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Terrestrial impact: The record in the rocks*

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Abstract—Approximately 130 terrestrial craters are currently known. They range up to 140 km, and perhaps as much as 200 km, in diameter and from Recent to ~2 billion years in age. The known sample, however, is highly biased to geologically young craters on the better known cratonic areas. The sample is also deficient in small (D < 20 km) craters compared to other planetary bodies. These biases are largely the result of active terrestrial geologic processes and their effects have to be considered when interpreting the record. The strength of the terrestrial cratering record lies in the availability of ground truth data, particularly on the structural and lithological nature of craters, which can be interpreted to understand and constrain large-scale impact processes. Some contributions include the definition of the concept of transient cavity formation and structural uplift during cratering events. Depths of excavation are poorly constrained, as very few terrestrial craters have preserved ejecta.

Unlike their planetary counterparts, terrestrial impact craters are mostly recognized not by morphology but by the occurrence of characteristic shock metamorphic effects. Their study has led to models of shock wave attenuation and an understanding of the character and formation of various impact-lithologies, including impact melt rocks. They, in turn, aid in interpreting the nature of extraterrestrial samples, particularly samples from the lunar highlands. The recognition of diagnostic shock metamorphic effects and the signature of projectile contamination through geochemical anomalies in impact lithologies provide the basis for recognizing the impact signature in K/T boundary samples. The record also provides a basis for testing hypotheses of periodic cometary showers. Although inherently not suitable to define short wavelength periods in time due to relatively large uncertainties associated with crater ages, the current record shows no evidence of periodicity. Future directions in terrestrial impact studies will likely continue to focus on the K/T and related problems, including the recognition of other impact signatures in the stratigraphic record. Some emphasis will likely be given to the economic potential of craters and individual large structures, such as Sudbury, will provide an increasingly better understood context for interpreting planetary impact craters. To live up to the full potential of the record to constrain impact processes, however, more basic characterization studies are required, in addition to emphasis on topical areas of study.

INTRODUCTION

PHYSICAL MODELS, THE NATURE OF PLANETARY SURFACES and the results of the study of lunar and meteorite samples all attest to the importance of hypervelocity impact as a geologic process in the context of solar system and planetary evolution. An integral component in understanding the geological and geophysical consequences of large-scale hypervelocity impact is the data supplied by terrestrial impact craters. As the prominence of impact as a process has grown, so has the study of terrestrial impacts. Today, the discovery and description of terrestrial impact structures is no longer limited to a small group of experts but is now within the domain of the larger geologic community.

Barringer or Meteor Crater, Arizona, is regarded as the classic example of a well-preserved, terrestrial simple crater in the kilometer size-range. This was not always the case. When the impact origin of Meteor Crater was proposed by D. M. Barringer (1906), it touched off a debate with the established geologic community that ended only in the late 1950s, with acknowledgement of an impact origin by the United States Geological Survey (Shoemaker, 1960). Similar debates occurred over the origin of other structures (*e.g.,* Boon and Albritton, 1936, 1937, 1938; Bucher, 1936) and the hypothesis of a cryptoexplosion or explosive volcanic origin for some impact structures continued to have adherents through to the early 1970s (*e.g.,* Bucher, 1963, 1965; Currie, 1970, 1971). While generally laid to rest through the cumulative evidence from lunar and terrestrial studies, this specter of cryptoexplosion structures has resurfaced recently with regard to the evidence for impact at the Cretaceous-Tertiary boundary (Officer and Drake, 1985, 1989; Loper and McCartney, 1988). Clear and cogent arguments against the revival of such suggestions have been made most recently by French (1990) and Sharpton and Grieve (1990).

This contribution is a modified and expanded version of the Barringer Address presented at the 53rd Meteoritical Society Meeting. As a graduate student, I argued vigorously with my contemporaries against the impact origin of the Sudbury and Brent structures. My conversion to a pro-impact stance was essentially instantaneous and occurred upon comparing the microscopic textures and structure of Apollo 14 breccias (Grieve *et al.,* 1975) with breccia lithologies from Sudbury. It was an interest in lunar petrology that developed initially my interest in the results of terrestrial impact and I came to the study of terrestrial impact craters at a time when there was a growing acceptance of their impact origin because of the results of the Apollo missions to the moon. I can recall only a few cases when I had to devote intellectual effort to defend the fact that the specific structure or phenomenon I was studying was produced by impact. For the freedom to devote my energies to the study of nature's most catastrophic geologic process, I owe a debt to those earlier workers who had argued previously the case of terrestrial impact craters.

This contribution attempts to summarize some of the basic characteristics of the impact history of the Earth, as recorded in the surface rocks, and is undoubtedly influenced by my prejudices from over fifteen years of study of terrestrial impact structures. An effort has been made to indicate the strengths and weaknesses of both the terrestrial record of impact and the studies that have been made of this record. The main strengths of the study of terrestrial impact structures to understanding cratering processes are the observation of ground truth data and their interpretation. Particular contributions include: the determination of subsurface structure, the spatial distribution and

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nature of allochthonous and autochthonous shocked lithologies, including melt rocks and various types of breccias, and the geophysical character of impact craters. More general contributions include estimates of the cratering rate and its variation with time and constraints on such hypotheses as impact-related extinction events, such as hypothesized for the Cretaceous-Tertiary (K/T) boundary, and speculations on the occurrence of periodic cometary showers, which may affect the evolution of the biosphere (Alvarez and Muller, 1984) and possibly the geosphere (Rampino and Stothers, 1984). In evaluating and making interpretations, based on the overall cratering record, it is important to realize that one of the basic characteristics of the terrestrial impact record is its bias.

BASIC CHARACTERISTICS OF THE RECORD

Following the terminology of Dence (1965) and Krinov (1966), small impact pits (< 10 m) formed by meteorites that remain intact are not considered here as impact craters. If impact sites with multiple craters produced by a fragmented and dispersed impacting body, *e.g.,* the Henbury craters in Australia, are counted as a single impact, then the current number of known terrestrial impact craters is ~130. Small impact craters often retain physical evidence of the impacting body, thus, readily establishing their impact origin. The largest known crater with associated meteoritical material is the 1.2 km diameter (D) Barringer Crater. At larger craters, the impacting body is not appreciably slowed by atmospheric retardation. As a result, the peak shock pressures generated on impact are sufficient to melt and vaporize the impacting body and destroy it as a physical entity. In these cases, the identification of a particular terrestrial structure as an impact crater rests on the principal criterion of the occurrence of diagnostic shock metamorphic effects.

Table 1 summarizes some of the holdings of a computerized data base on terrestrial impact craters maintained by the Geological Survey of Canada. This listing supercedes previous compilations (*e.g.,* Grieve, 1982; Grieve and Robertson, 1987) and reflects recent additions and corrections to the data base of previously known craters. Table 1 represents structures for which, in our judgement, there is sufficient evidence to ascribe an impact origin. All the structures listed in Table 1 have meteoritic material and/or shock metamorphic effects. In a few cases, the evidence for shock is the subject of discussion, *e.g.,* Vredefort—Reimold (1987), Grieve *et al.* (1990), Uvalde—Robertson (1980), Sharpton and Neilsen (1988). Conversely, in other cases, the evidence for shock is generally accepted but its interpretation as due to impact is questioned by some individuals, *e.g.,* Sudbury—Stevenson (1990), Slate islands—Sage (1978). In most cases, however, the evidence for shock metamorphism and the impact interpretation is straightforward and generally accepted. The data base from which Table 1 is derived is an evolving entity. For example, most recently Nazarov *et al.* (1991) have suggested that the twin structures Kara, 65 km, and Ust-Kara, 25 km, (Table 1) may, in fact, be a single structure ~120 km in diameter. The Ust-Kara structure is almost entirely under the waters of the Kara Sea and its identification is based on outcrops of impact lithologies along the shore (Masaitis *et al.,* 1980). Nazarov *et al.* (1991), however, now consider that these lithologies are related to the adjacent Kara structure. As this interpretation is contained only in abstract form, however, we have retained Ust-Kara, for the moment, as an impact crater (Table 1). If readers detect any errors in Table 1 or know of additional craters, we would appreciate receiving the information for the data base.

The spatial distribution of known terrestrial impact craters is biased towards the cratonic areas of Australia, N. America and western USSR (Fig. 1). This bias is a result of two factors. These areas have been relatively stable for considerable geologic periods and, thus, are the best available surfaces for the long term acquisition and preservation of impact craters. They are also areas where there have been active search programs to identify and study impact craters. There are a number of cratonic areas, *e.g.,* in Africa and S. America, where there are still relatively few recognized craters (Fig. 1a). This reflects the generally lower level of geologic knowledge of these areas combine, in part, with difficulties in terrain and access.

The effect of programs, either institutional or individual, to study impact craters is illustrated by the distribution of known impact craters as of almost twenty years ago (Fig. 1b) compared with the present day (Fig. 1a). Comparing Table 1 with the list of terrestrial craters compiled by Dence (1972) indicates that the known record of impact in N. America has increased (~50%), from 30 to 46 structures at present. Although the discovery rate has remained relatively low, at less than one structure per year, the additions include some scientifically important structures, *e.g.,* Beaverhead (Hargraves *et al.,* 1990), Haughton (Robertson and Mason, 1975) and Montagnais, the first totally submerged impact structure to be identified (Jansa and Pe-Piper, 1987). The largest additions to the known record have been made in Australia and the U.S.S.R. (Figs. 1a, b). The record in Australia has increased (~300%), from 5 (Dence, 1972) to 16 structures (Table 1) and the record in the USSR has increased (~2000%), from 2 (Dence, 1972) to 36 (Table 1). This large increase in the number of known structures in the USSR is to some extent influenced by the length of time it takes Soviet scientific literature to be generally known outside the USSR. For example, Dence (1972) noted the 100 km Popigai structure but did not list it as a confirmed impact structure; although, Masaitis (1971) had already established its impact origin. Nevertheless, the trend is clear. While the N. American record had increased only slightly, considerable numbers of craters have been identified and studied in geographical areas where there has been institutional will or support for individual efforts. The lead in the study of terrestrial impact craters, which was once the role of Canada and the United States, is now squarely with the programs within the USSR.

Active terrestrial processes such as erosion and sedimentation result in characteristics of the impact record that are not present on other planetary bodies. For example, although impact is a surface phenomenon, buried impact craters account for ~20% of the known terrestrial record. In general, these craters are buried by post-impact sediments deposited in epicontinental seas. Fifty percent of the known buried structures occur in the USSR. This is due to local geology with large areas of platform sediments and the systematic coverage of large areas by relatively closely spaced geophysical, particularly gravity, data. Most buried structures were initially detected as geophysical anomalies and later drilled to confirm their impact origin. Individual craters can have complex post-impact histories involving erosion, burial, exhumation and erosion. If unprotected from erosion, the geological signature of a 20 km diameter crater may

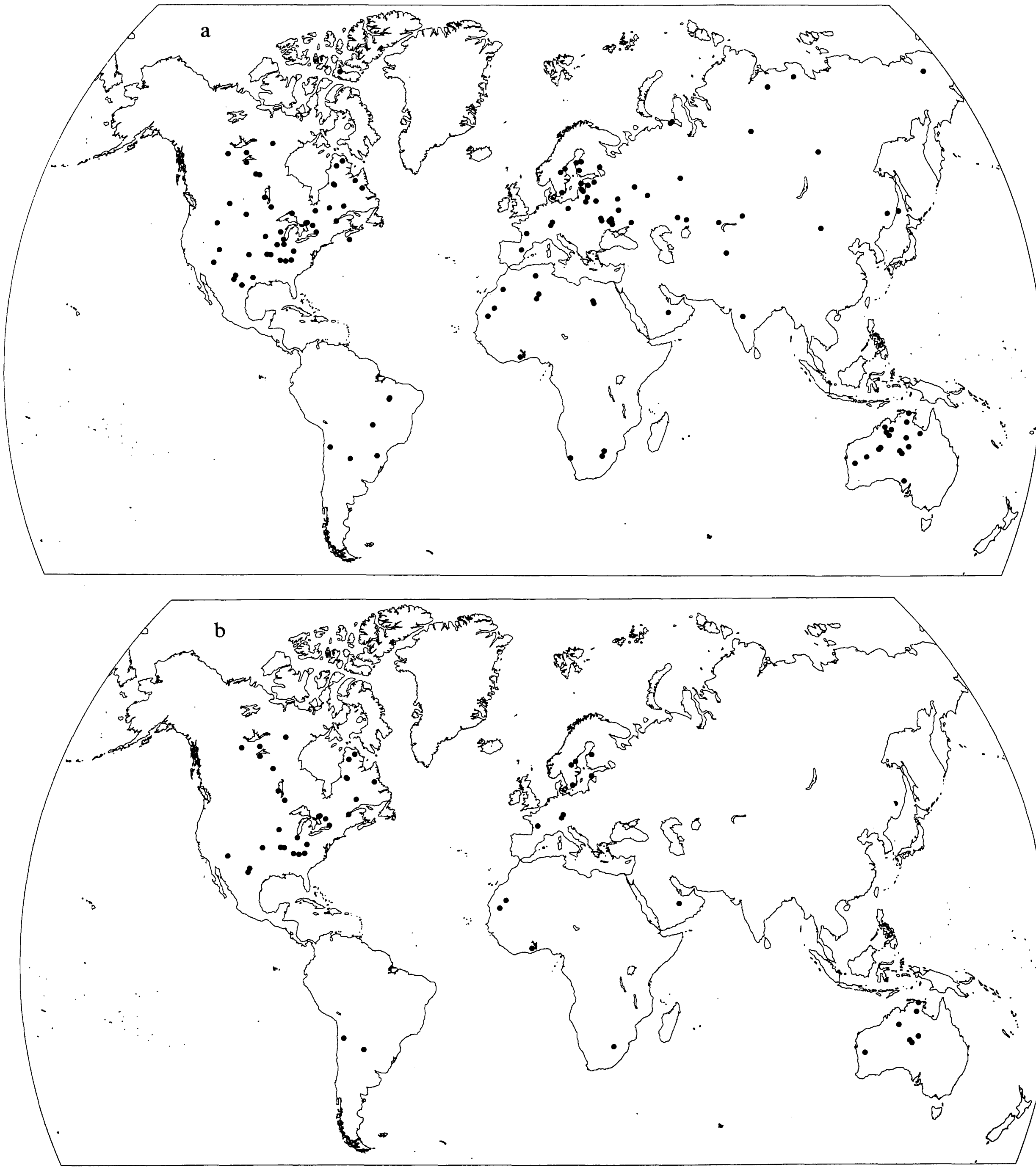

FIG. 1. World distribution of known impact craters a) current status: note concentrations in cratonic areas of Australia, North America and European USSR b) Dence (1972): compare with Fig. 1a.

be unrecognizable in the terrestrial record in as little as 120 Ma (Grieve, 1984). This is a very general statement and actual lifetimes for specific craters are dependent on local geologic history. The complication of post-impact burial can result in the preservation of relatively small craters, *e.g.*, Kelly West, D = 10 km (Shoemaker *et al.*, 1990) and Janisjärvi, D = 14 km (Masaitis *et al.*, 1980) are both of Proterozoic age (Table 1).

Only 25% of the known craters have been dated isotopically, generally by analyses of their impact melt rocks. Many others, however, have relatively well constrained bio-stratigraphic ages, generally from analyses of post-impact, crater-filling sediments. The remainder generally have minimum age estimates based on the age of the target rocks (Table 1). There is a bias in the record towards young structures, with 50% being less than 200 Ma old (Fig. 2). This reflects the low retention rate for craters due to active terrestrial processes. This bias towards young structures occurs in subsets of the known sample. In some cases, however, there may be a second order structure to the distri-

TABLE 1. Known terrestrial impact craters.

Name	Country	Diameter (km)	Age (Ma)	Lat.	Long.
Acraman	Australia	160	>570	S32°1′	E135°27′
Amguid	Algeria	0.45	<0.1	N26°5′	E4°23′
Aouelloul	Mauritania	0.39	3.1 ± 0.3	N20°15′	W12°41′
Araguainha Dome	Brazil	40	<249 ± 19	S16°46′	W52°59′
Azuara	Spain	30	<130	N41°10′	W0°55′
B.P. Structure	Libya	2.8	<120	N25°19′	E24°20′
Barringer	USA	1.2	0.049 ± 0.003	N35°2′	W111°1′
Beaverhead	USA	15	~600	N45°0′	W113°0′
Bee Bluff	USA	2.4	<40	N29°2′	W99°51′
Beyenchime-Salaatin	USSR	7.5	<65	N71°50′	E123°30′
Bigach	USSR	7	6 ± 3	N48°30′	E82°0′
Boltysh	USSR	25	88 ± 3	N48°45′	E32°10′
Bosumtwi	Ghana	10.5	1.3 ± 0.2	N6°32′	W1°25′
Boxhole	Australia	0.17	0.03	S22°37′	E135°12′
Brent	Canada	3.8	450 ± 3	N46°5′	W78°29′
Campo del Cielo	Argentina	0.05	<0.004	S27°38′	W61°42′
Carswell	Canada	39	115 ± 10	N58°27′	W109°30′
Charlevoix	Canada	54	357 ± 15	N47°32′	W70°18′
Chiyli	USSR	3	46 ± 7	N49°10′	E57°51′
Clearwater Lake East	Canada	22	290 ± 20	N56°5′	W74°7′
Clearwater Lake West	Canada	32	290 ± 20	N56°13′	W74°30′
Connolly Basin	Australia	9	<60	S23°32′	E124°45′
Crooked Creek	USA	7	320 ± 80	N37°50′	W91°23′
Dalgaranga	Australia	0.021	0.025	S27°43′	E117°5′
Decaturville	USA	6	<300	N37°54′	W92°43′
Deep Bay	Canada	13	100 ± 50	N56°24′	W102°59′
Dellen	Sweden	15	109.6 ± 1.0	N61°48′	E16°48′
Des Plaines	USA	8	<280	N42°3′	W87°52′
Dobele	USSR	4.5	300 ± 35	N56°35′	E23°15′
Eagle Butte	Canada	19	<65	N49°42′	W110°30′
El'gygytgyn	USSR	18	3.5 ± 0.5	N67°30′	E172°5′
Flynn Creek	USA	3.55	360 ± 20	N36°17′	W85°40′
Glasford	USA	4	<430	N40°36′	W89°47′
Glover Bluff	USA	10	<500	N43°58′	W89°32′
Goat Paddock	Australia	5.1	<50	S18°20′	E126°40′
Gosses Bluff	Australia	22	142.5 ± 0.5	S23°50′	E132°19′
Gow Lake	Canada	4	<250	N56°27′	W104°29′
Gusev	USSR	3.5	65	N48°21′	E40°14′
Haughton	Canada	20.5	21.5 ± 1.2	N75°22′	W89°40′
Haviland	USA	0.011	0	N37°35′	W99°10′
Henbury	Australia	0.157	<0.005	S24°35′	E133°9′
Holleford	Canada	2.35	550 ± 100	N44°28′	W76°38′
Ile Rouleau	Canada	4	<300	N50°41′	W73°53′
Ilumetsy	USSR	0.08	>0.002	N57°58′	E25°25′
Ilyinets	USSR	4.5	395 ± 5	N49°6′	E29°12′
Janisjärvi	USSR	14	698 ± 22	N61°58′	E30°55′
Kaalijärvi	USSR	0.11	0.004 ± 0.001	N58°24′	E22°40′
Kaluga	USSR	15	380 ± 10	N54°30′	E36°15′
Kamensk	USSR	25	65 ± 2	N48°20′	E40°15′
Kara	USSR	65	73 ± 3	N69°5′	E64°18′
Kar-Kul	USSR	45	<225	N38°57′	E73°24′
Karla	USSR	12	10 ± 5	N54°54′	E48°0′
Kelly West	Australia	10	>550	S19°56′	E133°57′
Kentland	USA	13	<300	N40°45′	W87°24′
Kjardla	USSR	4	455	N57°0′	E22°42′
Kursk	USSR	5.5	250 ± 80	N51°40′	E36°0′
Lac Couture	Canada	8	430 ± 25	N60°8′	W75°20′
Lac la Moinerie	Canada	8	400 ± 50	N57°26′	W66°36′
Lappajärvi	Finland	17	77.3 ± 0.4	N63°12′	E23°42′
Lawn Hill	Australia	18	>515	S18°40′	E138°39′
Liverpool	Australia	1.6	150 ± 70	S12°24′	E134°3′
Logancha	USSR	20	50 ± 20	N65°30′	E95°48′
Logoisk	USSR	17	40 ± 5	N54°12′	E27°48′
Lonar	India	1.83	0.052 ± 0.006	N19°58′	E76°31′
Macha	USSR	0.3	<0.007	N59°59′	E118°0′
Manicouagan	Canada	100	212 ± 2	N51°23′	W68°42′
Manson	USA	35	65.7 ± 1.0	N42°35′	W94°31′
Marquez Dome	USA	15	58 ± 2	N31°17′	W96°18′
Middlesboro	USA	6	<300	N36°37′	W83°44′

TABLE 1. Continued.

Name	Country	Diameter (km)	Age (Ma)	Lat.	Long.
Mien	Sweden	9	121.0 ± 2.3	N56°24′	E14°54′
Misarai	USSR	5	395 ± 145	N54°0′	E23°54′
Mishina Gora	USSR	4	<360	N58°40′	E28°0′
Mistastin	Canada	28	38 ± 4	N55°53′	W63°18′
Montagnais	Canada	45	50.5 ± 0.76	N42°53′	W64°13′
Monturaqui	Chile	0.46	1	S23°56′	W68°17′
Morasko	Poland	0.1	0.01	N52°29′	E16°54′
New Quebec	Canada	3.44	1.4 ± 0.1	N61°17′	W73°40′
Nicholson Lake	Canada	12.5	<400	N62°40′	W102°41′
Oasis	Libya	11.5	<120	N24°35′	E24°24′
Obolon'	USSR	15	215 ± 25	N49°30′	E32°55′
Odessa	USA	0.168	<0.05	N31°45′	W102°29′
Ouarkziz	Algeria	3.5	<70	N29°0′	W7°33′
Piccaninny	Australia	7	<360	S17°32′	E128°25′
Pilot Lake	Canada	5.8	445 ± 2	N60°17′	W111°1′
Popigai	USSR	100	35 ± 5	N71°30′	E111°0′
Presqu'ile	Canada	12	<500	N49°43′	W78°48′
Pretoria Salt Pan	South Africa	1.13	0.2	S25°24′	E28°5′
Puchezh-Katunki	USSR	80	220 ± 10	N57°6′	E43°35′
Ragozinka	USSR	9	55 ± 5	N58°18′	W62°0′
Red Wing	USA	9	200 ± 5	N47°36′	W103°33′
Riachao Ring	Brazil	4.5	<200	S7°43′	W46°39′
Ries	Germany	24	14.8 ± 0.7	N48°53′	E10°37′
Rochechouart	France	23	186 ± 8	N45°50′	E0°56′
Roter Kamm	Namibia	2.5	3.7 ± 0.3	S27°46′	E16°18′
Rotmistrovka	USSR	2.7	140 ± 20	N49°0′	E32°0′
Sääksjärvi	Finland	5	514 ± 12	N61°23′	E22°25′
Saint Martin	Canada	40	220.5 ± 18	N51°47′	W98°32′
Serpent Mound	USA	6.4	<320	N39°2′	W83°24′
Serra da Canghala	Brazil	12	<300	S8°5′	W46°52′
Shunak	USSR	3.1	12	N47°12′	E72°42′
Sierra Madera	USA	13	<100	N30°36′	W102°55′
Sikhote Alin	USSR	0.027	0	N46°7′	E134°40′
Siljan	Sweden	55	368 ± 1.1	N61°2′	E14°52′
Slate Islands	Canada	30	<350	N48°40′	W87°0′
Sobolev	USSR	0.053	0	N46°18′	E138°52′
Soderfjärden	Finland	6	550	N63°2′	E21°35′
Spider	Australia	13	>570	S16°44′	E126°5′
Steen River	Canada	25	95 ± 7	N59°31′	W117°38′
Steinheim	Germany	3.8	14.8 ± 0.7	N48°2′	E10°4′
Strangways	Australia	25	<470	S15°12′	E133°35′
Sudbury	Canada	200	1850 ± 3	N46°36′	W81°11′
Tabun-Khara-Obo	Mongolia	1.3	>120	N44°6′	E109°36′
Talemzane	Algeria	1.75	<3	N33°19′	E4°2′
Teague	Australia	28	1685 ± 5	S25°52′	E120°53′
Tenoumer	Mauritania	1.9	2.5 ± 0.5	N22°55′	W10°24′
Ternovka	USSR	12	330 ± 30	N48°1′	E33°5′
Tin Bider	Algeria	6	<70	N27°36′	E5°7′
Upheaval Dome	USA	5	<65	N38°26′	W109°54′
Ust-Kara	USSR	25	73 ± 3	N69°18′	E65°18′
Vargeao Dome	Brazil	12	<70	S26°50′	W52°7′
Veevers	Australia	0.08	<1	S22°58′	E125°22′
Vepriaj	USSR	8	160 ± 30	N55°6′	E24°36′
Vredefort	South Africa	140	1970 ± 100	S27°0′	E27°30′
Wabar	Saudi Arabia	0.097	0.006 ± 0.002	N21°30′	E50°28′
Wanapitei Lake	Canada	7.5	37 ± 2	N46°45′	W80°45′
Wells Creek	USA	14	200 ± 100	N36°23′	W87°40′
West Hawk Lake	Canada	3.15	100 ± 50	N49°46′	W95°11′
Wolfe Creek	Australia	0.875	<0.3	S19°18′	E127°47′
Zapadnaya	USSR	4	115 ± 10	N49°44′	E29°0′
Zeleny Gai	USSR	2.5	>140	N48°42′	E32°54′
Zhamanshin	USSR	13.5	0.9 ± 0.1	N48°24′	E60°58′

bution of ages. For example, in the USSR there are a relatively large number of structures with ages in the 200–400 Ma range. These are buried structures and have been protected from erosion.

The sizes of known terrestrial impact craters range up to 200 km (Sudbury, Table 1). Although previously listed as 140 km (Grieve and Robertson, 1987), recent analyses of the spatial distribution of shock features and outliers of original cover rocks

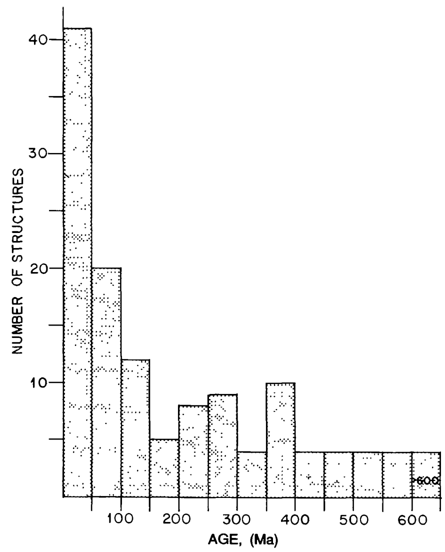

FIG. 2. Histogram, in 50 Ma bins, of ages of known terrestrial impact craters. Note decrease in number with age due to terrestrial processes.

at Sudbury suggest that it may have been originally as large as 200 km (Lakomy, 1990; Stöffler *et al.,* 1989). This illustrates another characteristic of the terrestrial record. In most cases, original rim diameters are reconstructed estimates (Fig. 3). For structures with minimal information, there is considerable uncertainty in individual estimates. For example, Shoemaker and Shoemaker (1990) estimate the original diameter of Kelly West (Table 1) to be anywhere between 10 and 20 km. In some cases, erosion has removed virtually all of the topographic relief associated with the original crater and all that remains is a circular geological anomaly reflecting the original central uplifted area, *e.g.,* Carswell, Siljan, Vredefort, and others. In other cases, there may be a remanent central peak. For example, Gosses Bluff is expressed topographically as a 5 km annular ring of erosionally resistant sandstones (Fig. 3) but other geological and geophysical data suggest the original diameter was of the order of 22 km (Milton *et al.,* 1972; Table 1).

There is a general bias in the sizes of known terrestrial craters. At large diameters, the cumulative size-frequency distribution can be approximated by a distribution similar to a production distribution on other terrestrial planets (Fig. 4). In the terrestrial case, it is more likely to be a steady-state distribution, rather than a production distribution. At diameters below ~20 km, the number of known craters falls below this distribution, with an increasing deficit of craters at smaller diameters. This reflects the relatively greater ability of erosion to remove the topographic and geologic signature of smaller craters and the inherent difficulties associated with recognizing smaller structures. Comparison of the size-frequency distribution of the currently known record with that calculated from previous listings of known craters suggests that this distribution is a basic characteristic of the terrestrial record (Fig. 4) and that it is not going to change dramatically as new structures are identified.

Planetary impact craters are recognized by their morphology. Terrestrial impact craters are recognized by their morphology and geologic structure but, in some cases, are recognized solely by geologic structure. Terrestrial craters show the basic progression of simple to complex forms with increasing diameter, as well as secondary forms such as central peak craters, peak ring craters and ring basins, observed on other terrestrial planets (Wood and Head, 1976). Caution must be exercised, however, in comparing the modified morphologic elements of craters within the terrestrial record and, in particular, comparing terrestrial and planetary craters (Pike, 1985). Original morphologic elements can be enhanced, modified or removed by the various types of erosional processes on earth. For example, Haughton (Table 1) has topographic rings, which has led to its description as a peak ring structure (Robertson and Sweeney, 1983). More recent analyses indicate that these rings are not reflected in the geology or geophysical data (Grieve, 1988) and suggest that they may be erosional artifacts rather than primary features. In addition, similar-sized craters can have different morphologies, depending on the depth of erosion and the degree to which target lithologies have contributed to differential erosion (Fig. 3).

Morphometric relations are affected by erosion to varying degrees. For example, the correlation between the physical height of the central uplift and crater diameter for terrestrial craters is poor, in contrast to the good correlation in lunar data (Pike, 1977). Only the most general conclusion can be reached: namely, that larger craters have higher central peaks (Fig. 5a). The lack of correlation reflects the variable erosion of topography. Other morphometric parameters are better defined; there is a more obvious correlation between the diameter of the central uplift and crater diameter for terrestrial craters (Fig. 5b). This is because the diameter of the central uplift, in addition to being defined topographically, can be defined geologically by mapping the occurrence of uplifted target rocks, even when the topographic expression has been completely removed (Fig. 3).

The intent of the discussion above has been not only to outline the basic character of the terrestrial impact record but also to emphasize that it differs from that on other planets in the degree to which it is complicated by endogenic geologic processes. Data compilations, such as in Table 1, should not be used without understanding the underlying character of the data. Unfortunately, this is not always the case. For example, much of the discussion over whether or not the terrestrial impact record contains a periodic signal deals with terrestrial data compilations as if they were planetary data (Rampino and Stothers, 1984) and fails to take account of the complications due to terrestrial processes. Terrestrial processes do have a positive side. Differential erosion of similar-sized structures, for example, allows direct observation of different structural levels without drilling. Such observations are required to build an integrated model of the third dimension of crater structure. It is impossible to give an in-depth discussion of all the attributes of terrestrial impact craters. Accordingly, what follows is a relatively brief review of some of the parameters and concepts of impact cratering that have been constrained by ground truth observations from the terrestrial impact record.

CRATERING MECHANICS

The basic physics behind crater excavation are well described (Gault *et al.,* 1968; Melosh, 1989). The cratering flow field induced in the target rocks has been determined experimentally (Gault *et al.,* 1968) and duplicated in continuum mechanics calculations (Orphal *et al.,* 1980). A simplified analytical description of particle movements during the formation of explosion craters has been described by Maxwell (1977) and applied to impact craters (Croft, 1980). Observational data are generally consistent with these models (Grieve *et al.,* 1981) but there is limited commonality in a temporal sense. Code calculations, analytical descriptions and experimental observations are dynamic and follow cavity growth with time. Observations at craters deal only with the final structure; although, it is possible to reconstruct motions to some degree if stratigraphic markers are present. A central concept to all cratering mechanics is the so-called transient cavity. The transient cavity is not observed directly in observational data; although, it was the interpretation of terrestrial data that originally defined the concept. As noted by Gault *et al.* (1968), however, when discussing the results of laboratory-scale experimental craters, "it is investigations of natural craters that (will) clarify our understanding of cratering mechanics".

Transient Cavity Formation

A typical simple crater consists of a bowl-shaped depression with an upraised rim. Terrestrial simple craters have an apparent depth to diameter ratio of ~0.12. They are shallower than lunar craters, where the depth to diameter ratio is ~0.20 (Pike, 1977). Beneath the surface of the apparent crater, there is a bowl-shaped breccia lens (Fig. 6). This lens does not occur in experimental craters and is not observed directly in planetary craters, where only surface information is available. Drilling and subsequent petrographic analyses of material from terrestrial craters indicates that the material in this breccia lens is mixed and largely unshocked; although, materials shocked up to and including melting occur, often in lenses and concentrated towards the base of the breccia lens (Shoemaker, 1960; Dence *et al.,* 1977). Beneath the breccia lens are autochthonous target rocks shocked to ~25 GPa (Robertson and Grieve, 1977).

As shock pressures attenuate radially from the point of impact, the apparent reversal in shock stratigraphy with the largely unshocked breccia lens overlying shocked target rocks in the crater floor indicates that the breccia lens materials are allochthonous. The generally low shock level of the breccia lens, however, indicates clearly that it is not fall back material. The formation of the breccia lens is a relatively late-time phenomenon in the cratering process and is not modelled in finite difference code calculations. Thus, its fundamental allochthonous nature is not always appreciated in the ballistic extrapolations of code calculations, which attempt to model the breccia lens by *in situ* fragmentation (Roddy *et al.,* 1980). The consensus is that the breccia lens is formed by slumping of the walls of the initial cavity formed by the cratering flow field. It is difficult to trace the origins of this concept but over the years the role of cavity slumping has featured more prominently in models of crater formation (Shoemaker, 1960; Dence, 1968; Dence *et al.,* 1977; Grieve *et al.,* 1977).

The initial cavity formed directly by the cratering flow field, which is more or less equivalent to that produced in experiments (Gault *et al.,* 1968; Stöffler *et al.,* 1975), is referred to as the transient cavity. As with breccia lens formation, the concept does not have a single source. It is explicitly discussed in Dence (1968), where it is referred to as the primary cavity. From drilling data at Barringer and Brent, Dence (1973) proposed that the transient cavity at terrestrial simple craters was parabolic in cross-section and had a depth-diameter ratio of ~0.33 (Fig. 6). This is somewhat deeper than laboratory experiments in sand, where the depth-diameter ratio is ~0.20–0.25 (Gault *et al.,* 1968; Stöffler *et al.,* 1975). This is presumably due to rock strength effects limiting radial growth in natural craters.

The Dence (1973) model for transient cavity shape (Fig. 7) has been tested by assuming that the floor of simple craters beneath the breccia lens approximates the floor of the transient cavity and that the only modification occurs through transient cavity wall failure (Grieve and Garvin, 1984; Grieve *et al.,* 1989). The testing essentially compares the observed volume of the breccia lens with that calculated from wall slumping and is numerically highly sensitive to small changes in dimensions. Calculated and observed results compare favourably. They do, however, exemplify another characteristic of the record. Of the seventeen known terrestrial simple craters with D ≥ 1 km, only seven had sufficient morphometric data to test the model and of these four had depth estimates based on gravity models not drilling data. This observation indicates that while the community may be discovering several new craters per year, it is not necessarily undertaking the detailed and in-depth studies at individual craters required to utilize the full potential of the ground truth data in the terrestrial record.

Complex Structures

Complex structures are characterized by an uplifted center, annular trough and outer faulted rim (Fig. 8). Terrestrial complex structures occur above diameters greater than 2 km and 4 km when the target rocks are sedimentary and crystalline, respectively. They represent a highly modified crater form, relative to simple craters, and have considerably smaller depth-diameter ratios (Fig. 8). Traditionally, it has been more difficult to convince some workers of the impact origin of complex structures compared to the more obvious craterform of simple craters. As complex structures cannot be duplicated readily in laboratory experiments and code calculations, the nature of cratering mechanics at complex structures has been more debatable. Several workers have suggested that the complex craterform results from the impact of a low density cometary body (*e.g.,* Roddy,

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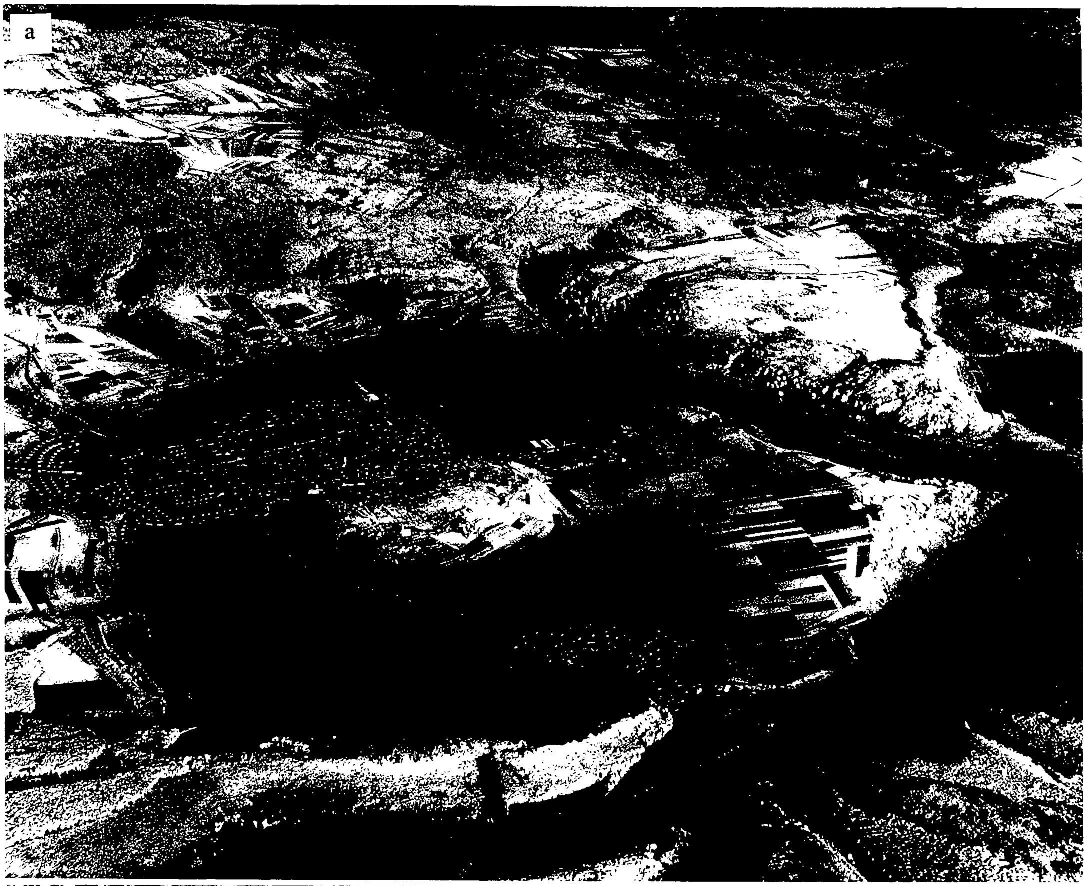
FIG. 3. Examples of terrestrial complex craters with varying appearance due to terrestrial processes. (a) Steinheim (D = 3.8 km, 14.8 Ma old) is partially filled by post impact sediments but original rim and central uplift are still well defined. (b) Haughton (D = 20.5 km, 21.5 Ma old) has the bulk of its post-impact sediments removed by erosion, exposing the interior breccia sheet. A central uplift is not visible and the rim area is not well defined morphologically. (c) Mistastin (D = 28 km, 38 Ma old) has no post-impact sediments and the bulk of the impact lithologies (impact melt and breccias) have been removed by glacial erosion. The target rocks of the floor of the crater and the central uplift are exposed and morphological expression of the original rim area is not present. (d) Gosses Bluff (D = 22 km, 142.5 Ma old) is eroded below the original crater floor. Essentially no allochthonous impact lithologies remain and the crater is expressed morphologically by a 5 km diameter annulus of hills consisting of erosionally resistant sandstones contained in the original central uplift. The original rim area is not expressed morphologically, although some expression of the original 22 km diameter structure is visible as annuli with albedo differences compared to the surrounding terrain.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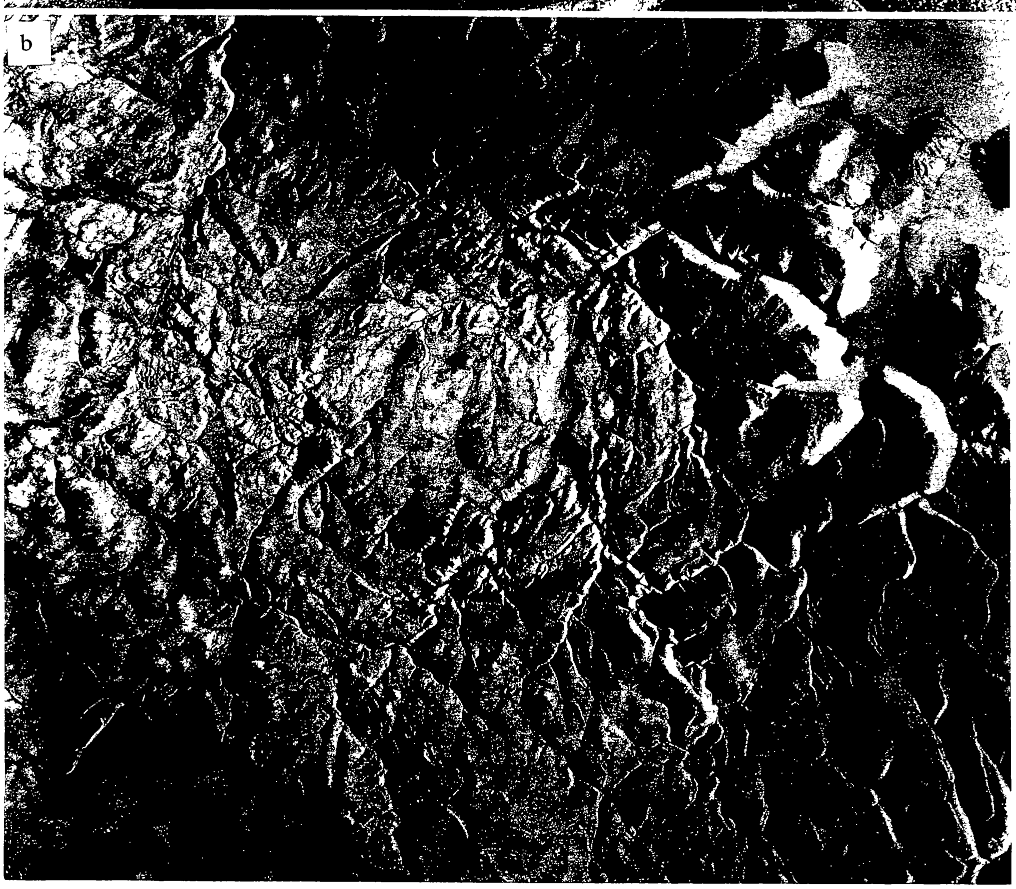
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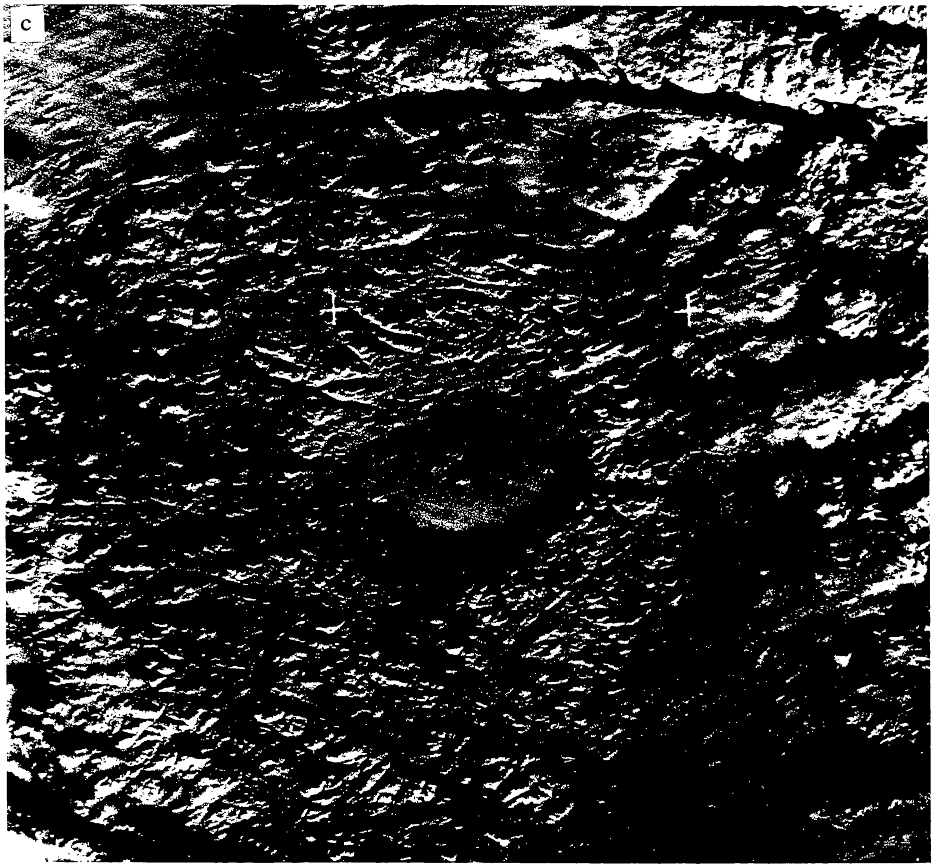
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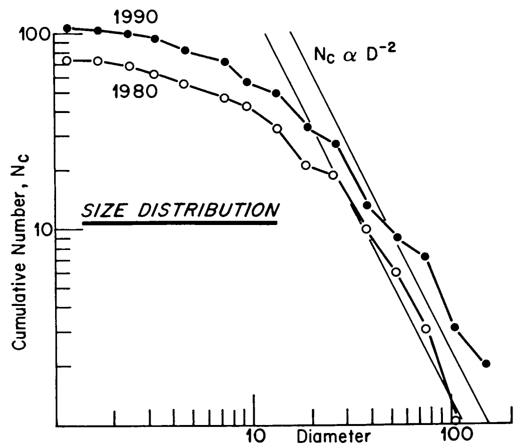

FIG. 4. Log cumulative number (N_c) of Phanerozoic-aged terrestrial impact craters against log diameter (D). Note deviation from $N_c \alpha D^{-2}$ distribution and deficit of craters at small diameters. This characteristic is present in both 1980 and 1990 compilations of the terrestrial cratering record.

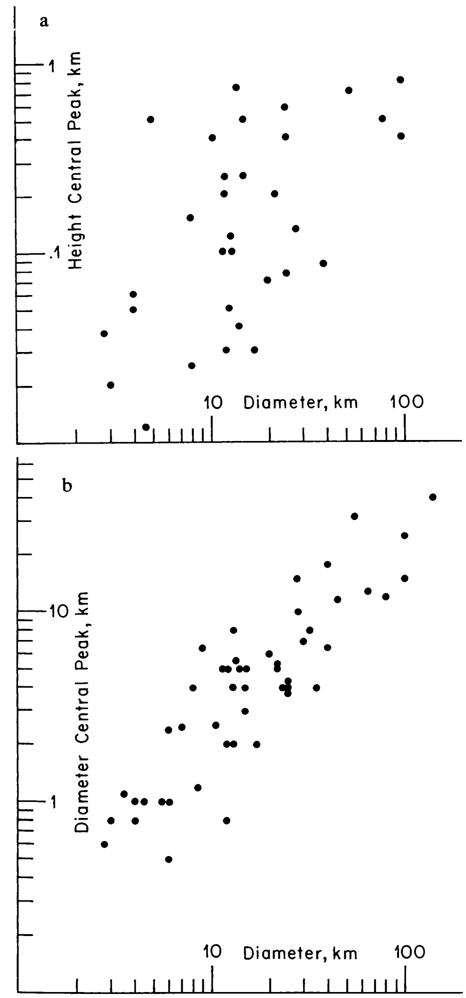

FIG. 5. a) Log height of central structures at complex craters against log diameter. Note lack of anything but the most general relationship. b) Log diameter of central structures at complex craters against log diameter. Note a better defined relationship than with log height of central structures (Fig. 5a).

1968). Others have suggested that somehow cratering mechanics fundamentally differ in complex craters and they undergo relatively shallower excavation compared to simple craters (Head *et al.,* 1975).

Observations at terrestrial complex craters indicate clearly that material in central structures is uplifted from depth and that the amount of uplift is a function of crater diameter. Similarly, complex craters with stratigraphic markers indicate that relatively deeper lithologies are missing only in the central area (Dence *et al.,* 1977). Excavation does not occur all the way out to the rim, with the diameter of obvious excavation being limited to 0.50–0.65 the final diameter (Grieve *et al.,* 1981). By reconstructing the cavity through repositioning beds in their original stratigraphic positions, it has been established that the original or transient cavity formed by the cratering flow field at complex craters had a geometry essentially equivalent to that at simple craters. This is the concept of proportional growth (Croft, 1981), as opposed to non-portional growth, where the transient cavity geometry varies with crater diameter (Head *et al.,* 1975). Although somewhat slow to accept the relevance of terrestrial data, the planetary community has now generally accepted this concept. The final turning point was the results of code calculations confirming a "deep" transient cavity at large crater diameters (O'Keefe and Ahrens, 1987). It is somewhat distressing, but not unexpected, that concensus on transient cavity geometry at large structures based on terrestrial observations was not accepted by the larger cratering community until some members of the various sub-disciplines (*e.g.,* planetary geologists, computer code modelers) had each, in turn, reached the same conclusion from their data or models.

With a general consensus on the first order attributes of cratering mechanics at larger craters, it is important to remember that there are also important second order effects. For example, as particles set in motion are affected in gravity, materials that may be excavated in small structures will have insufficient initial velocities to be excavated beyond the final rim at larger structures. Thus, larger structures are less efficiently excavated. As a result, the relative volume of impact melt retained in larger structures is greater (Lange and Ahrens, 1979), as confirmed by terrestrial observations. Similarly, impact melt volumes and transient cavity volumes do not scale at the same rate (Cintala and Grieve, 1984) and at very large structures the melt volume may intersect the base of the transient cavity (Melosh, 1989; Cintala and Grieve, 1991). This may affect final crater form (Croft, 1983; Cintala and Grieve, 1991).

Depth of Excavation

Observations at terrestrial craters have the potential to provide a direct measure of the depth of excavation. Unfortunately, very few craters have preserved ejecta blankets and, in the few cases available, stratigraphic markers are not necessarily present at the appropriate depths. The only simple crater where there is a well established stratigraphic section and preserved ejecta is Barringer. Here, blocks of Coconino sandstone in the ejecta (Shoemaker, 1960) give a depth of excavation (d_e) estimate of >0.08 D. There are only two complex craters with similar data. The Ries (Pohl *et al.,* 1977) and Ragozinka (Vishnevsky and

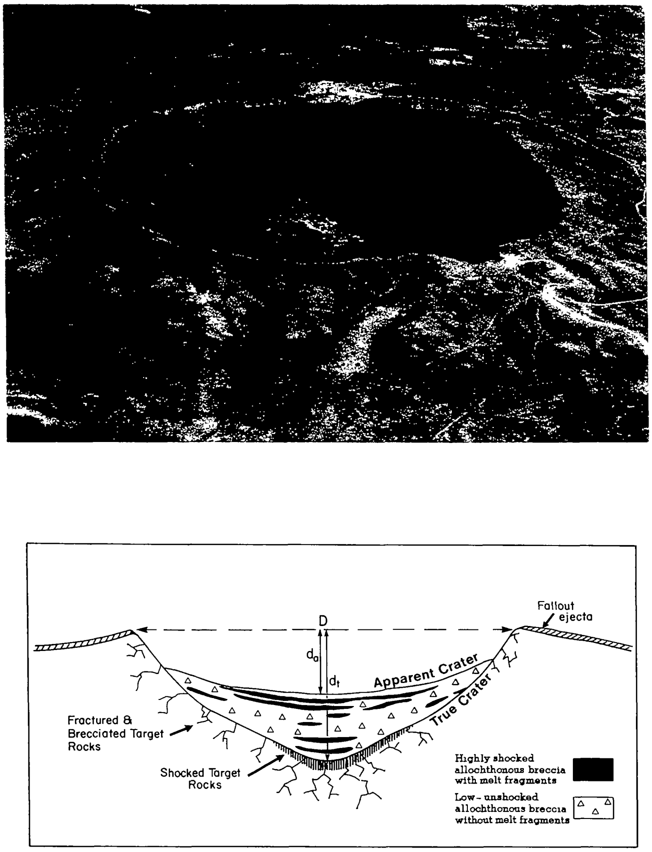

FIG. 6. (Top) Oblique aerial photograph of Barringer Crater, the canonical example of a young simple crater. (Bottom) Schematic cross section of a simple crater, showing the apparent (d_a) and true (d_t) depths and the interior breccia lens.

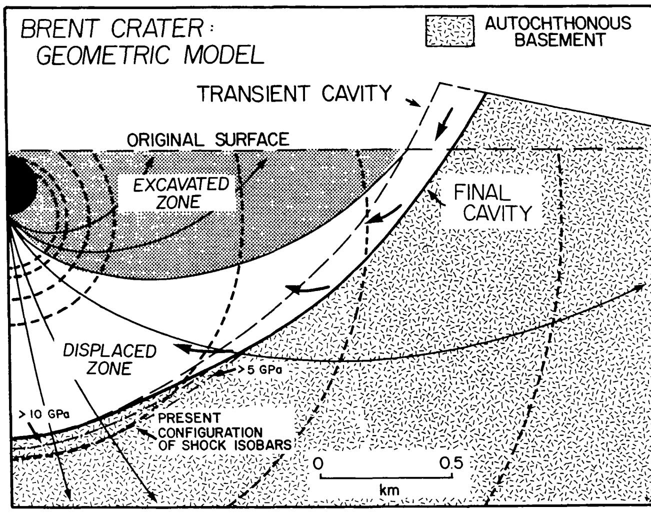

FIG. 7. Model of transient cavity formation and geometry for simple craters as exemplified by Brent. Note excavated and displaced zones, including downward displacement of shock isobars, followed by inward collapse of the transient cavity wall.

Lagutenko, 1986) both give d_e estimates of >0.03 D. There are a few other structures with preserved allochthonous breccia deposits, which have been interpreted as ejecta. Accepting such an interpretation, for example, gives d_e estimates of 0.13 D from both Decaturville (Offield and Pohn, 1977) and Steinheim (Reiff, 1977). These breccia deposits, however, are within the craters and may be analogous to those at other structures, such as Haughton, which are interpreted not as fallback but allochthonous material that never left the crater (Grieve, 1988). Croft (1980) estimated the depth of excavation from small scale laboratory and explosion experiments, as well as the results of analytical calculations, to be 0.1 D. The terrestrial observations are not inconsistent with his estimate. Particularly, as his estimate refers to transient cavity diameter and, in the case of complex terrestrial craters, transient cavity diameter is ~0.50–0.65 D, final crater diameter (Grieve *et al.*, 1981; Lakomy, 1990).

The depth of excavation is highly relevant in considering the provenance of lunar highland samples and their interpretation in terms of the lithological nature of the third dimension of the lunar crust. This applies particularly to those samples believed to be ejecta from the large multi-ring basins (Ryder and Wood, 1977). This has been controversial with, for example, maximum depth of excavation estimates for the Imbrium Basin varying from 30 km (Head *et al.*, 1975) to 60 km (Dence, 1976). In part, these widely varying estimates result from a misunderstanding of terminology (*e.g.*, the depth of excavation is not equivalent to the depth of the transient cavity (Grieve, 1980)) and of the concept that the diameter of excavation during transient cavity formation is considerably less than the final diameter. More

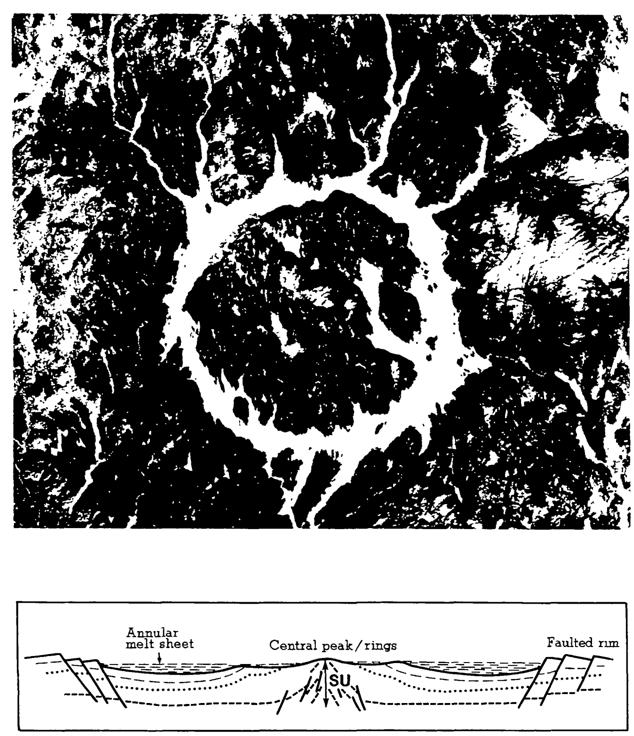

FIG. 8. LANDSAT image of the Manicouagan crater, the largest Phanerozoic aged crater in N. America. The annular lake is ~70 km in diameter and surrounds an inner plateau capped by impact melt rocks with an emergent central uplift. (Bottom) Schematic cross section of a large complex crater, showing faulted rim area, annular trough and central uplift structure. Note the shallower depth-diameter ratio compared to simple craters (Fig. 6) and the amount of structural uplift (SU) in the center.

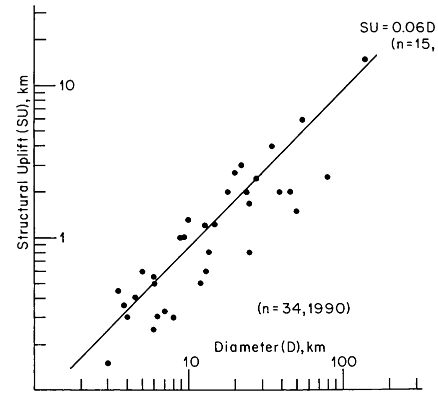

FIG. 9. Log structural uplift (SU) at complex craters against log diameter. Relationship SU = 0.06 $D^{1.1}$ is from Grieve *et al.* (1981) and was defined on the basis of 15 data points. More recent data are consistent with this relationship.

recent lunar work, using orbital geochemical data and sample petrogenesis (Spudis *et al.,* 1987), appears to be reaching a consensus with the experimental and terrestrial data.

While this approaching concensus is gratifying, depth estimates are still constrained by very few data. It is unlikely that this situation will improve with new discoveries in the terrestrial record. Young craters, which are likely to have preserved ejecta, are the most easy to recognize in the record. They, therefore, are most likely to have been already identified (Fig. 2). In addition, larger terrestrial complex craters yet to be discovered, which are of the most interest as analogies to large planetary craters, will have generally penetrated the sedimentary cover into basement. Thus, they will only supply minimum depth estimates. The most useful structures will be complex structures within the sedimentary column, which were buried soon after formation. To obtain the relevant information from the ejecta, the structures will have to be drilled off-structure, which has not been a priority in the exploration of buried impact structures. There is, however, another indirect method of constraining the depth of excavation; namely by considering the amount of structural uplift that has taken place in the centers of complex structures.

Structural Uplift

From terrestrial data, Grieve *et al.* (1981) suggested that SU = 0.06 $D^{1.1}$, where SU was the observed amount of uplift undergone by the deepest horizon now exposed in the center of complex structures (Fig. 8). This relationship was based on data from fifteen structures and, as more data have become available, the relationship has not basically changed (Fig. 9). The amount of uplift in large structures is considerable and the largest structures can expose lower crustal material at the surface, *e.g.,* at Vredefort (Slawson, 1976). On other planetary bodies, remote sensed observations of central peaks can provide insights into the composition of the crust at depth (Pieters, 1982) with, at least, the relative depth involved being constrained by the terrestrial relationship.

The amount of stratigraphic uplift is also an indirect measure of the depth of excavation. The target rock units, which lay stratigraphically above the uplifted material, are no longer present and were removed during the cratering process. The main agent for the removal of material is excavation. This is somewhat of an oversimplification. Some target rock material, such as that which becomes incorporated in impact sheets and breccia deposits, is allochthonous but is not ejected and the maximum depth of excavation occurs off-center, according to the geometry of the cratering flow field (Fig. 1 in Croft, 1981). Nevertheless, given the very small number of direct observations of depth of excavation, the amount of structural uplift provides some additional constraints. As the diameter of excavation at complex structures is 0.50–0.65 D, using SU to approximate d_e translates to $d_e \sim 0.09$–0.12 the diameter of excavation, which is consistent with constraints on d_e derived from experimental craters, calculations and direct observations.

SHOCK METAMORPHISM

The occurrence of diagnostic shock metamorphic effects is the principal criterion for the recognition of terrestrial impact craters. As noted earlier, terrestrial impact craters differ in this respect from other planetary craters, which are recognized solely on the basis of morphology. The recognition of diagnostic shock effects had a dramatic effect upon the rate of identification of terrestrial craters (Fig. 10). It is not the intention to describe here details of shock metamorphic effects. The nature and occurrence of shock metamorphic effects are well documented in such compendia as French and Short (1968) and Roddy *et al.* (1977) and in individual works, most notably Stöffler (1971, 1972, 1974). Beyond the basis characterization of shock metamorphism, studies of shock metamorphic effects have also led to the definition of a number of principles. A few are discussed briefly here.

As estimated from the occurrence of particular shock effects in minerals calibrated against the results of shock recovery experiments, recorded shock pressures in the para-autochthonous target rocks of the crater floor have been determined for a number of structures; *e.g.,* Brent (Robertson and Grieve, 1977), Charlevoix (Robertson, 1975), Kursk (Basilevsky *et al.,* 1983), Manicouagan (Dressler, 1990), Slate Islands (Robertson and Grieve, 1977) and others. From these systematic studies, it is possible to estimate the attenuation of shock pressure with radial distance. For Brent, the only simple crater for which there are data, the observed attenuation rate is $\sim R^{-20}$. This, however does not take into account material motion and compression following the passage of the shock wave. When the section is restored to its original, precompression, length the attenuation rate is R^{-2}-R^{-3}, depending on the exact restoration (Robertson and Grieve, 1977). This is similar to that determined for far-field attenuation using equations of state formulations in finite difference calculations (Ahrens and O'Keefe, 1977). Similar rates are obtained from Charlevoix and the Slate Islands (Robertson and Grieve, 1977). These latter structures are complex craters and, in these cases, samples were from the surface. Therefore, observed attenuation rate had to be corrected for movements during transient cavity modification. At Kursk, observations were from a vertical sequence obtained from drill core through the central uplift and give similar results. In this case, the physical position of the samples had to be corrected for movement during uplift (Basilevsky *et al.,* 1983).

Knowledge of the rate of shock pressure attenuation and the relative spatial distribution of specific shock features with respect to morphologic elements permits estimation of such parameters as SU in crystalline targets, where there are no stratigraphic markers (Grieve *et al.,* 1981). It also permits reconstructions of the original crater diameter at heavily eroded structures (*e.g.,* Stöffler *et al.,* 1989; Hargraves *et al.,* 1990). Care, however, must be exercised. There is a degree of circularity in such reconstructions. The calculation of attenuation rates from observations at complex craters also involves an element of cavity reconstruction and some additional independent data, such as the occurrence of preserved pre-impact cover rocks, are required to have confidence in the reconstructions and estimates of the original crater dimensions (Grieve and Head, 1983).

These types of studies cannot be used in interpreting the significance of shock levels in the lunar samples, as they are all essentially allochthonous in nature. Breccia and impact melt deposits at terrestrial craters are, however, more amenable for analog studies. For example, Stöffler *et al.* (1979) have established a classification and nomenclature scheme for lunar highland rocks, which is based on textural analogies with terrestrial impact lithologies. As there is no terrestrial complex crater solely in crystalline targets with a well-preserved ejecta blanket, most of the analogies dealing with ejecta from large lunar craters and basins have centered on the suevite and Bunte breccia ejecta deposits of the Ries. The target at the Ries consists of ~600 m of Mesozoic sediments capping crystalline basement (Pohl *et al.,* 1977). Some of the observations and interpretations at the Ries, and other terrestrial impact structures, which are relevant to the lunar highlands samples and large planetary craters in general, include: (1) ejecta blankets are dominated by local material incorporated into the ejecta as it flows outward as ground surge following initial ballistic ejection (Hörz *et al.,* 1977; Hörz and Banholzer, 1980). This agrees with the models of Oberbeck (1975) for ballistic transport, erosion and sedimentation in the overall ejection process. (2) by far the bulk of the ejecta is material from near the surface. At the Ries, material identified as coming from depths >0.02 D is <0.1% of the mass of the ejecta (Hörz and Banholzer, 1980). This is a ready explanation for the paucity of "deep-seated" samples related to basin-sized impacts in the lunar highlands (Spudis *et al.,* 1987). (3) ejecta blankets are relatively cold at the time of their deposition, being dominated by unshocked near surface and local material from outside the crater and (4) details of inverted stratigraphy and regular radial variations in shock metamorphic effects, as suggested by some workers (Ahrens and O'Keefe, 1978), are not present, having been destroyed by late-stage, turbulent ground flow.

Similarly, studies of terrestrial impact melt rocks have provided constraints on the interpretation of lunar impact melt rocks. These include: (1) impact rocks from the same event may have variable texture, due to variations in cooling history, but are, in general, chemically similar (Grieve, 1980). This attribute has been used to divide lunar impact rocks into what are interpreted as genetically related groupings (Ryder and Spudis, 1987). (2) the chemistry of impact melt rocks corresponds to a mixture of the target rocks. In some cases, the composition of lunar impact rocks has been used to back calculate possible progenitors (Morris *et al.,* 1986). Mixing models, however, only require the solution to be numerically correct and such models should also be constrained by as much lithological information and geologic reasoning as possible. (3) the lithic and mineral clasts contained in impact melt rocks do not accurately reflect the mineralogy of the source area of the impact melt matrix. They

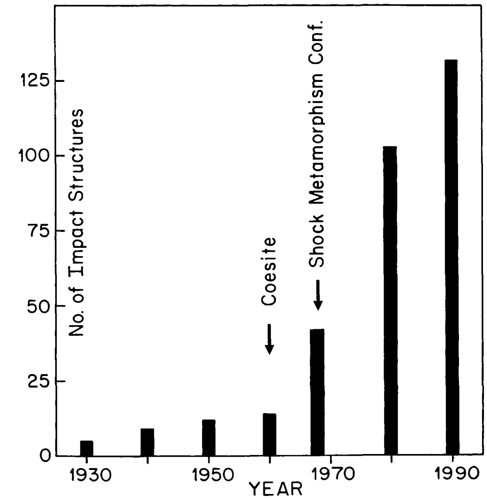

FIG. 10. Bar chart of number of known terrestrial impact craters with year. Note large increase in the rate of discovery with the recognition of shock metamorphic effects, such as coesite, and Conference on Shock Metamorphism of Natural Materials (French and Short, 1968).

represent clastic material incorporated during the movement of the melt to its present position and they are also biased towards more refractory types of clasts (McCormick *et al.,* 1989). This is an unfortunate characteristic of impact melt rocks, as it would simplify matters considerably if clast content mirrored the original lithologies in the melted volume of the target.

One of the tenets of the earlier literature on lunar samples is the resetting of isotopic systems as the result of impact. In most cases, this was interpreted as occurring due to thermal metamorphism in hot ejecta blankets. Thus, specific sample ages were related to the formation of a particular impact event (Maurer *et al.,* 1978). The terrestrial experience, however, indicates that shock metamorphism is not particularly efficient at resetting isotopic systems. Isotopic systems can be reset directly by shock but require pressures sufficient to result in melting (Bottomley *et al.,* 1989). At pressures below melting, partially reset ages are the norm (Jesseberger *et al.,* 1978; Scharer and Deutsch, 1989; Staudacher *et al.,* 1982). Extensive impact melt and hot impact deposits are largely confined to the interiors of craters. This suggests that the direct association of specific impact melt rocks and breccias and their ages with particular distant basins in the interpretation of lunar samples is open to considerable debate, a point made by Hörz and Banholzer (1980) but not yet generally recognized.

PROJECTILE IDENTIFICATION

Small terrestrial craters with physical remanants of the projectile are invariably formed by iron or stony iron bodies (Table 2). This is due to the atmospheric breakup of weaker stony bodies of similar dimension to iron projectiles that can penetrate the atmosphere (Melosh, 1981). Thus, the record of projectile types at small diameters is highly biased. At larger diameters, the impacting body is destroyed as a physical entity by the high shock pressures generated on impact. It can still be identified, in some cases, by its chemical signature now manifested as a siderophile element anomaly in impact melt rocks. Correlating siderophile element enrichments with particular impact events

TABLE 2. Known projectile types at terrestrial craters, ordered by increasing diameter.

Crater	Diameter, km	Projectile
Haviland	0.011	Pallasite
Dalgaranga	0.021	Mesosiderite
Sikhote Alin	0.027	IIAB
Campo del Cielo	0.050	IAB
Sobolev	0.053	Iron
Veevers	0.080	IIAB
Wabar	0.097	IIIAB
Morasko	0.100	IAB
Kaalijärvi	0.110	IAB
Henbury	0.157	IIIAB
Odessa	0.168	IAB
Boxhole	0.170	IIIAB
Macha	0.30	Iron
Aouelloul	0.39	Iron or Pallasite
Monturaqui	0.46	IAB
Wolfe Creek	0.87	IIIAB
Barringer	1.12	IAB

New Quebec	3.4	Chondrite
Brent	3.8	L or LL Chondrite
Gow	4	Iron?
Ilyinets	4.5	Iron
Sääksjärvi	5	Chondrite
Wanapitei	7.5	LL Chondrite
Mien	9	Stone?
Bosumtwi	10.5	Iron
Ternovka	12	Chondrite
Nicholson Lake	12.5	Nakhlite or Urelite?
Zhamanshin	13.5	Chondrite?
Dellen	15	Stone?
Obolon	15	Iron
Lappajärvi	17	Chondrite
El'gygytgyn	18	Achondrite
Clearwater East	22	Cl Chondrite
Rochechouart	23	Iron?
Ries	24	Aubrite
Boltysh	25	Chondrite
Kamensk	25	Chondrite
Ust-Kara	25	Chondrite
Mistastin	28	Iron?
Kara	65	Chondrite
Popigai	100	Chondrite

-------Represents division between physical evidence (smaller crater diameters) and solely chemical evidence (larger crater diameters) of projectile type. Some identifications, based on chemical evidence, are tentative or open to interpretation.

was first attempted for lunar samples (Gros *et al.,* 1976; Hertogen *et al.,* 1978). The need for examples was recognized and it was extended to samples of impact melt from terrestrial craters (Palme *et al.,* 1978; Wolf *et al.,* 1980), for which it has become the standard fingerprint for projectile identification (Table 2). The amount of projectile material incorporated in impact melt sheets is generally low ($\leq 1\%$). The highest recorded is $\sim 10\%$ at East Clearwater, where the siderophiles are carried in a sulphide phase (Palme *et al.,* 1979).

The reliability of the identification of projectile type at terrestrial craters is variable (Table 2) and, in some cases, the identification is controversial. For example, there is a modest siderophile element anomaly associated with some samples at the Ries, which has been interpreted as indicating an aubritic projectile (Morgan *et al.,* 1979; Horn *et al.,* 1983). However, enigmatic Fe-Ni-Cr metal veinlets occurring in amphibolite target rock below the crater floor have been interpreted by El Goresy and Chao (1977) as indicating a chondritic projectile. As the modest siderophile anomaly occurs in a number of lithologies, including both fallback and fallout suevite (Pernicka *et al.,* 1987), the interpretation of an aubrite projectile is favored (Table 2). In other cases, searches for siderophile anomalies at some impact structures have been largely unsuccessful (Palme *et al.,* 1978). This may be due to a variety of reasons, including: a sampling problem (Palme *et al.,* 1981), the impact of a differentiated projectile with little or no siderophile element enrichment with respect to terrestrial materials, or uncertainties in making corrections for indigenous siderophile elements. The terrestrial environment with its wide variety of potential target rocks, including mafic and ultramafic types, and the potential for post-impact alteration and mobilization of elements (Lambert, 1977), is not ideal for determining projectile type from siderophile data. Nevertheless, the interpretation of the data, which are a direct measure of the composition of large earth-crossing bodies, suggests that a variety of projectile types, including achondrites, chondrites, irons and stony-irons, is recorded in the terrestrial data (Table 2).

IMPACT AND THE GLOBAL ENVIRONMENT

The series of observations linking diagnostic shock metamorphic effects to terrestrial impact structures and the identification of siderophile anomalies at some of these structures has provided the background for the interpretation of the chemical (Alvarez *et al.,* 1980; Kyte *et al.,* 1980; and others) and physical (Bohor *et al.,* 1984; Izett, 1987; and others) evidence of a major impact event at the Cretaceous-Tertiary (K/T) boundary. Despite the evidence, a sector of the geologic community continues to argue against this interpretation (Officer and Drake, 1989; Loper and McCartney, 1988; and others) and the subject remains controversial (Silver and Schultz, 1982; Sharpton and Ward, 1990). While the record in the K/T boundary sediments is consistent with the impact interpretation, it is clear that many of the details are not known at the present time, particularly, with respect to the nature of the killing mechanism and associated mass extinctions (Alvarez *et al.,* 1980; Emiliani *et al.,* 1981; Prinn and Fegley, 1987; Wolbach *et al.,* 1985).

One of the arguments, albeit a weak one, of the opponents of the impact interpretation is that there is no known K/T crater. Without this "smoking gun", they have argued that the physical and chemical evidence is circumstantial and open to various interpretations (Officer, 1990). As a result, considerable effort has been undertaken recently to identify the K/T crater or craters. A number of known structures have been suggested as candidates: Manson (French, 1984), but at 35 km it is too small, and Kara and Ust-Kara (Masaitis and Mashchak, 1982), but their precise ages are the matter of debate (Nazarov *et al.,* 1989b,d; Koeberl *et al.,* 1990). Other large-scale structures have been suggested as being related to the K/T event, *e.g.,* the Amirante basin in the western Indian Ocean (Hartnady, 1986) and a 300 km structure in the Colombian Basin (Hildebrand and Boynton, 1990). Neither of these structures, however, are confirmed impact craters. The most recent candidate is an ~ 180 km diameter feature, known as Chicxulub, on the Yucatan peninsula (Hildebrand and Penfield, 1990). This is a buried structure and some evidence of shock, in the form of planar features in quartz in deposits interior and exterior to the structure, as well as possible impact melt rocks, has been reported (Hildebrand and Penfield,

TABLE 3. Impact craters, ordered by age, used to search for periodicity in cratering record.

Crater	Diameter, km	Age, Ma
Zhamanshin	13.5	0.9 ± 0.1
Bosumtwi	10.5	1.3 ± 0.2
El'gygytgyn	18.3	3.5 ± 0.5
Bigach	7	6 ± 3
Karla	12	10 ± 5
Ries	24	14.8 ± 0.7
Haughton	20.5	21.5 ± 1.2
Popigai	100	35 ± 5
Wanapitei	7.5	37 ± 2
Mistastin	28	38 ± 4
Logoisk	17	40 ± 5
Montagnais	45	50.5 ± 0.8
Ragozinka	9	55 ± 5
Marquez	15	58 ± 2
Kamensk	25	65 ± 2
Manson	35	65.7 ± 1.0
Kara	65	73 ± 3
Ust-Kara	25	73 ± 3
Lappajärvi	17	77.3 ± 0.4
Boltysh	25	88 ± 3
Steen River	25	95 ± 7
Dellen	15	109.6 ± 1.0
Carswell	39	115 ± 10
Mien	9	121.0 ± 2.3
Gosses Bluff	22	142.5 ± 0.5
Rochechouart	23	186 ± 8
Red Wing	9	200 ± 5
Manicouagan	100	212 ± 2
Puchezh-Katunki	80	220 ± 10

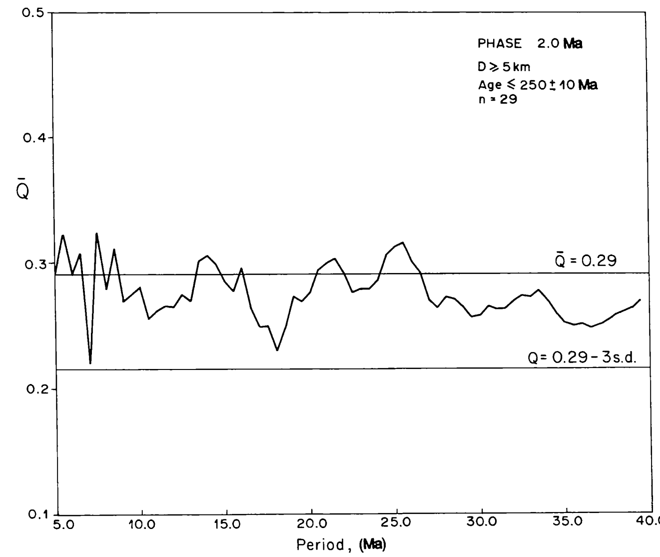

FIG. 11. Plot of $\bar{Q}$ against period of craters with D ≥ 5 km and ages ≤250 ± 10 Ma. The variable $\bar{Q}$ has a mean of 0.29 for random data and a value less than 0.29–3 s.d. is taken to indicate a statistically significant period (see Grieve *et al.*, 1985, 1987). For 0.5 Ma periods between 5 Ma and 40 Ma and phases of 0 to 20 Ma, no period was detected. This figure illustrates the lowest value of $\bar{Q}$, at 7.0 Ma, detected in all runs.

1990; Kring *et al.*, 1991). As there is some evidence from variations in the concentration and size of shocked quartz grains and the thickness of K/T boundary deposits (Bohor *et al.*, 1987; Izett, 1987) that points towards a source in the Americas, the Chicxulub structure may be a promising candidate. The Chicxulub structure, however, also has an element of controversy (Sharpton *et al.*, 1991). Whatever the case, the attention focused on K/T candidates has resulted and, no doubt, will result in important new data compilations and acquisitions at these structures (Hartung and Anderson, 1988; Nazarov *et al.*, 1989a,b,c,d; 1990a,b; 1991).

The potential association of one major impact event with a mass extinction has prompted suggestions that other impacts may have produced similar but reduced effects. For example, the formation of the Austral-asian tektite field has been linked to a climate change (Glass *et al.*, 1979), as has the formation of the N. American tektite field (Ganapathy, 1982), which has also been related to the extinction of several species of radiolaria (Glass and Zwart, 1979). The problem in making these associations with particular craters is generally temporal. For example, Popigai is the correct age (35 ± 5 Ma) for the N. America tektites. Isotopic ages, however, have natural and experimental uncertainties and, therefore, temporal equivalence is never exact. Even if they are equivalent, it can be simply coincidence. What is required is a direct compositional or physical link between the lithologies in the distal impact deposits and the potential source crater.

Age uncertainties are also a factor in assessing periodicities in the cratering record. Following the K/T impact hypothesis, Raup and Sepkoski (1982, 1986) have presented evidence for a periodicity in extinctions in the marine biological record. Others have applied time-series analysis to the terrestrial cratering record and suggest an equivalent periodicity and, thus, a causal relationship (Alvarez and Muller, 1984; Rampino and Stothers, 1984; and others). This has been extended to other aspects of terrestrial geologic history (Rampino and Stothers, 1985). Grieve *et al.* (1985) argued against this type of analysis, principally because of the biases in the terrestrial impact record and the observation that periods of different value can be obtained on different vintages of listings of crater ages. More recently, we have shown that, when actual age uncertainties are taken into account, it is essentially impossible to define any stable periodicity in the terrestrial record (Grieve *et al.*, 1987). Model calculations indicate that the period in a 50:50 mix of periodic and random crater ages can be detected >50% of the time, at the 99% confidence level, only if age uncertainties are <10% of the period being sought. For the ~30 Ma period discussed in the literature, this is unrealistic, given the age uncertainties on terrestrial craters (Table 1), and may never be achieved, given the nature of isotopic age dating techniques. Nevertheless, some authors, despite the statistical arguments concerning uncertainties and the biases in the record, persist in finding merit to the periodicity argument (Napier, 1989). For completeness, the currently known terrestrial record (Table 1) has been again analysed for periodicities. The methodology for time series analysis, for consistency, is that given in Grieve *et al.* (1985) and only structures with D > 5 km and ages <250 Ma with uncertainties of <±10 Ma were considered (Table 3). The results show no obvious periodicity (Fig. 11).

FUTURE DIRECTIONS

Investigations centered around the K/T event and the possible source crater(s) will continue to have a high profile and, thus, focus attention on the terrestrial cratering record. Much of utility of the record will remain, however, with the constrains it provides to understanding impact as a planetary process. It is clear that detailed studies of the character of individual craters are still required. When it is necessary to compile data from the record, it is often found that much of the literature has little depth and much needed information is not present (Grieve *et al.,* 1989; Grieve and Cintala, 1991). In some cases, the literature consists of little more than a discovery paper and various subsequent analytical papers based on samples collected during the initial discovery. Some impact craters, however, have a voluminous literature. These are the historically or presently controversial structures, *e.g.,* Barringer, Sudbury and Vredefort. This is a characteristic of the knowledge base of the record regulated by the manner in which scientific problems are addressed by individual workers and funded by institutions and agencies. Unless more effort is expended on basic characterization, the full potential of the terrestrial record will not be realized. Unfortunately, these types of basic studies cannot compete, in many instances, with high profile "topical" research goals.

The K/T debate has also focused attention on other deficiencies of the terrestrial record. No impact structures are known from the world's ocean basins. This problem can be addressed through the acquisition of better topographic data from the ocean floors. In this respect, high resolution gravity and acoustic data will be of assistance. In addition, much of the discussion on the K/T event focuses on distal ejecta deposits. Almost nothing is known about such deposits. The Acraman structure in Australia, which has distal deposits preserved in the sedimentary record up to distances of 500 km (Wallace *et al.,* 1989), may provide insights. Another candidate is the Montagnais structure on the continental shelf off Nova Scotia (Jansa *et al.,* 1989), which is in an area that has undergone a period of essentially continuous sedimentation since its formation. By the same token, the K/T debate has focused attention on the utility of large scale impact as a stratigraphic marker. The occurrence of tektite layers in ocean sediments (Glass *et al.,* 1979) and the evidence for a medium sized impact in Pliocene sediments in the Antarctic Basin (Kyte *et al.,* 1981, 1988) are examples. The increasing awareness by the larger community of the signature of impact may lead to other discoveries. Some caution will have to be exercised, as not all stratigraphic boundaries with, for example, a geochemical anomaly are necessarily due to impact (McLaren and Goodfellow, 1990), nor are all minerals with anomalous optical properties due to shock (Grieve *et al.,* 1990; Xu *et al.,* 1989).

Terrestrial impact craters are also scientifically useful beyond the information they bring to understanding impact processes. For example, as a fixed topographic form for a given diameter, they can be used as guides to local erosion rates. Also, as closed basins that quickly fill with post-impact sediments, they can provide a record of the paleoclimate and paleontology. For example, the Haughton structure contains a sequence of Miocene sediments not present elsewhere in the N. American Arctic (Hickey *et al.,* 1988). Similarly, sediments in more recent structures have the potential to provide information on global change.

Some recognition is currently being given to the economic potential of impact craters (Masaitis, 1989; Reimold and Dressler, 1990). Historically, it was economic interest that drove D. Barringer to undertake the initial intensive studies at Barringer Crater. Masaitis (1989) estimates that ~20% of the known terrestrial craters have some form of mineralization. The formation of viable economic deposits are due, in general, to the fact that impact craters are unusual terrestrial structures. In detail, however, they can arise from a variety of causes. The 4.5 billion tons of oil shales at the Boltysh (Bass *et al.,* 1967) are due to it being a closed sedimentary basin. The uranium ores at Carswell (Tona *et al.,* 1985) are due to the structural uplift of a Precambrian regolith. The hydrocarbons at Red Wing (Brenan *et al.,* 1975) are due to brecciation and the formation of a structural trap. The Ni-Cu and platinum group ores at Sudbury may be due to liquid immiscibility in a melt of unusually high SiO_2 composition (Naldrett, 1984).

Sudbury will be the site of an international conference on large scale impact and planetary evolution in 1992 and is the subject of a LITHOPROBE transect, which has high resolution reflection seismic investigations as its core. Thus, as much, or more, attention will be focused on Sudbury as was focused recently on the Vredefort structure (Nicholaysen and Reimold, 1990). There is mounting geochemical evidence that the so-called Sudbury Igneous Complex (Pye *et al.,* 1984) has the characteristic of melted crustal rocks (Faggart *et al.,* 1985; Deutsch *et al.,* 1990) and that the Sudbury (Impact) Structure is in the ~200 km range (Lakomy, 1990; Peredery and Morrison, 1984). Least-squares mixing models indicate that the average composition of the Sudbury Igneous Complex can be modelled as a mix of local predominantly Archean granite and greenstone terrain (Grieve *et al.,* 1991). This is consistent with the isotopic data and an origin of the Igneous Complex as an impact melt sheet (Faggart *et al.,* 1985; Grieve *et al.,* 1991; Lakomy, 1990; Stöffler *et al.,* 1989).

This interpretation exemplifies a number of points regarding the present status of our understanding of impact. Previously, the Igneous Complex was considered to be, in large part or totally, magmatic in origin on volumetric arguments. It was too voluminous for the size of the impact crater (French, 1970). We now understand, from a blend of experiment, observation and physics, that larger craters contain more impact melt (Lange and Ahrens, 1979; Cintala and Grieve, 1984; Melosh, 1989) and thus that this volumetric argument is no longer valid. The logic and various observations leading to the impact melt interpretation for the Sudbury Igneous Complex also illustrate the self-consistency and maturity of the knowledge base on impact phenomena. This is also illustrated by the essentially simultaneous, but independent, interpretation of the Igneous Complex as a melt sheet by three of the major groups working on terrestrial cratering problems (Grieve and Robertson, 1988; Masaitis, 1989; Stöffler *et al.,* 1989).

As an impact melt sheet, however, the Sudbury Igneous Complex is unique. It is differentiated. This can be attributed to its thickness, 2–5 km, which is an order of magnitude greater than any other studied impact melt sheet in the terrestrial record. An impact origin for the Igneous Complex has important ramifications for the lunar sample record. As a melt sheet, it presents a suite of relatively coarse grained, clast free differentiated igneous textured rocks. That impact melt rocks from large impact events may present such a sequence has been suggested previously (Grieve, 1980) but the Igneous Complex provides the observational confirmation. It is possible, therefore, that some of the so-called pristine samples from the lunar highlands (Warren and Wasson, 1977) could have the same impact origin. On a personal level, this interpretation closes a loop. It was samples from Sudbury that were my first analogies to lunar breccias and prompted my initial interest in terrestrial impact craters. Now, the reinterpretation of the Igneous Complex as an impact melt may prompt a reexamination of the lunar record. Considerable future work and interpretation are required at Sudbury to glean all the constraints it can provide on the nature of large planetary impact craters and their lithologies. It aptly illustrates, however, the merits of studying the record of impact in the rocks of this planet for what it can tell us about the other planets.

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