ESTIMATES OF IMPACT FLUX FROM TERRESTRIAL CRATER COUNTS: THE ROLE OF GRAVITY AND TARGET PROPERTIES

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ABSTRACT

Data from terrestrial craters is used to derive estimates of the rate of impact crater formation on Earth and illustrate how gravity and target properties influence crater size. The number of craters with diameter >20 km on the North American and northwest European stable cratons is taken as the flux over the last 500 Ma. This is an average rate of 0.15 ± 0.1 x 10 $^{-14}$ km^{-2} year^{-1} or one 20km crater per 1.1Ma for the whole Earth. For a given crater formed in crystalline rocks energy released is calculated from the rate of attenuation of shock waves below the impact point and the dynamic tensile fracture strength of the target materials as confined by overburden pressure at the base of the transient cavity. These results support diameter-energy relationships of the form D = aE^b where b is approximately 1/3.5, between energy and gravity scaling.

1. INTRODUCTION

In 1998 French[1] listed three questions that need to be resolved to determine impact flux on Earth: (1) How often is there an impact of a given size? (2) How much energy is released in a given impact? (3) How large a crater is formed? He also pointed out that "individual estimates of the frequency of impact on Earth for objects of the same size vary by factors of 5-10x, especially for larger objects". The recent detailed study by Bland and Artemieva [2] gives a similar five-fold range for craters with diameter >20km quoting the estimates of Grieve & Shoemaker [3] and of Hughes [4]. While some statistical uncertainty is inevitable, a more precise estimate should be possible by further consideration of Earth's impact crater record in the light of the set of questions raised by French.

Frequency of crater formation depends upon using geological criteria to assess the size and exposure age of the area selected for counting. This gives one measure of crater frequency but questions of energy release and impactor size for a given range of crater diameters must also be addressed. Melosh and Ivanov [5] note that "50 years of study … have not resulted in a predictive, quantitative model of crater formation". To determine energy released and resulting crater size requires an improved understanding of crater mechanics, the effects of target material properties and the role of gravity. This paper addresses cases where there is sufficient information from field and laboratory observations and recent experiments on the dynamic tensile strength of rocks to make such calculations.

2. RATE OF CRATER FORMATION

The Earth has a highly variable surface with a sparse population of impact craters concentrated in a few geologically stable regions. Among them, the exposed Canadian Shield, about 1% of the Earth's total area, is the largest with relatively homogeneous properties. C.S.Beals and colleagues first recognized its potential as a collecting surface after the New Quebec and Brent craters were brought to scientific attention in 1950-51. They realized that these craters are of similar size (3-4km) and form, but differ in age by ~450 Ma or more, suggesting that other impact scars should be preserved. Inspired by the views of Baldwin [6], they initiated the first systematic search over a large area for terrestrial craters that resembled those on the Moon [7].

By 1972 criteria for recognizing impact craters had advanced to the stage where an estimate of the rate of crater formation on the Canadian Shield was possible [8]. The scope of the estimate was enlarged by Grieve and Dence in 1979 to encompass the stable cratonic areas of North America and northwestern Europe [9], a combined area of 17 x 10^{6} km^{2}, or about 10% of the land surface of the Earth. By that date all large (>20km) craters recognized today in those regions had been identified. Their determination of crater production rate showed similar independent rates for the North American and European cratonic regions and a combined rate of 0.35 ± 0.13 x 10 $^{-14}$ km $^{-2}$ year $^{-1}$ for craters with D >20 km. This estimate was based on an accumulation period of 450 Ma. They note that for craters >22.6 km the distribution slope approximates $N \propto D^{-2}$ with some variation for craters >45km where the sample size is small. For craters < 22.6 km the distribution slope is much lower, an effect initially attributed to shorter preservation times and difficulty of recognizing small craters. It is now apparent [2] that breakup in the atmosphere of bodies < 1km in diameter prevents many small craters from being formed.

Three decades of growth in the terrestrial impact database has provided an increase in the number of craters recognized in the North American and Europe but few are within the area originally considered. Those that have been added are too small to affect flux calculations based on the number of craters with D >20 km. Grieve and Dence [9] adopted a -2 size distribution slope and derived an estimate of an average of one impact capable of forming a 20km crater on Earth every 560,000 years. Likewise, for an impact that would be capable of creating a 100km crater on Earth, they project one occurrence on average every 14 Ma. Bland and Artemieva [2] give very similar estimates for this size range in their Table 2 but adopt different distribution slopes of ~ -1.8 for craters <70 km and ~ -2.6 for larger craters. Their curve lies close to that of Hughes [4] and converges at D near 125km. On the other hand, French [1], in his Table 2.1, based on estimates from Grieve and Shoemaker [3] and Neukum and Ivanov [10], calculates one 20km crater per 350,000 years but only one 100km crater per 26 Ma. His figures imply a mean distribution slope of $N \propto D^{-2.4}$.

However, the most robust portion of the distribution curve for all terrestrial craters is in the interval D > 20 < 90 km where the slope approximates -2 [9]. The rate for craters with D > 20 km proposed in [9] gives a projection of 28 craters in the combined North American and NW European cratons, for an accumulation period of 450 Ma. On the other hand only 15 craters with diameter >16km were recognized at the time of the study with the deficiency mainly for D <32km. With no significant change in the numbers taking place since then a more accurate estimate for the rate for crater formation is obtained if the actual number of known craters with D>20km is used. Furthermore, as some craters in Scandinavia and possibly Canada are >450 Ma old [11], the accumulation period for these areas may approximate 500Ma. The rate of accumulation then becomes $0.15 \pm 0.1 \times 10^{-14}$ km^{-2} year^{-1} or one 20km crater per 1.1Ma for the whole Earth. Using a distribution slope of -2, an impact capable of forming a 100km crater would occur on Earth once per 28 Ma. This is in harmony with current knowledge of three craters with D >80km formed on land since the end of the Palaeozoic 250 Ma ago.

3. DERIVING ESTIMATES OF ENERGY RELEASE FROM IMPACT CRATER SIZE

To convert from crater size to impactor size requires an estimate of kinetic energy released on impact. Various approaches have been made in the last half-century with differing results. In the case of Barringer Meteor Crater early estimates of energy released by the iron meteorite range over three orders of magnitude, from 3×10^{14} to 5×10^{17} J [12] and in recent papers still vary by a factor of 5-10x [1,2]. Estimates have been derived by scaling from craters formed by nuclear explosion [13], from observations of the volume of fractured [14] or shock-melted rock [15] or by extrapolation from experiments under controlled conditions. More recently calculations have taken into account a number of parameters. Thus Bland and Artemieva [2] convert from crater size to impactor mass by using the scaling relationships of Schmidt and Housen [16] and selecting for impact at 45°, velocity of 18 km.s^{-1} and densities of 3000 kg.m^{-3} for the target, 3400 kg.m^{-3} for stones and 7800 kg.m^{-3} for irons.

3.1 Stages of crater formation

It is now well recognized that crater formation can be discussed as a three-stage process involving initial contact, excavation of a transient cavity and collapse of the cavity to form the final structure [17]. The size of the fully developed transient cavity provides the most accurate expression of the energy released on impact so a prime aim of the observer in analyzing terrestrial craters is to recover the form and size of the transient cavity. This requires deciphering complications that arise from the processes that produce the final crater. In using an array of observations from selected terrestrial craters it is useful to take each parameter into account according to the stage of crater development that it represents. The present discussion specifically relates to targets comprising strong crystalline rocks of low porosity.

In the earliest stage the target is compressed by shock waves generated on contact and the resulting imprint of shock metamorphism is a direct measure of the reaction of the target materials to the energy deposited [18]. Gravity is not an important factor at this stage but becomes so in two ways as shock waves are reflected from the trailing edge of the impactor and from the free surface to unload and modify the elastic and plastic effects of dynamic compression. Gravity is a control on the volume of melted and fragmented material retained within the crater. In addition, where the target retains strength below the zone of total melting the resulting dynamic tensional regime leads to fracturing and fragmentation, allowing the shocked material to flow freely as the cavity is excavated. At this point, as outlined below, gravity acts through the weight of overburden to regulate the limit of fracturing and fragmentation and hence the depth of the cavity.

In the third stage, the rocks of the uplifted crater rim collapse under gravity, enlarging the rim diameter and either partly filling the cavity with breccia to form a simple crater or by enhancing the upwards motion of the center assists in complex crater formation. Craters occurring in crystalline rocks show a morphological and structural progression with size from simple through flat-floored and complex with central peak to peak ring forms, as are recognized in other planets.

3.2 Information needed for the calculation of energy released on impact

The method employed here for calculating released energy uses the dimensions of the imprint of shock metamorphism as a direct expression of the initial shock compression and its subsequent attenuation. It then considers the extent of fragmentation resulting from the reflected shock waves, particularly as seen directly under the point of impact. Most important is the shock level at the limit of down axis fragmentation at the base of the breccia lens. In simple craters where breccias are preserved this information is obtained by drilling at the center; in complex craters the equivalent fragmentation limit is taken as the maximum shock level at the top of the central peak. Comparative information is needed from laboratory or nuclear explosion experiments along with calculations to derive estimates of the rate of shock pressure decay and expressions of how confining pressure modifies dynamic tensile strength.

In subsequent sections observations from selected impact craters on the Canadian Shield are used as examples of craters formed in crystalline rocks. As discussed in previous papers [19, 20] craters formed in crystalline rocks have the advantage of being formed in relatively homogeneous target materials in which the development and preservation of shock metamorphism in quartz and feldspars is generally well preserved and can be calibrated against laboratory experiments using similar materials. By contrast, data for sedimentary rocks are sparse, more diverse and subject to considerable uncertainty in terms of the effects of variable porosity and contrasts in physical properties across bedding planes.

4. SOURCES OF OBSERVATIONS

4.1 Observations from natural terrestrial craters

An examination of craters formed in the crystalline rocks of the Canadian Shield indicates that the most complete and direct reconstruction of the transient cavity can be made in the case of the Brent crater, the largest known simple crater [19]. Its diameter prior to erosion is estimated as 3.8km. Extensive drilling has provided a detailed cross-section from which the depth to the base of the breccia lens from the original surface is estimated at 1,150m. The Charlevoix crater is taken as the representative complex crater with diameter before erosion estimated at 54km. From the analysis given in [20] the rocks forming its central uplift have risen from below the level of the down axis fragmentation limit. They conform to the general model by moving as large blocks along discrete shear zones rather than as dispersed fragments. At the present level of exposure shock metamorphism at Charlevoix indicates the fragmentation limit in the center was at a shock level of about 25GPa at an original depth of about 11km.

Additional data on shock levels at the limit of fragmentation comes from the craters listed in [20]. They form the basis for the relationship first noted in [19] that the level of shock metamorphism, P (GPa) increases with increasing final crater diameter, D (km) according to the relationship:

$$P = 3.5\ D^{0.5} \qquad (1)$$

By comparing reconstructions of the transient cavity stage at Brent and Charlevoix the striking difference in the size of the excavation relative to the imprint of shock metamorphism is apparent (Fig.1). In large craters substantially more elastic energy is stored below the fragmentation limit and expands during uplift.

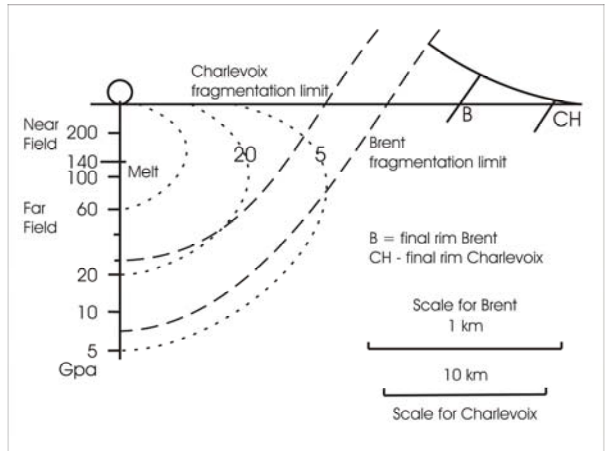

Fig.1 - Comparison of the transient cavities at Brent and Charlevoix normalized to the imprint of shock metamorphism. Note scale difference and that the final rim at Brent (B) is much closer to the transient cavity rim than that of Charlevoix (CH).

In Table 1 the estimated final diameter, transient cavity (TC) diameter and depth from the original surface and shock pressure at the fragmentation limit (FL) is given for the Brent simple crater, Nicholson Lake central peak crater and three peak ring craters.

Table 1 - Craters providing data for energy calculations

Crater	Final diameter (km)	TC diameter (km)	TC depth (km)	Shock pressure at FL (GPa)
Brent	3.8	3.0	1.15	7
Nicholson Lake	14	10	3.5- 4	12.5
Clearwater Lake W	32	18	6.5- 7	20
Charlevoix	54	32	11-13	25
Manicouagan	80	48	17-19	30

The complex craters share the presence of early Paleozoic sedimentary rocks, mainly limestone that formed a relatively thin (<200m) cover over the crystalline rock basement at the time of the respective impacts. These pre-impact sedimentary rocks form the upper part of the transient cavity rim and have first been lifted away from the center and then brought downwards and inwards during late stage collapse. They are thereby preserved in a structural trough around the center. Their inner limit provides a measure of the radius of the transient cavity after correction for movement towards the center. Knowing the radius allows calculation of a range of values for the depth of the transient cavity according to whether the depth/diameter ratio is about 1/2.5 as at Brent or 1/3 as suggested by the Charlevoix restoration (Fig.1).

Some estimates of the final rim diameter used here differ from those of other authors. Rather than the outermost visible circumferential fracture the rim preferred here is the dominant shear zone activated during late stage collapse. It is generally based on structural, gravity anomaly and other geophysical evidence.

4.2 Experimental data bearing on energy release calculations

Two sets of data are of direct relevance to energy calculations. Hugoniot data and shock experiments that produce distinctive shock effects are required for calibration of shock metamorphism. The data used here are based on the measurements summarized in [1,19] and reinterpreted in [20]. In addition Hugoniot data form the basis for the calculations of Ahrens and O'Keefe [21] to determine shock pressure attenuation in crystalline rocks under various conditions of hypervelocity impact. Their results are used in the next section

In addition, Ai and Ahrens [22] have determined shock pressures the onset of fracturing and at the limit where fracturing results in complete fragmentation in two strong crystalline rocks and in Coconino sandstone. Their results place important constraints on the dimensions of transient cavities in similar materials and provide an explanation for the difference in shock levels at the fragmentation limit noted in fig. 1 and Table 1. However, the shock pressures measured experimentally for the onset of dynamic fracturing and complete fragmentation at room temperature and pressure are 100-500 MPa. These are much lower than the shock pressures of 7-30 GPa inferred from shock metamorphism at the limit of brecciation in terrestrial craters.

Ai and Ahrens note that the fracture and fragmentation limits will be affected by confining pressure. It follows that in natural impact events the confining pressure imposed by gravity and the density of the enclosing rocks governs the extent of dynamic fragmentation and hence the depth of the transient cavity. In Fig.2 the results from experiment are compared apparent dynamic tensile strength from observed shock pressures at selected craters. Confining pressures are calculated as the pressure at the base of transient cavities reconstructed as illustrated in Fig.1 with average basement rock density of 2700 kg.m^{-3}.

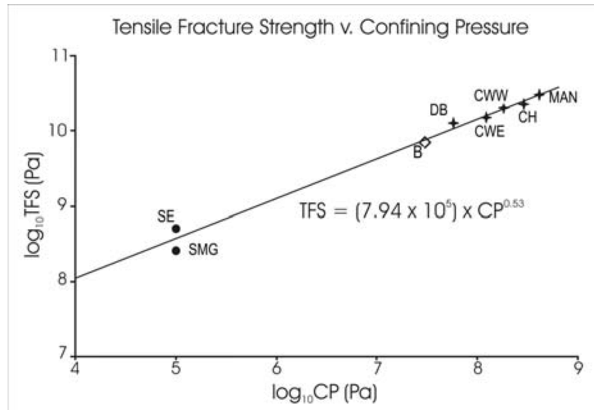

Fig.2 - Dynamic tensile fracture strength v. calculated confining pressure at the fragmentation limit from laboratory data (dots) after [18] and representative Canadian craters: B= Brent, DB = Deep Bay, CWE & CWW = Clearwater East & West, CH = Charlevoix, MAN = Manicouagan

The resulting relationship between confining pressure (CP) and dynamic tensile fragmentation strength (TFS) is remarkably consistent and is given (in Pa) by

$$TFS = 7.94 \times 10^5 \, CP^{0.53} . \qquad (2)$$

4.3 Energy calculations from shock metamorphism in simple impact craters

The calculation of energy release on impact can be made from Ahrens and O'Keefe's [21] calculations of shock-wave attenuation from equations of state. They investigate attenuation in low porosity crystalline rocks for impact by spheroidal iron and stony meteorites striking vertically over a range of velocities. In the calculation that most closely simulates a typical asteroidal impact on Earth, a spheroidal stony body of density about 3900 kg.m^{-3} impacting at 15 km.s^{-1} generates a shock wave with pressure on contact near 300 GPa. Initial decay in the near field is slow as the projectile is embedded but then the calculated rate of attenuation down axis in the far field becomes approximately -2. This result is in good agreement with attenuation rates of shock waves generated in similar rocks by nuclear explosions [19,20]. Since the calculations normalize the centerline distance from the point of impact to the radius of the impactor, R_o(m) the size of the impactor can be calculated from the formula

$$P = 2512 \, (R/R_o)^{-2} \qquad (3)$$

P is the shock pressure in GPa at depth R (m) down axis. Application thus depends on being able to determine the shock pressure at a given depth below an impact point.

This can be most readily done in simple craters if the shock level can be determined at the base of the transient cavity. At Brent the central drill hole penetrates the limit of fragmentation that marks the base of the breccia lens at R = 1,150m where shock metamorphism indicates a mean shock pressure of 7GPa, with an uncertainty of about 3GPa. Then, from Eq.1 R_o = 60.7 m, and for the given conditions [21] the mass of the impactor is 3.65×10^9 kg and the energy released on impact is calculated to be 4.1×10^{17} J.

4.4 Energy calculations for complex craters

Reconstructing the transient cavity in complex craters with central peak or peak ring is more difficult as the axial region is strongly distorted by the late stage central uplift process. However, in the cases noted in Table 1 the diameter of the transient cavity can be estimated from the preserved remnants of the thin layer of pre-impact sedimentary rocks that covered the Precambrian basement at the time of impact. In the case of Charlevoix the distribution of shock zones at the present level of erosion gives a basis for reconstructing the transient cavity [20] and demonstrates that as at Brent it has a depth to diameter ratio of about 1:2.5 to 1:2.8. If this ratio is accepted for the other craters that are listed in Table 1 a depth can be calculated in each case. As the shock pressure at the respective fragmentation limit (FL) is also known from the level of shock metamorphism at the center, Eq. 3 can be applied with results given in Table 2.

Table 2 - Impact energy calculations for complex craters

Crater	Final Diameter (km)	Impactor diameter (km)	Energy released (J)
Nicholson L.	14	0.57	4.13 x E 19
Clearwater Lake West	32	1.28	4.87 x E 20
Charlevoix	54	2.55	3.83 x E 21
Manicouagan	80	4.2	1.7 x E 22

5. GENERALIZATION OF IMPACT ENERGY CALCULATIONS FROM CRATER SIZE

For comparative purposes the diameter to energy relationship is commonly cast in the form $D = a\ E^{\ b}$. Here D is the final diameter, a is a function of target properties, E is the energy released on impact and b varies from 1/3 for energy scaling to 1/4 where gravity dominates. Cooper [23] and others have found that nuclear explosion craters conform to b = 1/3.4. Examples of this relationship include that of Shoemaker [24] whose formulation is based on nuclear explosion data and can be expressed as $D = 1.435 \times 10^{-5}\ E^{\ 1/3.4}$, when D is in km and E in joules. Another example for large craters in crystalline rock also based on nuclear explosion results is $D = 1.96 \times 10^{-5}\ E^{\ 1/3.4}$ [12], while French [1] employs simple energy scaling with results that can be expressed as $D = 2.79 \times 10^{-6}\ E^{\ 1/3}$.

The results in Table 2 along with the result for Brent can be compared to give values for a and b. Taking each pair in turn values for b range from 1/3 to 1/3.9 with an average of 1/3.5; the mean value of a is about 3×10^{-5}. Exponent b is in good agreement with 1/3.4 as obtained from nuclear explosion craters [23] and reinforces indications the importance of gravity in determining the size of the transient cavity and the final diameter of impact craters on Earth. As shown in Table 3, where calculations for D=20km are compared with those in the papers quoted, the method employed here gives energy estimates close to those of other approaches.

Table 3 - Representative energy calculations for craters with final diameter D = 20km

Author	Ref.	Formula	Calculated Energy (J)
Shoemaker	[24]	$D = 1.435 \times 10^{-5}\ E^{\ 1/3.4}$	7.8×10^{20}
Dence et al.	[12]	$D = 2.75 \times 10^{-5}\ E^{\ 1/3.4}$	2.7×10^{20}
French	[1]	$D = 2.79 \times 10^{-6}\ E^{\ 1/3}$	3.7×10^{20}
This paper		$D = 2.87 \times 10^{-5}\ E^{\ 1/3.44}$	1.3×10^{20}
Bland & Artemieva	[2]	$D = 2.16 \times 10^{-4}\ E^{\ 1/3.85}$	1.33×10^{21}

Note that Bland and Artemieva calculate for 45º impacts while all other calculations take the vertical impact case.

6. CONCLUSION

The rate of crater formation adopted here implies that, in the thoroughly explored terrestrial cratons, the terrestrial crater record is essentially complete for craters >20 km over the last 500Ma. This is similar to the position of Hughes [4] and Bland and Artemieva [2] though they extend the record to craters >2-3 km but restrict it to the last 120Ma. A further implication is that the North American and NW European cratons may be slightly over endowed with large (>32km) craters for the area they encompass. Certainly the eastern Canadian Shield is relatively rich in large impacts [8].

It must also be recalled that the database in the two cratons consists largely of craters formed in crystalline rocks. This allows close comparisons with craters on other stony bodies in the Solar system. However, such comparisons must allow for differences in gravity not only in its effect on impact velocity and ejecta distribution but also on the role of confining pressure in determining the depth of transient cavities. In addition, as approximately half the craters on Earth are formed in sequences of sedimentary rocks >1km thick, other complications must be considered in making interplanetary comparisons. The strength of sedimentary materials is generally lower than that of crystalline rock [22] so tensile fracturing and fragmentation will extend to lower shock levels for a given size of impactor. In addition, stratification and porosity of sedimentary materials may have substantial effects. Attenuation of the initial shock wave is greater in porous media [23] and a larger proportion of the energy is partitioned as heat. Likewise the role of water and may be a significant factor, particularly at the late stage of collapse and central uplift. Although a detailed comparison is beyond the scope of this paper, a general statement can be made to the effect that craters formed in sediments are commonly substantially shallower than craters formed in crystalline rock. Crater by crater evaluation is needed for detailed comparisons between the terrestrial crater database and those for other members of the Solar System.

7. REFERENCES

1. French, B.M. 1998. *Traces of catastrophe.* Vol. LPI contribution no. 954. Houston: Lunar and Planetary Institute. 120.
2. Bland, P.A., Artemieva, N.A. 2006. *Meteoritics and Planetary Science*. 41: 607-631.
3. Grieve, R.A.F., Shoemaker, E.M. 1994. In *Hazards Due to Comets and Asteroids*. University of Arizona Press: Tucson.. 417-462.
4. Hughes, D.W. 2000. *Monthly Notices of the Royal Astronomical Society*. 137: 429-437.
5. Melosh, H.J.,Ivanov, B.A. 1999. *Annual Review of Earth and Planetary Sciences*. 27: 385-415.
6. Baldwin, R.B. 1949. *The face of the moon*. Chicago: University of Chicago Press. 239.
7. Beals, C.S., Ferguson, G.M., Landau, A. 1956. *Journal of the Royal Astronomical Society of Canada*. 50: 203-211; 250-261.
8. Dence, M.R. 1972. *The nature and significance of terrestrial impact structures*. in *Proceedings, 24th International Geological Congress*. 1972. Montreal.
9. Grieve, R.A.F. and Dence, M.R. 1979. *Icarus*. 38: 230-242.
10. Neukum, G., Ivanov, B.A. 1994. In *Hazards due to comets and asteroids*, T. Gehrels, Editor. University of Arizona: Tucson.. 359-416.
11. Abels, A., et al. 2002. In *Impacts in Precambrian shields*, J. Plado and L.J. Pesonen, Editors. Springer: Berlin.. 1-58.
12. Dence, M.R., Grieve, R.A.F., and Robertson, P.B. 1977. In *Impact and explosion cratering*, D.J. Roddy, R.O. Pepin, R.B. Merrill, Editors. Pergamon Press: New York. 247-275.
13. Shoemaker, E.M. 1960. In *Report of the XXIst International Geological Congress, Norden*: Copenhagen. 418-434.
14. Innes, M.J.S. 1961. *Journal of Geophysical Research*. 66: 2225-2239.
15. Grieve, R.A.F., Dence, M.R., Robertson, P.B. 1977. In *Impact and explosion cratering*, D.J. Roddy, R.O. Pepin, R.B. Merrill, Editors. Pergamon Press: New York.. 791-814.
16. Schmidt, R.M. and Housen, K.R. 1987. *International Journal of Impact Engineering*. 5: 543-560.
17. Melosh, H.J. 1989, *Impact Cratering: A Geological Process*. Oxford University Press: Oxford. 245.
18. Collins, G.S., Melosh, H.J., Ivanov, B.A. 2004. *Meteoritics & Planetary Science*. 39: 217-231.
19. Dence, M.R. 2002. In *Impacts in Precambrian Shields*. J. Plado, L.J. Pesonen, Editors. Springer. Berlin 59-79.
20. Dence, M.R. 2004. *Meteoritics and Planetary Science*. 39: 267-286.
21. Ahrens, T. and O'Keefe, J. 1977. In *Impact and Explosion Cratering*, D.J. Roddy, R.O. Pepin, R.B. Merrill, Editors. Pergamon Press: New York. 639-656.
22. Ai, H.A. and Ahrens, T.J. 2004. *Meteoritics and Planetary Science*. 39: 233-246.
23. Cooper, H.F.J. 1977. In *Impact and Explosion Cratering*, D.J. Roddy, R.O. Pepin, R.B. Merrill, Editors. Pergamon: New York. 11-44.
24. Shoemaker, E.M. 1977. In *Impact and Explosion Cratering:* D.J. Roddy, R.O. Pepin, R.B. Merrill, Editors. Pergamon Press: New York. 617-628.