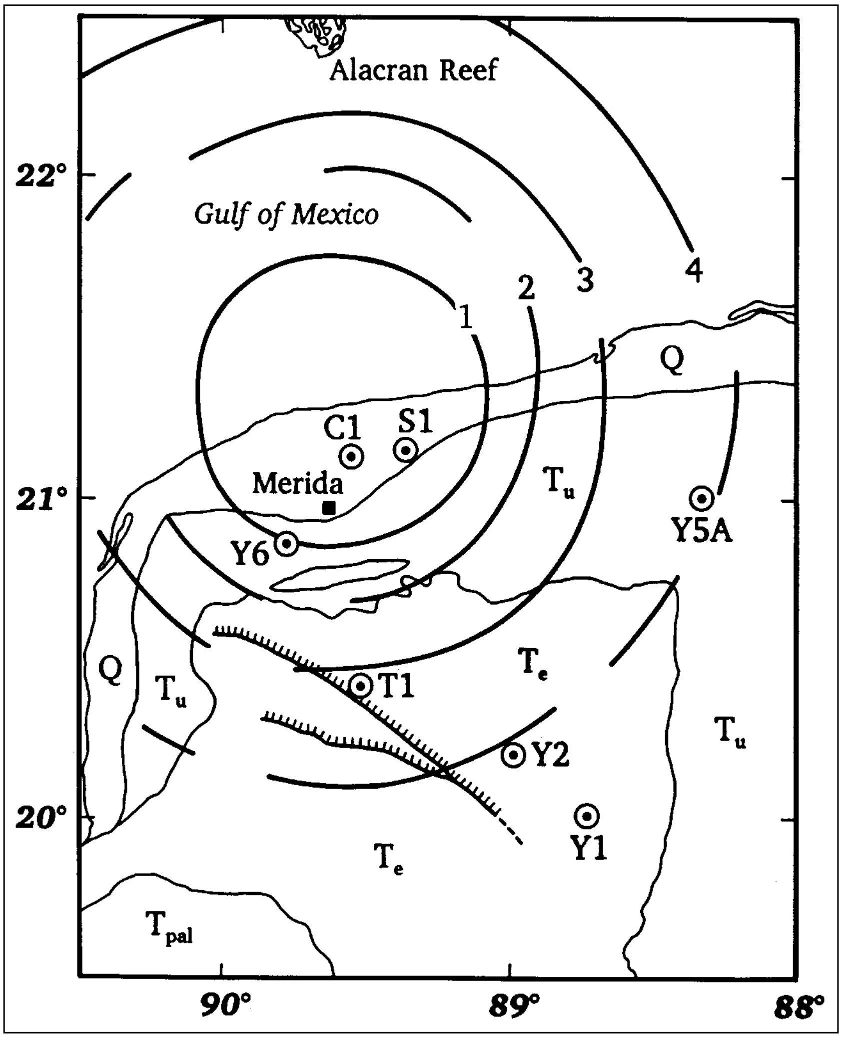
8. The Chicxulub Impact Structure

Several lines of evidence support the interpretation of Chicxulub not only as having an impact origin, but also of being t h e K-T boundary impact crater. While PENFIELD & CAMARGO (1981) noted that the circular geophysical anomaly, which had been known for some time and had been the subject of earlier petroleum exploration drilling, may be due to the presence of a subsurface impact structure, interest in the feature gained momentum only after HILDEBRAND et al. (1991) related it to the K-T boundary extinctions. In the following paragraphs I will attempt a short summary of various key points of evidence in favor of an impact origin of the Chicxulub structure.

8.1. Gravity Anomaly

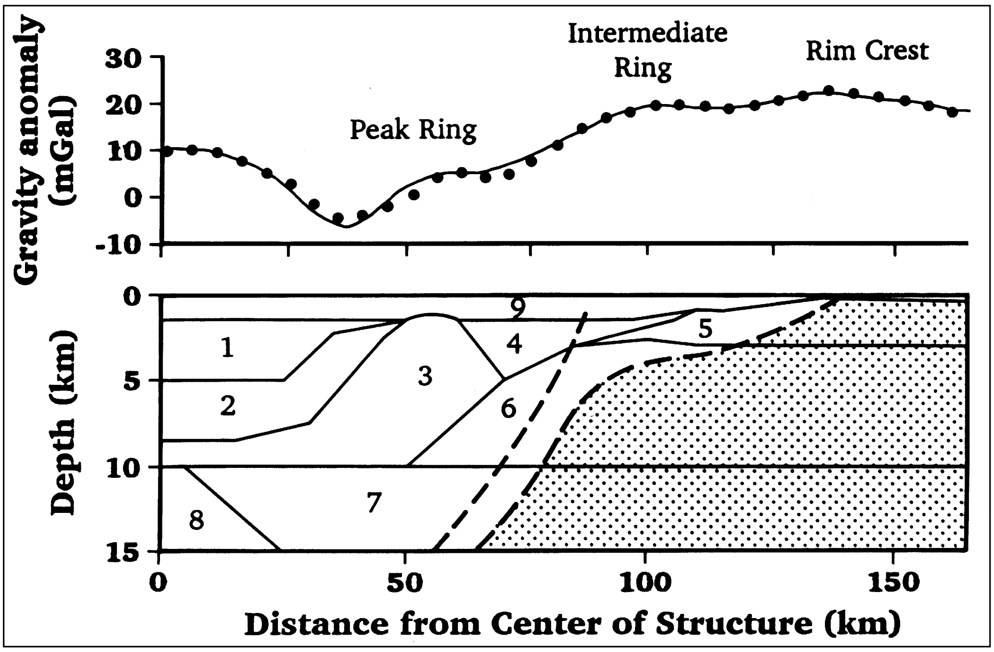
The gravity signature of the structure was recognized by PENFIELD & CAMARGO (1981) as being similar to that of known impact structures (cf. PILKINGTON & GRIEVE, 1992). The Bouguer gravity anomaly at the Chicxulub crater is predominantly negative, probably as a result of extensive brecciation. Initial data showed an extensive circular –30 mgal negative anomaly of about 180–200 km diameter, with a central 20 mgal high (HILDEBRAND et al., 1991). SHARPTON et al. (1993) compiled a new gravity map of the structure from 3134 offshore gravity measurements and 3675 land stations, which, after removing spurious points, resulted in a gravity range from –16.4 mgal to +53.6 mgal. In the Chicxulub basin, the gravity values are about 20 to

Text-Fig. 13.
Schematical diagram of the circular gravity anomalies at Chicxulub and the related surface geology. The hachured lines indicate the Ticul fault system. Carbonate rocks at the surface are: Q = Quaternary (<2 Ma), Tu = Upper Tertiary (2–35 Ma), Te = Eocene (35–55 Ma), and Tpal = Paleocene (55–65 Ma). Also indicated are several well sites at the crater area: C1 = Chicxulub 1 (near the town of Merida), S1 = Sacapuc 1, Y6 = Yucatán 6, Y1 = Yucatán 1, Y2 = Yucatán 2, Y5A = Yucatán 5A, and T1 = Ticul 1.
(After SHARPTON et al., 1993).



30 mgal lower than regional values. Besides the central 15–20 mgal gravity high, SHARPTON et al. (1993) recognized three major rings and some evidence of a fourth fractured outer ring structure (Text-Fig. 13). The central gravity high most likely is associated with the dense uplifted rocks from the basement and the impact melt rocks. The concentric gravity anomalies within the Chicxulub crater follow a $20^{0.5}$-spacing rule that has also been observed for large multiring basins on other planets and satellites (SHARPTON et al., 1993). The inner ring that is visible in Text-Fig. 13 has a diameter of 105 km and may correspond to the central peak-ring of large complex craters.

The gravity profile obtained by SHARPTON et al. (1993) and a related gravity model is shown in Text-Fig. 14. The second ring has a radius of about 77 km, while the third prominent ring extends to about 100 km radius. The broad general gravity low compared to regional values that marks the Chicxulub basin extends to about 140 km radius from the center of the structure. The gravity signature at this distance is marked by a series of subtle gravity highs that average about 2.5 mgal in amplitude. This signature, and agreement with the spacing rule mentioned above prompted SHARPTON et al. (1993) to interpret the feature as marking the outer limit of the basin rim crest, yielding a diameter of about 280 km for the Chicxulub structure.



Text-Fig. 14.
Top part of the figure shows the average radial gravity profile (solid line) obtained at the Chicxulub structure, which was constructed from traverses taken at 10° intervals through the crater, excluding the gravity-high zones between azimuths 310° and 360° (NNW sector). The bottom part shows a forward gravity model, with the following components and gravity contrasts (in g/cm³): 1 = impact melt rock and breccia (0.37), 2 = inner allogenic breccia unit (0.25), 3 = fractured uplifted crystalline basement (0.31), 4 = outer allogenic breccia (0.23), 5 = Cretaceous platform sediments (0.18), 6 = uppermost crystalline basement (0.40), 7 = intermediate basement (0.60), 8 = uplifted deep basement (0.80), and 9 = Tertiary carbonate rocks (0; all density contrasts are relative to this rock unit, which has a density of 1.8–2.0 g/cm³). The parabolic bold dashed line shows the approximate extent of the transient crater. The dotted area represents undisturbed rock units of the same nature as the labelled fields. The results of the gravity model are plotted on the top part of the figure as the heavy dotted line.
(After SHARPTON et al., 1993).

This interpretation has been challenged by PILKINGTON et al. (1994), who maintain that their modeling does not show any evidence of a fourth outer ring and, thus, that Chicxulub has a diameter of only 180 km. Some expressions on the surface, e.g., the distribution of sinkholes (PERRY et al., 1995) seem to support a diameter larger than 180 km, probably on the order of 240 km (POPE et al., 1993). A recent study of the horizontal gradient of the Bouguer anomaly over the structure by HILDEBRAND et al. (1995) was interpreted to be consistent with a 180 km crater diameter. On the other hand, another interpretation of the gravity data, in combination with drilling information, was made by SHARPTON et al. (1996). These authors conclude that the steep gravity gradients located between 75 to 105 km from the crater center are the result of the transient crater collapse rather than the crater rim, and are, thus, indicative of a total basin diameter of about 300 km. While some of the problem in estimating the diameter of Chicxulub may be of semantic nature (i.e., related to the interpretation of what constitutes the crater rim, and if Chicxulub is a multi-ring impact basin or simply a large complex crater), the determination of the dimensions of the Chicxulub structure will have to await the results of various drilling programs.

8.2. Other Geophysical Anomalies

Total magnetic field data indicate an anomaly about 180–210 km in diameter. PILKINGTON et al. (1994) show three zones, an outer zone with a radius from about 90–45 km, containing low-amplitude, short-wavelength anomalies (5–20 nT), followed by an intermediate zone with an average radius of about 45 km, which contains abundant large-amplitude (>100 nT), short-wavelength dipolar anomalies. The innermost zone has a diameter of about 40 km, is centered in the same area as the gravity high, and contains larger magnetic intensities, but a smoother structure (PILKINGTON et al., 1994). The central anomaly has been interpreted as evidence of a large impact melt body. Some on- and offshore seismic coverage of the structure is available, but only limited data have yet been released. No detailed interpretation of the seismic measurements have yet been published, with the exception of CAMARGO-ZANOGUERA & SUÁREZ-REYNOSO (1995), who conclude that their seismic data indicate a diameter of the transient crater of about 170 km.

8.3. Petrographical Evidence and Shock Metamorphism

One of the main problems of studying Chicxulub and its outcrops is that it is currently covered by several hundred meters of post-impact Tertiary sediments, requiring drilling investigations. Following the interest raised in the Chicxulub structure due to the geophysical studies, studies were made on a limited amount of drill core samples obtained from Petróleos Mexicanos (PEMEX; the Mexican state oil company) exploration drilling in the 1950s and 1960s. The location of the boreholes is indicated in Text-Fig. 13. Two of the boreholes at the center of the structure penetrated in a dense crystalline rock, which was initially thought to represent andesitic volcanic material, but was more recently recognized to be related to the impact structure (HILDEBRAND et al., 1991; see also KRING, 1993, and KOEBERL, 1993).

So far, samples from several intervals in the Y6 borehole, located about 60 km from the center of the structure at the flank of the 105 km diameter inner ring, and from the centrally located C1 borehole were studied in most detail.

Silicate rocks recovered from the Y6 core consist of a well-sorted, graded polymict breccia sequence from about 1100 m below sea level to >1400 m, with clasts in the upper parts being heavily altered, while well-crystallized coherent melt rocks follow closer to the bottom of the sequence. In the interval Y6–N17 (about 1295 to 1299 m) and Y6–N19 (1377–1380 m), abundant melt rock fragments and glass remnants were found (SHARPTON et al., 1992). These polymict breccias contain about 40–60 vol.-% of angular to rounded clasts, several cm in size, of fine-grained to glassy, often altered, melt rock in a medium- to coarse grained melt matrix composed mainly of feldspar and pyroxene with little evidence of alteration (SHARPTON et al., 1992). The C1–N10 interval (1393–1394 m) is different from the other samples, as it has a coarse-grained matrix and does not contain any unmelted clasts (SCHURAYTZ et al., 1994). The matrix contains predominantly subhedral to euhedral pyroxene (up to 0.7 mm in size) and plagioclase with a variety of morphologies (Text-Fig. 15a–d). The melt rock contains typical fine-grained feathery spherulitic devitrification textures (Text-Fig. 15c,d). The overall appearance and texture of the breccia and melt rock samples are very similar to those observed in known impact craters.

The most important findings in the breccia and melt rock samples are clear, abundant evidence for the presence of shock metamorphism, including the following:

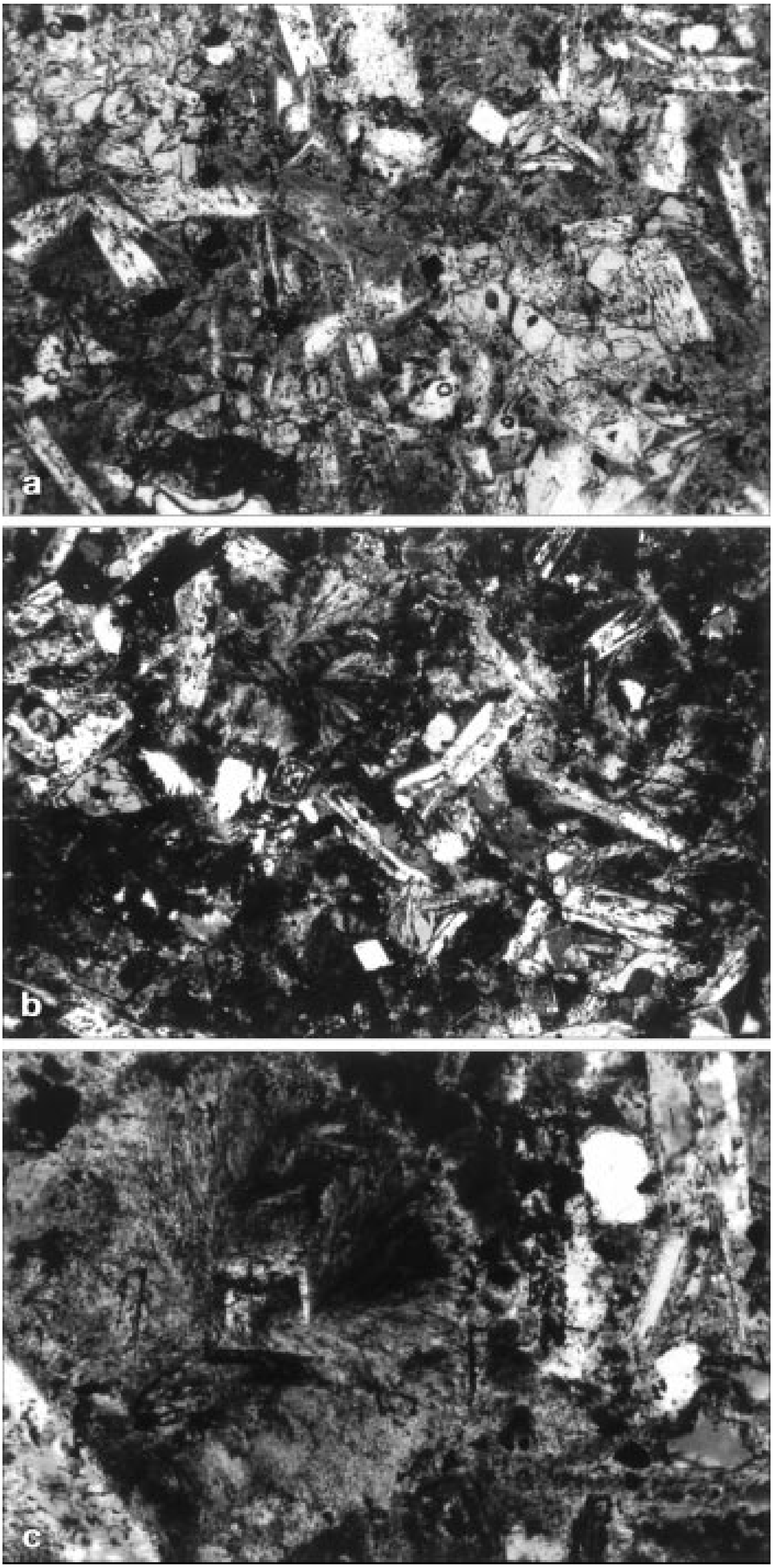
a) abundant planar deformation features (PDFs) in quartz and feldspar crystals from crystalline basement clasts within the breccias and melt rocks (SHARPTON et al., 1992); up to five sets of PDFs were found and measured by universal stage methods, showing the presence of the impact-characteristic orientations, with $\{10\bar{1}3\}$ or ω, and $\{10\bar{1}2\}$ or π orientations, being the most common (Text-Fig. 16);
b) shock mosaicism in quartz and feldspar;
c) diaplectic glass occurs as partly digested inclusions within glasses and melts; and
d) impact melts, glassy and fine-grained recrystallization products (e.g., Text-Fig. 15c). In many breccias from the intervals mentioned above about one third of the quartz grains and most of the feldspar grains are shocked (SHARPTON et al., 1992).

These findings provide unequivocal evidence for an impact origin of the Chicxulub structure.

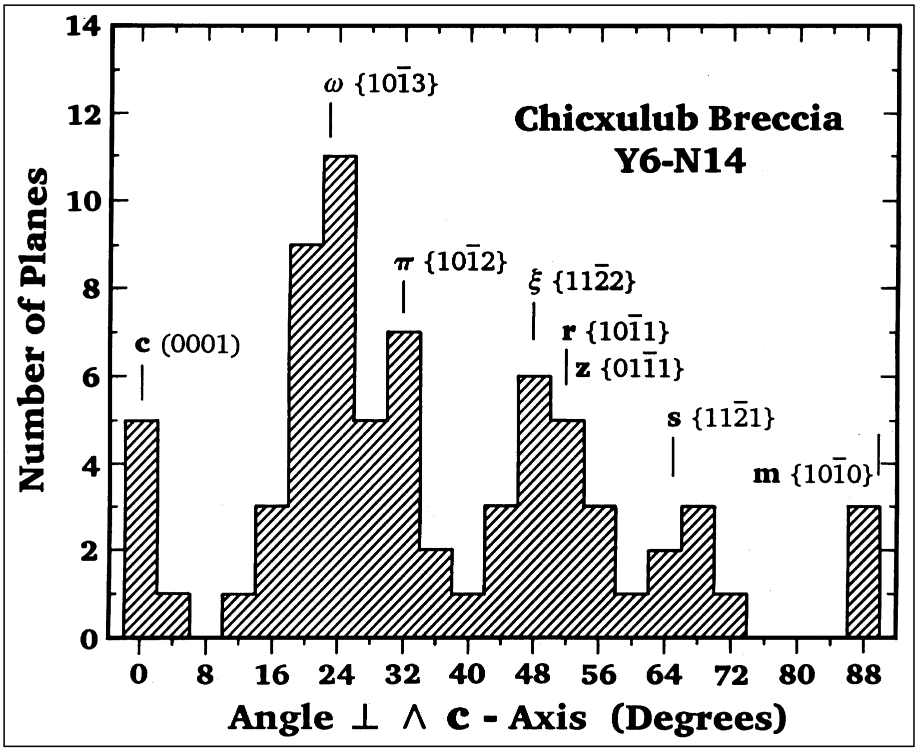
8.4. Meteoritical Signature in Impact Melt Rocks

In some of the melt rocks from the C1 and Y6 drill cores, Ir contents of up to 13.5 ppb were found, indicating the presence of an extraterrestrial component (SHARPTON et al., 1992; KOEBERL et al., 1994c; SCHURAYTZ et al., 1994). However, the Ir distribution (and that of other siderophile elements) was found to be very inhomogeneous. SCHURAYTZ et al. (1994) found pyrite crystals with high and variable contents of Ni and Co, and some opaque mineral grains also contained high Ir, which they interpreted as evidence for extensive post-impact hydrothermal activity that may have led to a redistribution of the siderophile elements. Some of the inhomogeneity may also be due to inhomogeneous distribution within the impact melt body.

Text-Fig. 15.
Microphotographs of the impact melt rock from the Chicxulub C1 drill core, N10 interval.
a) Fine intergrowth of plagioclase and pyroxene crystals in fine-grained unaltered matrix.
2.4 mm wide, parallel polars.
b) Zone with feathery spherulitic devitrification texture in center, surrounded by larger plagioclase and pyroxene crystals.
2.4 mm wide, crossed polars.
c) Close-up of fine-grained spherulitic (swallow-tail) devitrification texture.
1.2 mm wide, crossed polars.

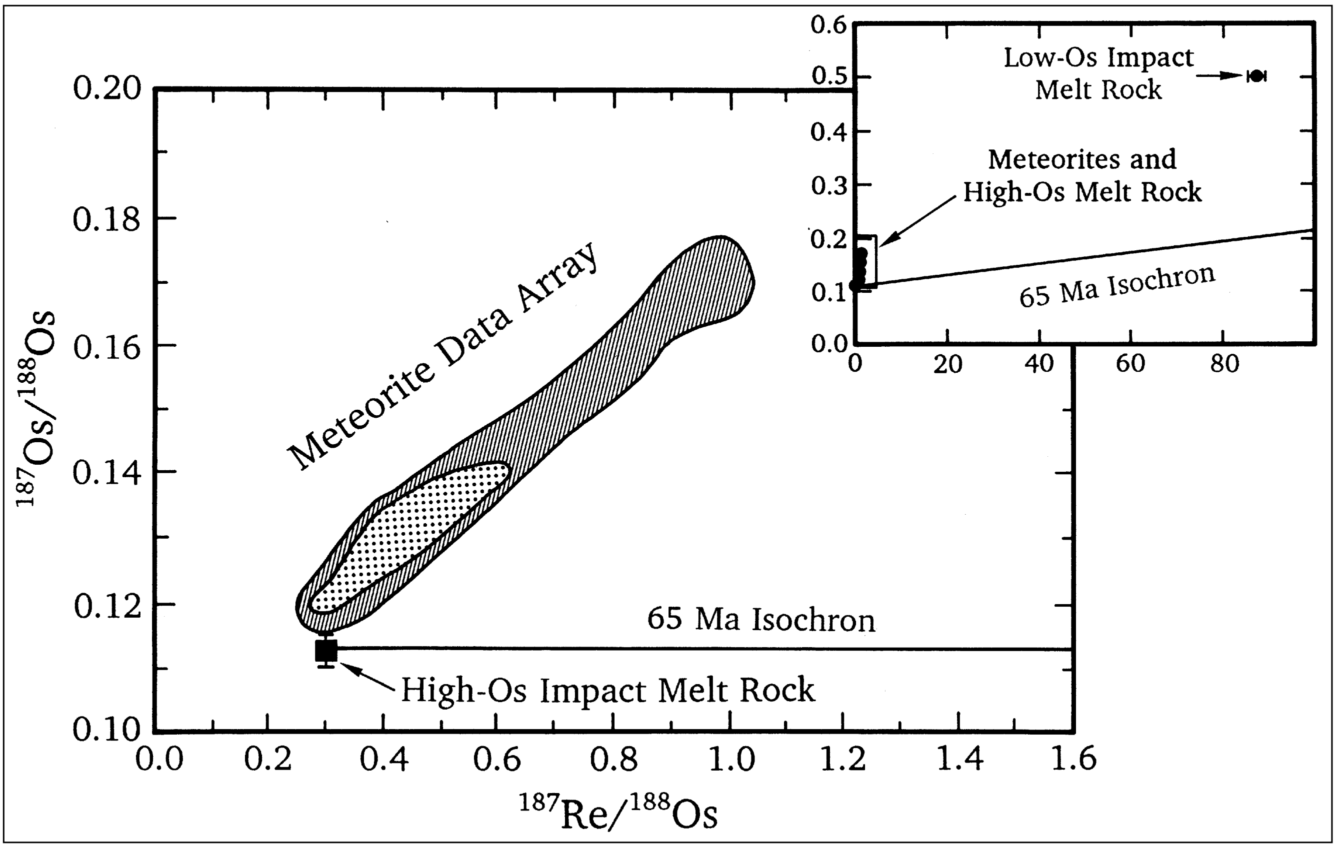


As the distribution of the PGEs may be ambiguous, KOEBERL et al. (1994c) studied the Re and Os abundances and Os isotopic compositions of melt rocks from the C1 core. One of the melt rock samples contained 25.2 ppb Os, and very low $^{187}Os/^{188}Os$ and $^{187}Re/^{188}Os$ ratios of 0.113 and 0.305, respectively. These values are inconsistent with derivation from old continental crust, but very close to the meteoritic data array (Text-Fig. 17). Another melt rock has only 0.056 ppb Os and a high $^{187}Os/^{188}Os$ ratio of 0.51 (Text-Fig. 17), similar to continental crustal values. This result supports the conclusions of the elemental studies, that the meteoritic component in the melt rock is inhomogeneously distributed. As mentioned before, the similarity between mantle and meteorite Os isotopic compositions makes the interpretation of Os isotope data difficult, and requires supporting studies to exclude the presence of any mantle components. In the case of Chicxulub, the large size of the structure may suggest that mantle material could have been excavated. However, impact models indicate that the Chicxulub-forming impact event excavated to a depth of 17–20 km, which is within the upper part of the crust. The depth of the transient crater, about 45–60 km, includes excavation plus downward displacement of the target beneath the impact locality. Hence, it is unlikely that the crater-forming event could have mixed mantle material into the Chicxulub melt rocks. In addition, trace element, Rb-Sr, and Sm-Nd isotopic characteristics of the samples are typical of



Text-Fig. 16.
Histogram showing the frequency (number of PDF sets) versus crystallographic orientation of the poles of the PDFs relative to the c-axes of the respective crystals in quartz grains from the Chicxulub impact breccia (interval Y6–N14). 71 sets in 30 grains were measured.
After SHARPTON et al. (1992).

rocks from the continental crust (see below). Thus, major contributions from basaltic, ultramafic, or other mantle-derived material are excluded, which is confirmed by the petrographical observations. MORB and related basalts contain only sub-ppb amounts of Os and $^{187}Os/^{188}Os$ ratios that are slightly higher than that observed in the Chicxulub melt rock (KOEBERL et al., 1994c). Depleted lithospheric mantle xenoliths are the only terrestrial rocks known with subchondritic $^{187}Os/^{188}Os$ ratios, but Os abundances in xenoliths are too low (2–3 ppb) to explain the high abundances observed in the Chicxulub melt rock. In addition, no basalts or ultramafic bodies have been observed in the Chicxulub target area or in impact breccias (SHARPTON et al., 1992, 1993; SCHURAYTZ et al., 1994). As a result, it has to be concluded that no mantle component in the Chicxulub samples is present. Consequently, the Os abundance and isotopic data suggest that the Chicxulub melt rocks contain up to 3 wt.-% of a chondritic component, which is within



Text-Fig. 17.
Ratios of $^{187}Os/^{188}Os$ versus $^{187}Re/^{188}Os$ for impact melt rock from the Chicxulub impact structure, Mexico.
(After KOEBERL et al., 1994c).
The data array for meteorites (hatched area = iron meteorites, dotted area = carbonaceous chondrites) is also indicated. This impact melt sample also has a very high Os content (25 ppb), clearly indicating a meteoritic contamination. The inset shows – at an expanded scale – the data for a low-Os melt rock sample, which has crustal $^{187}Os/^{188}Os$ and $^{187}Re/^{188}Os$ values, and seems to lack an extraterrestrial component.

the range of meteoritic components reported for large craters (see, e.g., PALME, 1982).

In a recent development, SCHURAYTZ et al. (1996) report on the discovery of two minute particles (dimensions on the order of a few µm) in Chicxulub impact melt (from the C1–N10 and Y6–N19 core sections) that consist of almost pure Ir. One particle seems to be pure Ir (99 wt.-%), while the other one contains a few wt.-% of other PGEs (e.g., Os; B. SCHURAYTZ, personal communication [1996]). In addition, KYTE (1996) described an unusual inclusion embedded in K-T boundary sediments in DSDP drill core 576 (western North Pacific). This altered fragment has almost chondritic (within a factor of <2) Ir, Fe, and Cr abundances and was interpreted by KYTE (1996) as a possible fragment of the K-T boundary impactor.

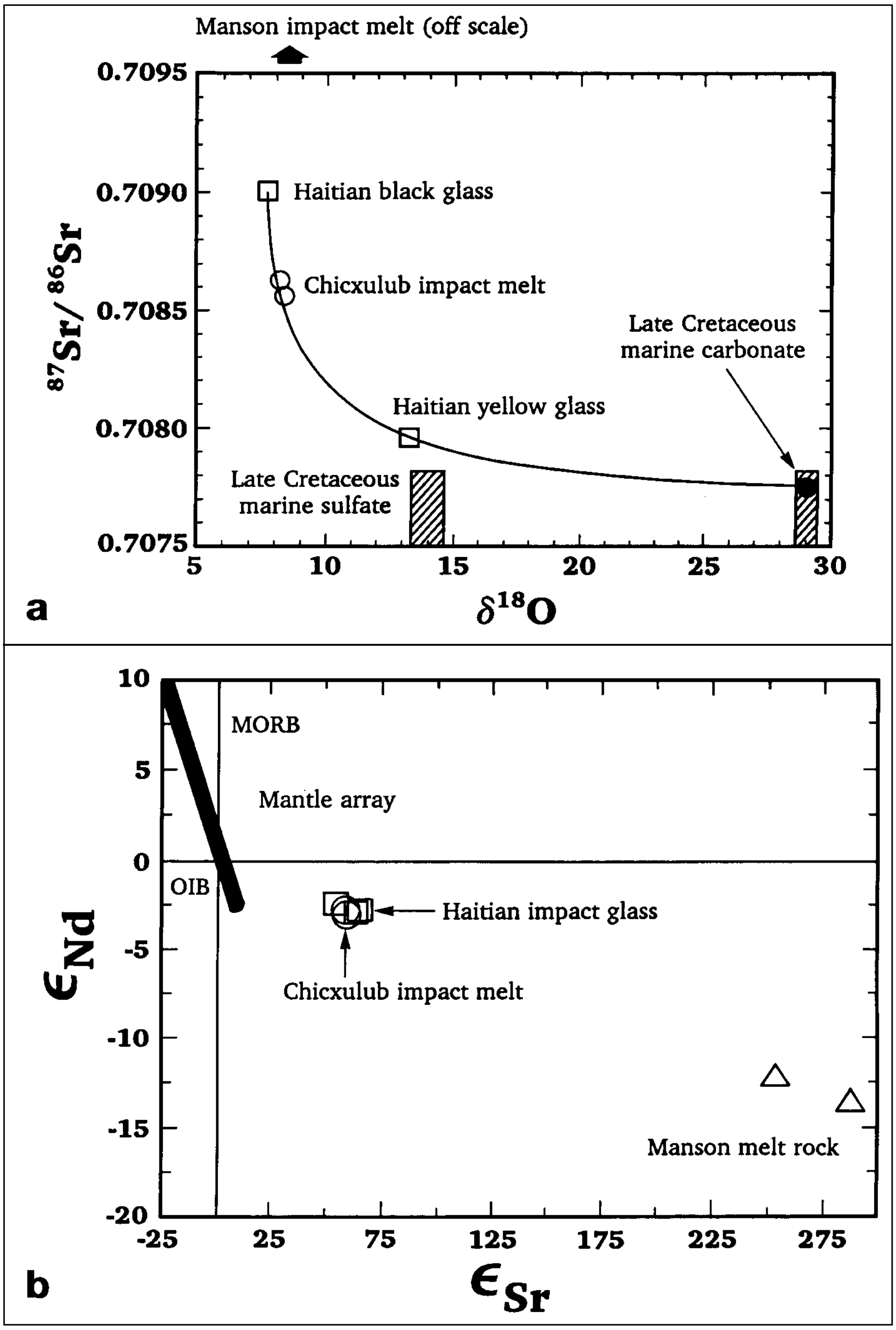
However, at this time this interpretation should be taken with caution, as other important elemental abundances have either not yet been determined, or are non-chondritic (e.g., Au has an abundance of 1000 times the chondritic concentration, which is on the order of >100 ppm (!); KYTE, 1996).

8.5. Geochemical Signature of Impact Melt Rocks and Connection to K-T Boundary Deposits

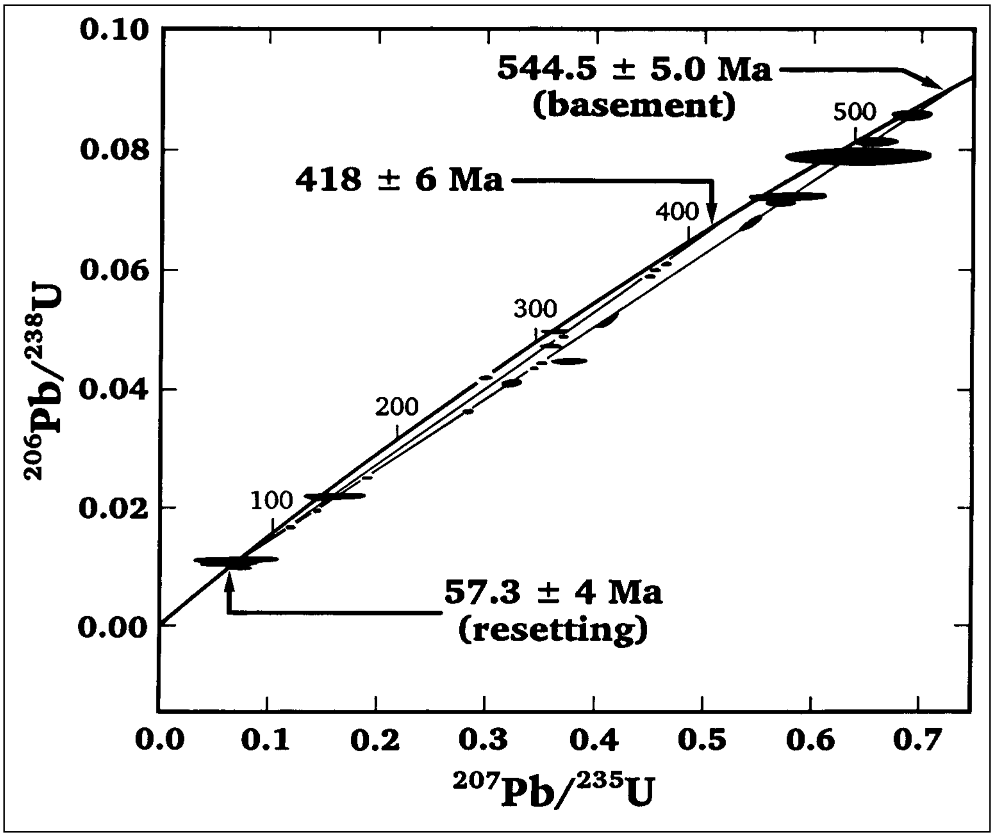
Major and trace element compositions of breccias and impact melt rocks from the Chicxulub structure are very similar to values obtained for average sediments and average crustal rocks (e.g., KOEBERL, 1993; SCHURAYTZ et al., 1994). Isotopic measurements are of great importance to establish a link between the Chicxulub crater and the impact deposits at the K-T boundary. BLUM et al. (1993) report on the measurement of the Rb-Sr, Sm-Nd, and oxygen isotopic composition of melt rocks from Chicxulub. They found that in a diagram of the oxygen versus the strontium isotopic composition (Text-Fig. 18a), the data for the Chicxulub melt rocks fall on a mixing hyperbola between the various types of impact glasses from the Haitian K-T boundary and a carbonate endmember (representing the carbonate platform rocks at Yucatán).

Data for impact melt rocks from the Manson crater are inconsistent with the Haitian impact glasses or Chicxulub melt rocks. This result indicates a common source for the Haitian impact glasses and the Chicxulub crater rocks.

A similar result is obtained if the Sm-Nd isotopic composition is considered as well. A plot of $\epsilon_{Sr}{}^{65Ma}$ versus $\epsilon_{Nd}{}^{65Ma}$ for Chicxulub melt rock samples and Haitian impact glasses (Text-Fig. 18b) shows that these two materials have similar values at about +58 and –3 for $\epsilon_{Sr}{}^{65Ma}$ and $\epsilon_{Nd}{}^{65Ma}$, respectively.



Text-Fig. 18.
Isotope plots for Chicxulub melt rocks (after BLUM et al., 1993).
a) Plot of $^{87}Sr/^{86}Sr$ (recalculated to 65 Ma) versus $\delta^{18}O$ for Chicxulub impact melt rock (open circles) and impact glass from Haiti (open squares). Also plotted are the fields for average carbonate (hatched area in lower right; average is shown by solid circle) and Late Cretaceous marine sulfate. The Chicxulub melt rock plots exactly on a mixing hyperbola defined by the two types of Haitian impact glass and the carbonate endmember, indicating a common source. On the other hand, the Manson crater rocks plot off the scale and are unrelated.
b) Plot of $\epsilon_{Sr}{}^{65Ma}$ versus $\epsilon_{Nd}{}^{65Ma}$ for Chicxulub melt rock samples and Haitian impact glasses, as well as some Manson crater rocks. The values for Chicxulub melt rock and impact glasses are similar at about +58 and –3 for $\epsilon_{Sr}{}^{65Ma}$ and $\epsilon_{Nd}{}^{65Ma}$, respectively, and significantly different from mantle compositions and Manson crater data.



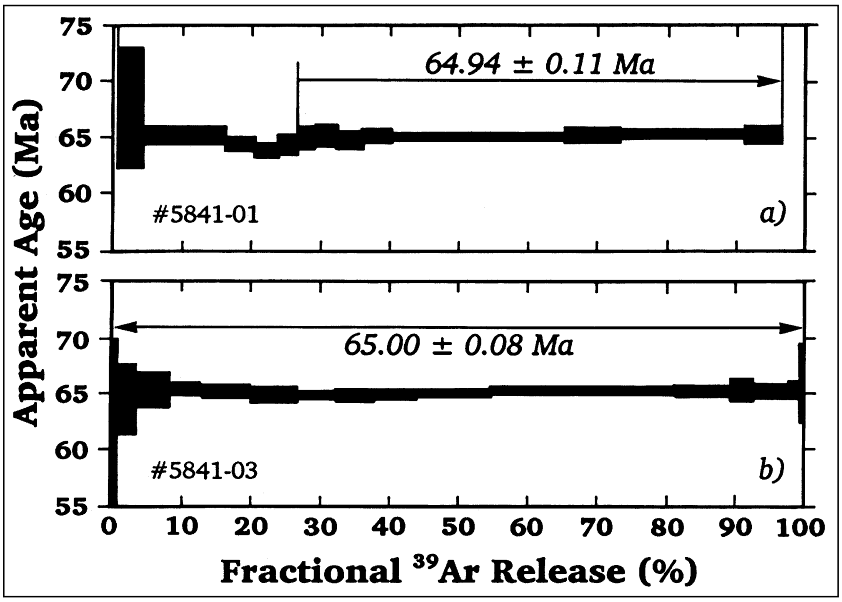
Text-Fig. 19.
Concordia diagram for U-Pb data of shocked single zircons from the K-T boundary sediments in Colorado and Haiti, and from Chicxulub impact breccia (after KROGH et al., 1993).
The results for the three sites are indistinguishable. Most zircons define an upper intercept of about 545 Ma, with a lower intercept (defined by the most severely shocked zircons) of about 57 Ma, which is very close to the impact age of 65 Ma. A minor component, defined by Chicxulub and Haiti zircons, has an intercept at about 418 Ma. The results demonstrate a common source for the K-T boundary ejecta and the Chicxulub impact breccias.

Manson crater rocks have significantly different values. The depleted mantle Nd model ages of the Chicxulub rocks fall in a tight range of about 1040–1080 Ma, suggesting that the silicate endmembers of these breccias and melt rocks had a source with a middle Proterozoic average crustal residence age, but younger sedimentation and crystallization ages (BLUM et al., 1993). The data are also inconsistent with derivation from the mantle, because values for rocks derived from the upper mantle generally fall in a narrow range of ϵ_{Nd} of +4 to +10 and ϵ_{Sr} of –10 to –30.

Another link between the K-T boundary impact ejecta and the Chicxulub rocks is obtained from single zircon U-Pb ages. KROGH et al. (1993) and KAMO & KROGH (1995) have been able to analyze the U-Pb isotopic composition of single zircons (some weighing as little as 1 µg) from various K-T boundary sites and from the Chicxulub crater. Text-Fig. 19 shows a concordia diagram for shocked zircons extracted from an impact melt breccia from Chicxulub and from K-T boundary deposits in Colorado and Haiti. All data show a similar result, with a major intercept at about 545 Ma, indicating a Pan-African basement age. Data from a Canadian K-T boundary site yield the same results (KAMO & KROGH, 1995). The more the zircons are shocked, the lower (more reset) their age is, leading to a lower intercept of 57 Ma, which is remarkably close to the impact age at 65 Ma. The U-Pb data on single zircons agree with the Rb-Sr and Sm-Nd data and demonstrate a geochemical link between the Chicxulub structure and impact ejecta at the K-T boundary.

8.6. Age of the Chicxulub Structure

As the studies mentioned in the previous sections have shown, without reasonable doubt, that Chicxulub is not only one of the largest (if not the largest) impact structure presently known on earth, and that the materials found at the K-T boundary sites around the world are very likely derived from the Chicxulub crater, the only question that remains is, does the age fit. Two detailed studies of the radiogenic age of the impact melt rock and impact glasses from the Chicxulub structure have been published.



Text-Fig. 20.
^{40}Ar-^{39}Ar age spectra, obtained by stepwise heating, of two impact glass samples that were extracted from the Chicxulub C1–N9 impact melt breccia. The patterns show good plateaus and the results are indistinguishable from those for the impact glasses from Haiti, and demonstrate a K-T boundary age for the Chicxulub impact crater. After SWISHER et al. (1992).

SHARPTON et al. (1992) used hand-picked fragments of fine-grained melt rock from the Y6–N19 and C1–N10 intervals, weighing 0.46 to 1.83 mg, for ^{40}Ar-^{39}Ar stepwise heating age determinations. They found that most spectra show evidence of low-temperature alteration, but that selected higher-temperature increments in the stepwise heating spectra of a sample from C1–N10, which result in an age of 65.2±0.1 Ma, are likely to represent the crystallization age. SWISHER et al. (1992) were able to separate individual small glass fragments (0.4–0.5 mm in size) from the C1–N9 interval. These samples weighed about 0.2 to 0.3 mg and were also studied using the ^{40}Ar-^{39}Ar stepwise heating technique. As SWISHER et al. (1992) seem to have succeeded in obtaining fresher samples, they obtained much better plateau ages than SHARPTON et al. (1992). Two typical examples of the Chicxulub glass age spectra are shown in Text-Fig. 20, with plateau ages of 64.94±0.11 and 65.00±0.08 Ma. SWISHER et al. (1992) found an average age of 64.98±0.05 Ma for their samples, which is indistinguishable from the age of 65.07±0.10 Ma that they obtained for impact glasses from the Beloc (Haiti) and Arroyo el Mimbral (Mexico) K-T boundary layers. These results lead to the rather obvious conclusion that Chicxulub is indeed of K-T boundary age.

9. Summary and Outlook

Interest in impact cratering studies was recently stimulated by research related to the events marking the K-T boundary. However, impact cratering still remains to be one of the least studied and least appreciated geological processes, even though over the past three decades, researchers have studied impact cratering in nature, in the laboratory, and by computer models. So far about 150 impact structures have been identified on the earth, and several are added to our lists every year, but many more must exist. The detailed research leading first to the conclusion that an impact event must have taken place at the end of the Cretaceous, and later to the identification of the Chicxulub crater in Yucatan, Mexico, as “the” K-T boundary source crater, serves to illustrate the necessity – and success – of interdisciplinary studies.

As mentioned above, several lines of evidence demonstrate that Chicxulub is the K-T boundary crater:

a) geophysical evidence (gravity, magnetic, and seismic anomalies) shows the presence of a large (200–300 km diameter) structure with geophysical characteristics that are identical to those of other known large impact structures;
b) drill core sample studies led to the discovery of shock metamorphic effects in the Chicxulub breccias and melt rocks, providing unambiguous evidence for an impact origin of the structure;
c) the presence of a meteoritic component in some impact melt rocks, as shown by PGE and Re-Os isotope studies, provided further evidence for the impact origin;
d) isotope geochemistry studies, using Rb-Sr, Sm-Nd, O, and U-Pb isotopic data, have shown that the impact debris at the K-T boundary sites around the world and the impact breccias and melt rocks have a common source, and
e) radiometric age determinations on melt rocks and impact glasses from Chicxulub and impact glasses from the Haitian K-T boundary gave identical results, at 65.0 Ma, which is indistinguishable from the age of the K-T boundary.

The specific geographic location of the Chicxulub crater, with abundant carbonate and anhydrite rocks of the Yucatan platform, must have had important consequences for the biosphere. During the impact event, large amounts of CO_2 and SO_2 must have been ejected into the atmosphere. Preliminary estimates (assuming about 300–2000 km^3 of vaporized sediments) yield sulfur masses of $3.5 \cdot 10^{16}$ to $7.0 \cdot 10^{17}$ g, and about 10^{19} g of CO_2 that were released almost instantaneously into the atmosphere (CHEN et al., 1994; POPE et al., 1994). While any detailed discussion or speculation on the long-term effects of such enormous amounts of gas are beyond the scope of this paper (but see, e.g., CHEN et al., 1994; POPE et al., 1994; PIERAZZO et al., 1996; LYONS & AHRENS, 1996), it is obvious that there must have been short-time, as well as long-time climatic changes that most likely led to the mass extinctions that mark the K-T boundary.

Future work on the Chicxulub structure will involve detailed investigations of several shallow drill cores that have been obtained in 1994 and 1995 by Mexican researchers. Currently (1996) these cores are being documented and will be made available shortly. These cores are of significant importance for understanding the crater diameter. Preliminary studies seem to be in agreement with the 300-km-diameter estimate (V.L. SHARPTON, personal communication [1996]). In addition, studies of proximal ejecta within a few crater radii are currently being undertaken. For example, an unusual deposit of breccia-like material with a thickness of several tens of meters was found at a quarry at Albion Island, in northern Belize, which was interpreted as the most proximal ejecta deposit of the Chicxulub crater found so far (POPE et al., 1996; OCAMPO et al., 1996). In addition, several cores were drilled in early 1996 during Leg 165 of the Ocean Drilling Program to study the proximal Chicxulub ejecta in the Caribbean Sea (SIGURDSSON et al., 1995). The results from the studies, which are vigorously pursued, will allow detailed understanding not only of the properties and size of the Chicxulub crater, but also of the important interactions of this major impact event with the geo- and biosphere.

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