ABSTRACTS FOR THE INTERNATIONAL WORKSHOP ON METEORITE IMPACT ON THE EARLY EARTH

Perth, Australia

September 21-22, 1990

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PREFACE

This volume contains abstracts that have been accepted for presentation at the International Workshop on Meteorite Impact on the Early Earth, September 21-22, 1990, in Perth, Western Australia. Responsible for the program are the co-conveners: Bevan French, NASA Headquarters; Richard Grieve, Geological Survey of Canada; and Virgil L. Sharpton, Lunar and Planetary Institute.

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COOLING AND PRECIPITATION OF AN IMPACT INDUCED HOT ATMOSPHERE; Y. ABE, Water Research Institute, Nagoya University, Nagoya, 464-01 JAPAN.

Very large impact heats up near surface materials and a hot transient atmosphere and a transient magma ocean would be formed. In this study, we investigate cooling and precipitation of the transient hot atmosphere.

We calculate the cooling history of the transient hot atmosphere by using the atmosphere model described by Abe and Matsui [1988][1] (For simplicity, we assumed a plane-parallel atmosphere unlike the previous study). Following factors are taken into account, (1) non-ideal behaviors of gases, (2) wavelength-, pressure-, temperature-, and path length dependence of the absorption coefficients, (3) Rayleigh scattering at shorter wave length, and (4) heating due to solar flux. We consider an H_2O rich atmosphere [2] with $H_2O : CO_2 = 9 : 1$ in mole ratio. Results are insensitive to the atmospheric composition, whenever large amount of H_2O is contained in the atmosphere.

Figure 1 shows time variation of the surface temperature. Dotted and hatched region in the Figure indicates existence of a transient magma ocean and a water ocean on the surface, respectively. We neglect cooling stage at higher surface temperature and assume that the initial surface temperature and pressure are 1700K ($\approx$ peridotite liquidus) and 10^6Pa, respectively. We consider two cases corresponding to two different assumptions for the planetary albedo. In one case, we assume the minimum planetary albedo due to the Rayleigh scattering. Obviously this yields the maximum estimate of the net solar flux. In the other case, we assume the maximum albedo (= unity). In this case, no solar flux reach the ground surface. The former case gives the maximum estimate of the cooling time and the latter gives minimum. We assume that the solar flux on the proto-planets is 70% of the present value (960W/m^2 on Earth and 1830W/m^2 on Venus).

Cooling rate decreases with decreasing the surface temperature. It takes several hundred years before starting precipitation. Precipitation rate including circulating water is comparable to or larger than the precipitation rate at the present Earth's equator. Since precipitating water is much warmer than the present, efficient chemical and physical erosion is expected. High precipitation of hot water induced by large impacts are expected to enhance the decrease of CO_2 in the proto-atmosphere and earlier formation of N_2 rich atmosphere[3].

PRECIPITATION FROM A HOT ATMOSPHERE: Y. ABE

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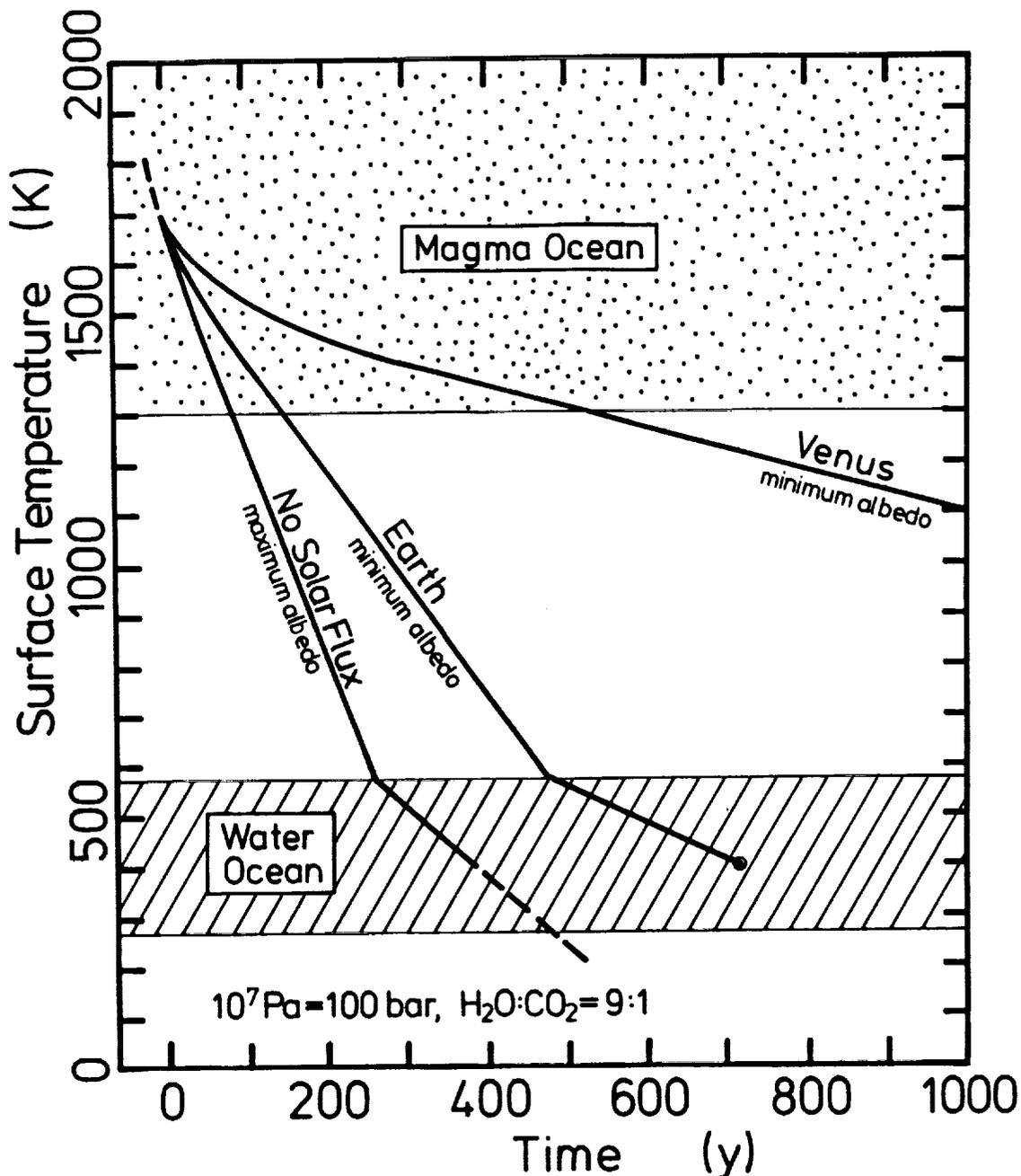
Figure 1.

LARGE IMPACT CRATERS ON THE EARTH, Thomas J. Ahrens and John D. O'Keefe, Seismological Laboratory 252-21, California Institute of Technology, Pasadena, CA 91125.

The effects of impact of large bolides on the atmosphere, ocean, and land as inferred from the results of computer calculations is reviewed. The impact of projectiles on the earth has been studied under a wide range of conditions to establish scaling relations for the maximum depth of penetration, excavation, and vaporization, and the transition diameter from simple to complex crater shapes. We modelled the impact of projectiles on a planetary surface under a wide range of conditions. These include impact velocities (U) of 12 to 38 km/s, impactor radii (a) of 5 to 5×10^6m, gravities (g) of 0 to 10^9 cm/s^2 and strengths of zero to 140 kbar. Planetary deformations and ejecta trajectories are delineated by placing massless tracer particles in planes below the point of impact. We determined the time evolution of crater depth and scaling relationships for the maximum depth of penetration and excavation. The depth of penetration and the crater depth as a function of dimensionless time (Ut/a) grow linearly for dimensionless times <5. For times > 5, the crater grows at a slower but nearly constant rate until it either stops or rebounds. The similitude observed for a wide range of impact parameters agrees with the scaling relationships of Holsapple and Schmidt. The late time motions obey gravity scaling. The transition from simple to complex craters occurs when the gravitational forces due to the size of the transient cavity dominate over material strength. Toroidal flow fields result in rebounding of the crater floor. The transition scales as g^{-1} for rate independent solids, and scales as $g^{-0.8}$ for ice and $g^{-1/3}$ for linearly viscous materials. The transition depth for strength controlled to gravity controlled planetary craters is assumed to occur when the change in the slope of the crater depth versus diameter data of Pike occurs for the terrestrial planets. Using this correlation, average crustal strengths, at kilometer depths of ~5.5 kbar are inferred for the Earth and Mars and ~10 kbar for the Moon and Mercury.

ESTIMATING THE TERRESTRIAL CRATER PRODUCTION RATE DURING THE LATE HEAVY BOMBARDMENT PERIOD. N. G. Barlow, SN21, NASA/Johnson Space Center, Houston, TX 77058.

Estimating the crater production rate at various times and locations in the solar system is important to our understanding of the geologic evolution of planetary surfaces as well as to the origin and evolution of impact populations. Much discussion has ensued regarding the recent cratering rate on the Earth as a result of the theory that mass extinctions of biota are associated with large meteorite or comet impacts. However, the lunar, mercurian, and martian cratering records indicate that a period of intense impact bombardment existed early in solar system history (approximately 4.5-3.8 x 10^9 years ago). The scars of this period, henceforth called the period of late heavy bombardment, have been erased from the terrestrial record through the action of geologic and atmospheric processes. Recent studies exploring the effects of high impact rates on such terrestrial processes as the formation of continents (1, 2), abiogenesis onset (3), and atmospheric erosion (4) have led to increasing interest in this early time period. One of the first problems needing to be addressed as this interest increases is a better determination of the crater production rate on Earth during the period of late heavy bombardment.

The similarity of the crater size-frequency distribution curves on inner solar system bodies indicates that the Earth and Moon have experienced essentially the same bombardment histories. Therefore, an estimate of the terrestrial cratering rate during the late heavy bombardment period can be made by referring to the lunar cratering record. The number of craters observed on the heavily cratered regions of the Moon is 3107 craters in the 16 to 181 km diameter range over 1.232 x 10^7 km^2 and 76 craters between 181 and 1440 km diameter over 3.8 x 10^7 km^2 (the entire surface area of the Moon). Subtracting out the number of post heavy bombardment era impacts as indicated by the mare record and scaling to the entire surface area of the Moon suggests that approximately 9323 craters formed during the period of late heavy bombardment across the lunar surface. The lunar crater density from the period of late heavy bombardment is thus (9323)/(3.8 x 10^7) or 2.453 x 10^{-4} craters $\geq$16 km diameter per km^2.

We can estimate the lunar crater production rate during late heavy bombardment by assuming that the flux rate was approximately constant throughout this period. Although variations in the impact flux rate are proposed for post heavy bombardment time (5) and likely existed during the heavy bombardment period, the lack of information regarding any periodicity of the flux rate during the early part of solar system history forces us to assume a constant flux rate here. The other piece of information needed to compute the crater production rate is the length of time spanned by the period of late heavy bombardment. Age information derived from the lunar samples and the lunar cratering record suggest that much of the highlands began retaining the cratering record about 4.2 x 10^9 years ago (6, 7). The estimates for the termination of heavy bombardment vary from 3.8 to 3.4 x 10^9 years ago, resulting in periods of 0.4 x 10^9 to 0.8 x 10^9 years for the duration of the late heavy bombardment period on the Moon. The corresponding crater production rates are 6.13 x 10^{-13} and 3.07 x 10^{-13} craters $\geq$16 km km^{-2} yr^{-1}, respectively.

The larger size and mass of the Earth compared to the Moon causes meteorites to have a larger impact velocity on the Earth, resulting in larger crater diameters for terrestrial craters than lunar craters given the same

sized object. However, the larger surface gravity of the Earth causes craters to be smaller than lunar craters formed by similar sized meteorites. The combined effects of impact velocity and surface gravity cause meteorites to form craters on the Earth approximately 1.1-1.2 times as large as lunar craters (1, 8). Thus, an object creating a 16 km diameter crater on the Moon would form an approximately 18 km diameter crater on Earth. In addition, the larger gravitational cross section of the Earth causes approximately 1.3-1.5 times as many craters per unit area on the Earth than on the Moon (1). Applying these adjustments to the above derived lunar crater density values results in a terrestrial crater density during the late heavy bombardment period between 3.2×10^{-4} and 3.7×10^{-4} craters ≥ 18 km km^{-2}. The corresponding range in crater production rates computed for both 0.8×10^9 and 0.4×10^9 years duration between 4×10^{-13} and 9×10^{-13} craters ≥ 18 km km^{-2} yr^{-1}.

The estimates of the present day (i.e., post heavy bombardment era) terrestrial crater production rate range from 1.8×10^{-15} craters ≥ 22 km km^{-2} yr^{-1} (9) to 3.6×10^{-15} craters ≥ 20 km km^{-2} yr^{-1} (10). The rates calculated above for the late heavy bombardment period are approximately 111-500 times the present rate. This is likely a lower limit to the late heavy bombardment period crater production rate on the Earth for two reasons: (a) a number of lunar craters from this period have probably been destroyed by subsequent geologic activity, and (b) the Earth may have been subjected to collisions with larger bodies than any recorded on the Moon.

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THE LATE HEAVY BOMBARDMENT CRATER SIZE-FREQUENCY DISTRIBUTION FUNCTION IN THE INNER SOLAR SYSTEM.   N. G. Barlow, SN21, NASA/Johnson Space Center, Houston, ⟨ 77058.

One benefit of spacecraft exploration of the planets has been to provide information about the cratering history during the first billion (10^9) years of solar system existence, a period for which the terrestrial record is essentially non-existent because of the Earth's active geologic and meteorologic environment. Within the inner solar system, the Moon, Mercury, and Mars exhibit surface regions which retain impact scars from the period of late heavy bombardment, a period extending roughly between 4.5-3.8 Gyr and consisting of a time when impact rates where substantially higher than at present. Younger geologic units are found on the Moon, Mars, Earth, and Venus. The different cratering records retained on each of these bodies are analyzed using well-established crater size-frequency distribution techniques.

Two techniques for the distribution of crater size-frequency distribution information are recommended by the Crater Analysis Techniques Working Group (1). The most common of these two techniques is the cumulative crater size-frequency distribution technique. Use of this technique indicates that craters superposed on the lightly cratered lunar mare and martian northern plains follow a power law function with a slope of approximately -1.8 or -2. Cumulative plots of heavily cratered regions also suggest that these regions can be approximated by this same power law function except at the small crater diameter end (Figure 1; P—lightly cratered regions; H—heavily cratered areas). These observations have lead to the general statement that the crater size-frequency distribution of all cratered surfaces approximates a power function with a -2 slope.

A disadvantage of the cumulative plotting technique is the leveling of frequency variations within a particular size range, caused by the cumulative nature of the technique itself. The relative crater size-frequency distribution technique avoids this problem by only determining the frequency of craters within each specified size range. In addition, the relative plot is a differential plot normalized to a -3 differential (equivalent to a -2 cumulative) slope; thus any variations from the power law function with -2 cumulative slope will be readily apparent on a relative plot.

Figure 2 displays the relative plots for the heavily cratered regions of the Moon, Mars, and Mercury (H) and for the lightly cratered lunar mare and martian northern plains (P). The lunar mare and martian plains approximate a horizontal line, which indicates that the cratering record on these regions does follow the -3 differential or -2 cumulative slope of a power law function. However, the heavily cratered regions do not follow a single sloped distribution function at all crater diameters. Analysis of the lunar curve indicates that craters between 8 and 45 km diameter approximate a curve with -2 differential (-1 cumulative) slope, between 45 and 90 km diameter the curve is close to a -3 differential (-2 cumulative) slope, between 90 and 362 km the curve exhibits a -4 differential (-3 cumulative) slope, and above 362 km diameter the curve fluctuates through a variety of slopes. Thus a single slope power law function cannot be used to accurately describe the crater size-frequency distribution function of heavily cratered regions in the inner solar system.

Controversy still exists regarding the cause of the multi-sloped distribution function exhibited by heavily cratered surfaces in the inner solar system (2). However, a number of arguments now exist which support the theory of production populations, i.e., the crater size-frequency distribution

CRATER S/F DISTRIBUTION FUNCTION: Barlow N.G.

theory of production populations, i.e., the crater size-frequency distribution curves accurately reflect the size-frequency distribution of the impacting objects (3). According to this theory, the multi-sloped curves seen in heavily cratered regions of the Moon, Mercury, and Mars reflect the size-frequency distribution of impactors dominating early in solar system history, during the period of late heavy bombardment. The flatter curve exhibited by craters on the lunar mare and martian northern plains has been formed by objects (probably asteroids and comets) which have dominated the cratering record since the end of heavy bombardment (about 3.8 Gyr on the Moon).

Terrestrial craters display a size-frequency distribution curve similar to that of the post heavy bombardment population recorded on the lunar mare and martian plains. The heavy bombardment-era cratering record does not exist on the Earth. However, there is no reason to expect that the Earth escaped the period of high impact rates--the record has simply been erased over time by the Earth's active obliterative environment.

The realization that the Earth was exposed to a period of intense bombardment early in its history was made a number of years ago, and several studies have attempted to relate present geologic, biologic, and atmospheric observations to events initiated during this period of high impact rates (4, 5, 6). However, each of these studies have approximated the crater size-frequency distribution function with a power law function of -2 cumulative slope. This results in an overestimate of the number of craters less than about 45 km in diameter from the number actually observed. More importantly for the results of these studies, however, is the effect at the large crater diameter end of the curve. The dramatic fluctuation in slope among craters greater than about 100 km in diameter implies that the number of craters will be overestimated in certain diameter ranges and underestimated in others. The accuracy of studies trying to determine the effects of high impact rates on terrestrial processes can only be improved with the use of a polynomial function which fits the observed heavy bombardment-era crater size-frequency distribution function.

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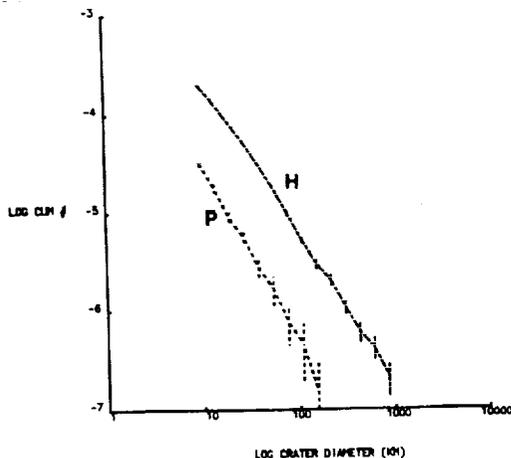
Figure 1
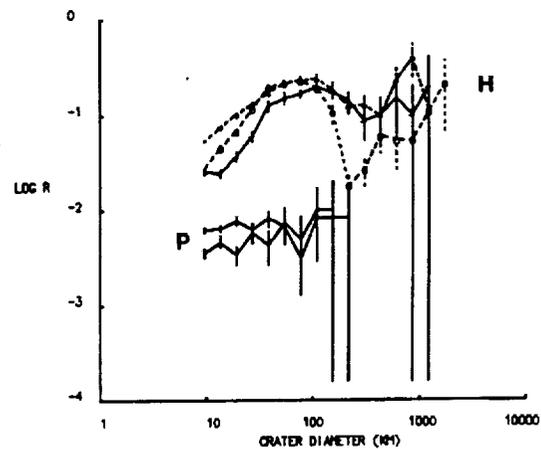
Figure 2

IMPLICATIONS OF THE SIZE DISTRIBUTION OF EARLY EARTH IMPACT FLUX; Clark R. Chapman, Planetary Science Inst., Tucson, AZ 85719 USA

Impact rates in early geologic time depend on the size of impact under consideration. This presentation concentrates on the issues connected with the *size distribution* of the impactors, and how that varied with time, independent of how the overall rate varied in time. Second-order effects of scaling for impact velocity and so on are ignored here.

Evidence on the shape of the size distribution prior to the late heavy bombardment (LHB) is indirect, at best. There is good evidence, however, that the size distribution responsible for the cratered terrains of the Moon, Mars, and Mercury was characterized by a very shallow "slope" (of the power-law representation) for craters of several km diameter up to nearly 100 km diameter. A somewhat steeper slope characterized the LHB at diameters of a couple hundred km, but the overall slope from 1 to 2000 km is nevertheless shallower than the -3 equal-area reference curve. It is believed that the same population of impactors struck the Earth.

A post-mare size distribution is recorded on all moderately to lightly cratered lunar maria beginning shortly after the Orientale event, also reflected on other terrestrial planets. This flux continues today on the Earth as well, and is believed to be caused by asteroids and comets. It is characterized by an appreciably steeper slope (about -3) than the LHB over the size range of several km to 100 km diameter.

Both size distributions, but especially the LHB one, have a trait characteristic of size distributions that are appreciably shallower than -4: the volumetric effects of impacts of the biggest magnitude dominate the cumulative effects of all smaller impacts. In other words, the size distribution is inherently "catastrophic," as discussed by Chapman and Morrison (1). Therefore, early impact cratering of the Earth should be considered not so much as a continuous process but rather as a process punctuated by rare, enormous individual events, of which the very largest one in some relevant time-interval has by far the most profound effects. (The most profound early example is the hypothesized event that led to the formation of the Moon; but analogous domination by big events later in the early history of the Earth is to be expected.) These size distributions appear to be the inevitable result of (a) accretionary processes involving planetesimals and (b) collisional fragmentation processes. But until a definitive explanation for the LHB size distribution is accepted, the specific cause of the catastrophic early history of the Earth will not be known.

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EARLY MILESTONES IN THE EVOLUTION OF THE TERRESTRIAL CRUST

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ABSTRACT

The most ancient terrestrial material identified to date consists of detrital zircon grains from the Mt Narrier Quartzite, the oldest of which yields an ion probe U-Pb age of 4275 Ma [1]. Records of the oldest intact rocks found to date indicate that already as early as 4.0 Ga ago the crust was tectonically diversified and at least locally geochemically evolved. Thus, the oldest units include banded tonalite / amphibolite gneiss and porphyritic granite gneiss (Acasta Gneiss, U-Pb zircon age 3962+/-3 Ma [2]); tonalitic banded orthogneiss (Enderby Land, Antarctica, U-Pb zircon age 3870+/- 10 Ma [3]); anorthositic to gabbroic enclaves in gneiss (Manfred Complex, northwestern Yilgarn Block, U-Pb zircon age 3730+/-6 Ma [4]); monzogranite gneiss (Meeberrie gneiss, Northwestern Yilgarn Block, U-Pb ages of 3678+/-6 Ma [4]) and 3.9-3.8 Ga old tonalitic gneiss-greenstones (metamorphosed volcanic and sedimentary supracrustals) systems in southwest Greenland and east Labrador [5]. The early Archaean gneisses invariably contain enclaves of older basic rocks, namely volcanics and/or layered basic to ultrabasic intrusions. Isotopic and geochemical evidence identifies the bulk of the oldest felsic igneous rocks as the products of partial melting of basic rocks or fractionation of basic magmas [6,7], as evidenced by (A) the Na-rich tonalitic/trondhjemitic composition of the gneisses. More fractionated K-rich early Archaean granites also occur [8,9] but are comparatively minor; (B) generally low initial Sr^{87}/Sr^{86} ratios, which place limits on the age of derivation from mantle-type source; (C) positive initial Nd^{143}/Nd^{144} ratios, which indicate light REE-depleted mantle-type source, and (D) non-radiogenic initial Pb compositions. When combined these observations constitutes compelling evidence for progressive nucleation of sial from basic source materials with mantle-type isotopic signatures. Although Arohaean (pre-2.6 Ga) crustal relics occupy less than 10 percent of the Earth surface, reconstructions of the vertical and lateral crustal structure allowed by studies of the relationships between deep-seated high-grade metamorphic suites and supracrustal greenstone domains provide spatial and temporal constraints on models of early crustal evolution. The data indicate that early Archaean tectonic regimes included older sialic nuclei, simatic crustal domains, younger continental nuclei and rifted domains where clastic sediments accumulated. Vertical superposition and lateral accretion of these domains is indicated by (A) the extensive occurrence of basic enclaves in Archaean gneisses and (B) the occurrence of xenocrystic zircons in Archaean basic volcanic rocks (for

example in Kambalda and Norseman, eastern Yilgarn Block [10]), suggesting that the foundation of younger greenstone belts may include remnants of older granite-greenstone systems and/or derived sediments. Conceivably some greenstone belts evolved as simatic Red Sea-type rifts within older cratons. Detailed studies of Archaean granite-greenstone systems allow the unravelling of their three-dimensional anatomy. Uplifted older high-grade metamorphic blocks, in part consisting of deep crustal root zones of older granite-greenstone systems, are in places juxtaposed with younger rifted and/or accreted volcanic belts, for example the Barberton greenstone belt with the Swaziland gneisses [11] and the Murchison granite-greenstone terrain with the Mt Narrier gneiss complex in the northwestern Yilgarn Block [12]. A continuous transition from an Archaean greenstone belt into amphibolite and granulite facies root zones occurs in the Dharwar craton, southern India, exposing a complete Archaean crustal cross section [13]. Uplifted domal cores of granitic gneiss emplaced into greenstone belts represent plutonic igneous activity contemporaneous with felsic volcanic activity within the volcanic greenstone sequences which they intrude [14]. The uplift of these batholiths was associated with subsidence and rifting of the denser greenstone belts where thick granite-derived clastic sedimentary wedges have subsequently accumulated, for example the Gorge Creek Group in the Pilbara and the Moodies Group in the eastern Transvaal [15]. The intrusion of late post-kinematic K-rich granites into the tonalite-greenstone systems has often taken place along older granite-greenstone boundaries, including roof zones [7]. The intersection and isolation of Archaean cratons by younger Proterozoic mobile belts renders geotectonic interpretations the subject of palaeomagnetic and theoretical considerations. Archaean structural trends display a parallel alignment on Gondwanaland and Pangea reconstructions, an observation interpreted in terms of mantle convection cell patterns [16]. The observed sequential age plots of Proterozoic and Archaean palaeomagnetic poles (from 3.5 Ga in Australia) on single apparent polar wander paths [APWP] which coincide within and between Precambrian shields, as well as between continents [17,18], cannot be interpreted as a mere coincidence, posing strict limits on lateral plate motions during at least parts of the early Precambrian. The overlap of APWP for early Proterozoic age segments in Australia, Africa and Canada has been regarded as a major enigma with potential implications for the question of early Earth radii [19]. The question is reinforced by the scarcity of evidence for ophiolites, flysch and molasse-type lithologies in early to middle Proterozoic supracrustal domains, suggesting that geotectonic processes during much of the Precambrian were distinct from those manifested in Himalayan and Alpine orogenic plate tectonic-related regimes [20]. The marked evolution of geochemical and isotopic signatures from oceanic and island arc-type to continental-type culminating during 2.8-2.5 Ga represents advanced sima-sial transformation. The manifest episodicity of Precambrian thermal history, indicated by strong clustering of isotopic ages with peak basic volcanic events

about 3.5-3.4 Ga and 2.8-2.6 Ga remains little understood. The development toward the late Archaean of a network of mobile belts - the loci of tectonic and thermal activity peaking about 1.9-1.7 Ga, has been interpreted in terms of changes in mantle convection cell patterns [16]. Where major enigmas arise, so does the promise of new discoveries. The recent evidence for major extra-terrestrial impacts in the early Archaean [21] requires thorough re-appraisal of the Archaean crustal record [22]. An attempted summary of the above concepts on early terrestrial evolution is portrayed in Figs 1 and 2.

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Figure 1 - Significant events in the Archaean evolution in Australian, south African and Indian cratons.

Figure 2 - Stages in the evolution of the Precambrian crust.
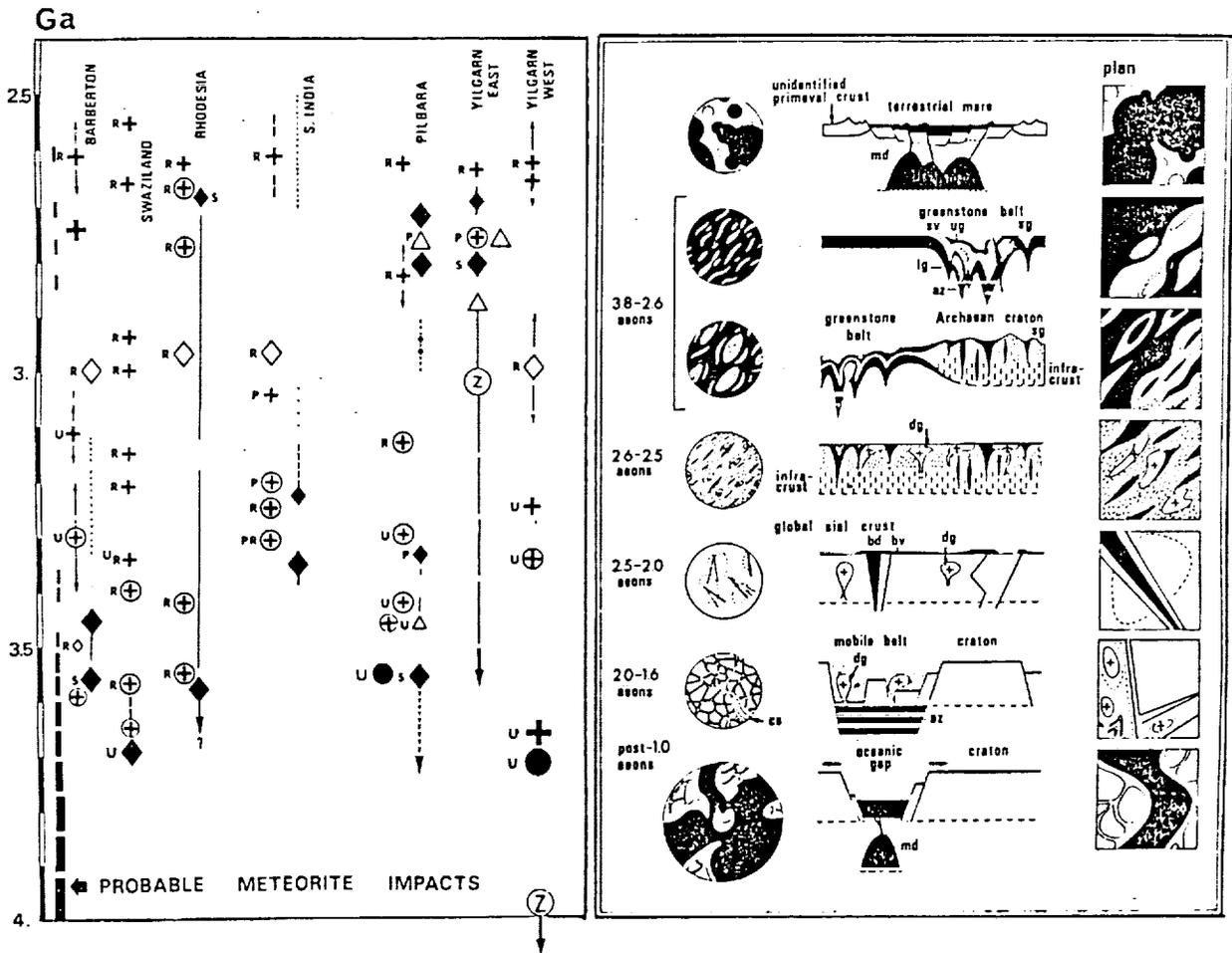
legend to Fig. 1: ●- layered intrusions (occurring as enclaves in gneiss; ◆ - basic/ultrabasic volcanic episodes; △ - felsic volcanic episodes; ⊕- felsic plutonism dominated by Na-rich gneisses; ✚ - felsic plutonism dominated by K-rich granites and gneisses; Z - detrital or xenocrystic zircons; ◇ - thermal resetting episodes; U - U-Pb zircon age; S - Sm-Nd isochron age; R - Rb-Sr isochron age; P - Pb-Pb age.

legend to Fig. 2: md - mantle diapir; lg - lower greenstones; ug - upper greenstones; az - anatectic zone; av - acid volcanics; sg - sodic granites; dg - differentiated granites; bd- basic dyke; bv - basic volcanics; cs - crustal suture

TESTS OF THE ARCHAEAN GREENSTONE BELT-TERRESTRIAL MARIA MODEL

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ABSTRACT

Green [1] pointed out the catastrophic nature of mantle fusion required to produce peridotitic komatiite magmas, suggesting that these events may have been genetically related to extraterrestrial impacts and adiabatic mantle diapirism and melting. Glikson [2] suggested that early Archaean basic/ultrabasic volcanic remnants, for which no basement has been directly identified, may constitute relics of terrestrial maria. These models have remained hypothetical in the absence of direct evidence in the Archaean record of large scale impacts, namely of shock-induced structural and metamorphic impact effects in pre-volcanic sialic terrains laterally juxtaposed with the greenstone belts. Had major impacts been responsible for triggering the voluminous basic and ultrabasic volcanic activity, extensive high-velocity impact effects would be expected. Since the original basement which occurred directly beneath greenstone belts is rarely observed, it is conceivable that the extruded volcanics completely inundated and extended beyond impacted crustal perimeters and that subsequent isostatic subsidence of the thick basic/ultrabasic volcanic cover resulted in the near-complete burial of impacted crust. In such a model the predominantly intrusive granitoids associated with greenstone belts, as well as felsic volcanic components of the latter, represent the products of near-contemporaneous fusion of underlying impacted crust. Pertinent to the testing of this model are comparisons with the proposed impact-related lopoliths of the Sudbury [3,4] and Bushveld [5-7] structures, where high velocity impact has been suggested on the basis of shatter cones, brecciation and extrusion of superheated felsites, namely the Onaping Formation and the Rooiberg felsite respectively. The areally extensive nature of the Bushveld Complex [65 000 square km], as contrasted to the apparent absence of impact effects in neighbouring basement terrains of the Kaapvaal shield, represents a situation consistent with interpretations of greenstone belts in terms of impact-triggered volcanism totally inundating impacted crust. In this line of reasoning, the scarcity of impact features in older Archaean gneiss terrains may suggest that the preserved blocks have been originally laterally removed from impacted sites and were subsequently juxtaposed with the greenstone belts through block/plate tectonics. This is consistent with the common occurrence of major faults between Archaean granite-greenstone terrains and high grade gneiss terrains. A view of greenstone depositories as terrestrial maria would be consistent with the episodic peaks of Archaean basic volcanism about 3.5-3.4 Ga and 2.8-2.6

Ga, conceivably genetically related to major meteorite impact. The near-contemporaneity of basic and felsic igneous acitivity represented by the greenstone/granite systems, as indicated by zircon ion probe data [8], can be interpreted in terms of concomitant fusion of the mantle and of impacted crust - the former related to deep fracturing and pressure rebound and the latter to dynamothermal effects of high-velocity impact. Tests of the terrestrial maria model require (A) a re-examination of and a search for possible structural and shock metamorphic effects along the rarely preserved basal unconformities between Archaean greenstone sequences and underlying crust; (B) a search for impact-induced mineralogical and textural features in Archaean clastic sediments, including basement-derived conglomerates, and (C) a search for iridium anomalies in sedimentary units associated with and immediately above greenstone belts. The recent identification of glass spherule condensate horizons displaying iridium anomalies, chondritic Pd/Ir ratios [Fig. 1] and chondritic-type spinels in the 3.4-3.3 Ga old Fig Tree Group (Transvaal) and similar units in the Warrawoona Group (Pilbara Block, Western Australia) by Lowe and his coworkers [9,10] represents a major breakthrough in this respect. Further identification of stratigraphic evidence and isotopic dating of condensate and ejecta units should allow a precise geochronology of the early terrestrial impact record and facilitate correlations between major Archaean tectonic and volcanic events and extraterrestrial impacts. The occurrence of distal impact ejecta within greenstone belts on the one hand, and the lack of observations to date of impact-induced deformation and large scale brecciation on the other hand, may hint that Archaean crustal blocks may represent the intact allochtonous remnants of an early crust otherwise largely obliterated by mega-impact events.

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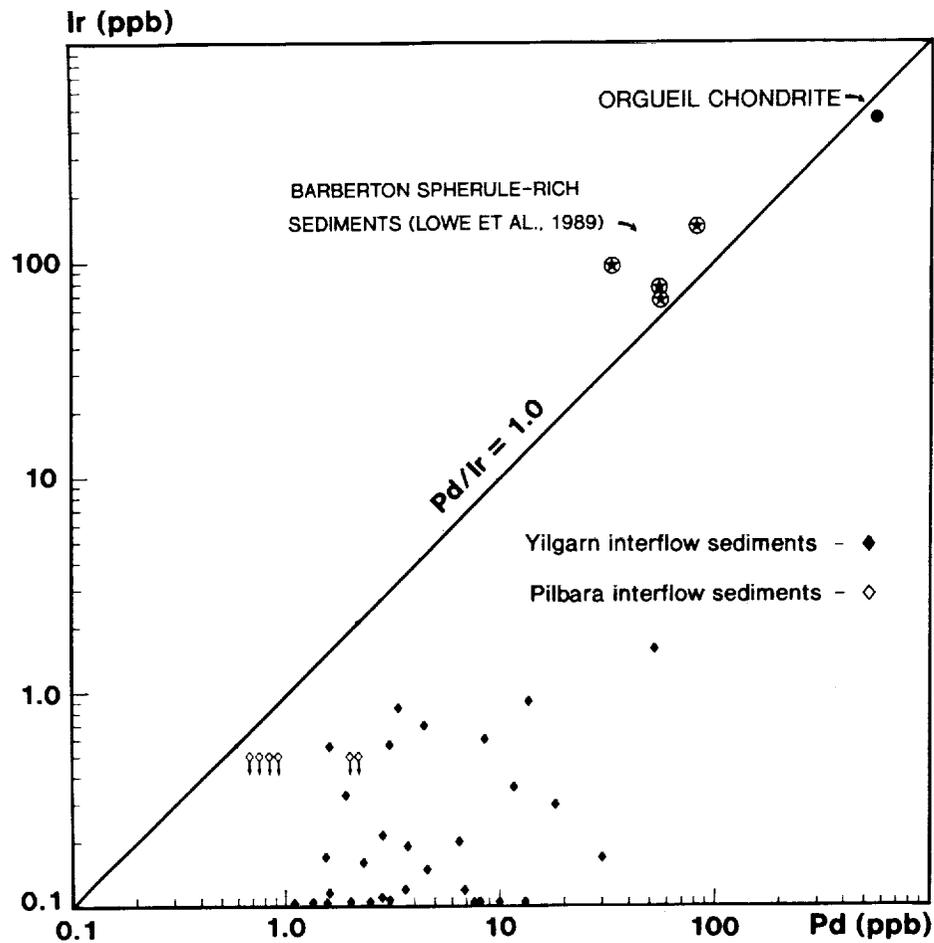
Figure 1 - Pd/Ir plots for Archaean sediments from the
Barberton Mountain Land (eastern Transvaal [10])
interflow sediments from the Kambalda greenstone succession
(eastern Yilgarn Block, Western Australia [11], [◆]) and
interflow sediments from the Pilbara Block, Western Australia
[◇], compared with chondritic values.

SEDIMENTOLOGY AND GEOCHEMISTRY OF THE BUNYEROO IMPACT EJECTA HORIZON, SOUTH AUSTRALIA; V.A. Gostin & M.W. Wallace, Department of Geology & Geophysics, University of Adelaide, GPO Box 498, Adelaide, S.A. 5001, Australia, and R.R. Keays, Department of Geology, University of Melbourne, Parkville, Vic. 3052, Australia.

Discovery of a widespread horizon of shock-deformed volcaniclastic ejecta preserved in late Proterozoic (~600 Ma) shales in South Australia (1-3) and its probable link to the Acraman impact structure in the middle Proterozoic Gawler Range Volcanics (1,4,5) provides a rare opportunity to study the effects of a major terrestrial impact, including the sedimentology and distribution of an ejecta blanket and its precious metal signature.

The ejecta horizon occurs in the Bunyeroo Formation at many localities within the Adelaide Geosyncline (1,2) including the Wearing Hills which are ~350 km NE of the Acraman impact site. Following search at the same stratigraphic level in other basins in South Australia, the ejecta has been located within the Lower Rodda beds of the Officer Basin, extending the limits of the ejecta to ~470 km NW of the Acraman impact structure (3). The ejecta is therefore widely dispersed (Fig.1), and provides an important chronostratigraphic marker enabling precise correlation of Proterozoic sequences in southern Australia.

The ~600 Ma Bunyeroo Formation consists of maroon and green shales, with minor concretionary carbonates,deposited in an outer marine-shelf setting. The ejecta horizon (Fig. 2) comprises mainly angular clasts of acid volcanics ranging from boulder (up to 30 cm diameter) to fine sand size. Its thickness varies from 0 to 40 cm. The basal layers consist of poorly sorted, angular, sand and pebble-sized volcanic fragments set in a mud matrix. These layers were formed by vertical settling of clasts through the water column, followed by subsidence and slumping into the muddy sea bed. The finer grained fragments and clay that have a lower settling velocity usually drape the coarse fragments, and are overlain in turn by lenticular layers of sandstone, usually <10 cm thick, that were formed through reworking by storm waves and current activity. All large fragments and most sand-grade material were derived from a pink to red porphyritic volcanic rock, like that at the Acraman impact site.
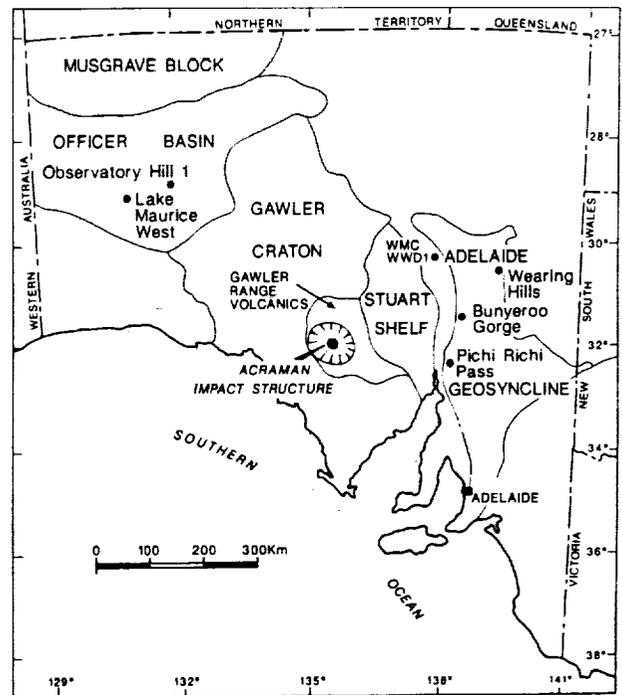
Fig 1 Location of key ejecta localities (solid dots) in Adelaide Geosyncline and Officer Basin, and Acraman impact site.
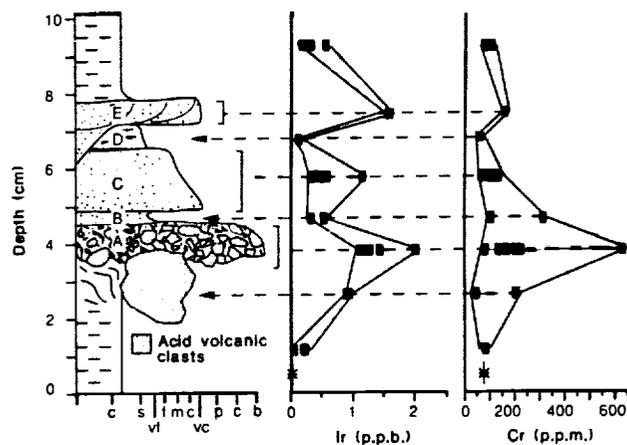
Fig.2 Generalised Bunyeroo ejecta stratigraphy, and Ir and Cr values (filled rectangles) for several localities. A, breccia; B, sandy mudstone; C, sandstone; D, siltstone; E, coarse cross-bedded sandstone. Shaded area = total range in Ir and Cr.

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Evidence supporting an impact origin for the horizon includes: the abundance of shattered mineral grains, the presence of multiple sets of shock lamellae in quartz grains, the presence of small shatter cones on large clasts, the local abundance of altered, tektite-like spherules, and anomalous Ir and other PGE values. The correlation of the Bunyeroo ejecta with the Acraman impact structure is further supported by U-Pb ages obtained from severely shocked, euhedral zircons within the ejecta (6); the dominant age of 1575 ± 11 Ma for the ejecta is consistent with derivation from the Gawler Range Volcanics, which has a U-Pb zircon age of 1592 ± 2 Ma. The geographic distribution of the ejecta (Fig. 1) and the lateral variation of clast size within the horizon also are consistent with the Acraman impact site as the source.

The Bunyeroo ejecta is enveloped in green shales that are several cm thick (1). These shales and the sandy layers of the ejecta horizon are enriched in Cu carbonates, barites and Fe oxides, minerals that are widespread in sediments of the Adelaide Geosyncline. Geochemical profiles of the ejecta horizon (Fig. 2) indicate anomalously high Ir, Au, Pt, Pd, Ru and Cr relative to the host shales of the Bunyeroo Formation (Ir up to 2.0 ppb, Pt up to 270 ppb). Ir enrichment up to 100 times background value for the host shales has been recorded. As iridium values for the volcanic rocks that crop out at the Acraman impact site are <0.005 ppb, the high values for Ir and for other PGEs and Cr in the ejecta horizon strongly suggest derivation from the impactor itself. The marked enrichment in Ir in the Bunyeroo ejecta is similar to that in sediments at the Cretaceous-Tertiary boundary, which has been attributed to a major impact event. The strong evidence for an impact origin of the Bunyeroo ejecta also points to a cosmic source for its PGE signature.

The shales above and below the Bunyeroo ejecta horizon also show Ir and Pt enrichments (0.073-0.45 ppb Ir, 3.1-313 ppb Pt), suggesting post-depositional mobilisation of Ir and Pt. Inter-element ratios of the PGEs within the ejecta horizon from different sites are also quite variable, again suggesting post-depositional, low temperature mobilisation of these elements. Indeed, all green shale horizons in the Bunyeroo Formation which were analysed, regardless of their stratigraphic position, have relatively high levels of Ir and other PGEs. The diagenetic origin of these anomalies is indicated by their association with enrichments in Cu-V-Zn-Co-Ni in thin, permeable green coloured reduced beds in a predominantly red bed sequence. A redox precipitation model similar to that invoked for red bed Cu-U-V deposits has been proposed to explain the PGE anomalies in the green shales (7).

In summary, the Bunyeroo ejecta is unique as the only known example of a widely dispersed, coarse-grained ejecta blanket that is, moreover, strongly linked to a known major impact structure. The marked Ir-PGE anomalies in the ejecta horizon provide support for the hypothesis that meteorite impact events can produce Ir anomalies in terrestrial sediments. The findings also indicate that Ir can be mobilised and concentrated in sediments by low-temperature diagenetic processes. The identification of ejecta horizons in sedimentary rocks therefore should be based on the coincidence of shock-metamorphic features in the detritus and clear iridium anomalies.

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LARGE IMPACT BASINS AND THE EARLY EARTH; R. Grieve, M. Pilkington, Geol. Surv. Canada, Ottawa, CANADA and E. Parmentier, Brown Univ., Providence, R.I., U.S.A.

The oldest terrestrial rocks date at 3.96 b.y. (1) They and 4.0-4.3 b.y. detrital zircons (2) suggest that some form of crust existed during, at least part of, the period of intense bombardment indicated by lunar history. The lack of sizeable areas of identifiable early crust may indicate thermal/mechanic instability due to intense convection and/or bombardment. The effects of bombardment on the earliest terrestrial crust would be, to a first order, similar to those in the lunar highlands, modified by the effects of a relatively higher thermal gradient and the high level of endogenic geologic activity.

The greatest uncertainty in estimating the number of impacts on the early earth is the assumed approach velocities, which affect lunar-terrestrial adjustments for variations in gravitational cross-section. For approach velocities of 6-10 km s^{-1} (3), and generally accepted early lunar cratering rates (4), we estimate that at least ~ 200 impact basins with D > 1000 km could have been formed on the early earth in the period 4.6-3.8 b.y. This estimate may be unrealistic as it requires a large number of residual planetesimals (3), compared to those required to account for the lunar record. Although this cumulative number of impacts add considerable exogenic energy, over 50% results from the largest impact, assuming a standard size-frequency distribution. When averaged over 800 m.y., the additional exogenic energy and impact melt production is of the same order as present-day internal energy losses and island arc volcanism, respectively.

Individual basins would serve to localize enhanced geologic activity. In addition to producing a topographic and structural anomaly, post-shock heating and uplift could result in an enhanced sub-impact thermal regime, sufficient to result in basaltic eruptions due to adiabatic decompression. This leads to the suggestion that large scale impact may have played a role in the formation of proto-oceanic crust (5). Conversely, others have suggested that the effects of such impacts ultimately give rise to differentiated lithologies, i.e. continental type material (6).

We have expanded some previous modelling (7,8), where we consider the convective and conductive heat losses from basin-forming impacts. The initial target conditions were a 80 km thick lithosphere with a 15°C km^{-1} gradient overlying an asthenosphere with a 0.1°C km^{-1} adiabat. This relatively thick and cold starting condition represents an end member for early earth history. The initial thermal anomaly was modelled to include post-shock heating and uplift constrained by observations at large terrestrial complex craters. For basin-forming impacts sufficient to excavate the entire lithospheric column in the center, post-shock heating and detailed modelling of uplift have little effect on the temperature distribution a short time after impact, due to rapid smoothing of the sub-impact isotherms by convection. Accordingly, we simplified the initial temperature distribution to conform to uplift at the center - 0.1 D and no uplift beyond 0.5 D. Cooling was calculated using finite difference solutions for the equations for viscous flow and thermal energy conservation (8).

Once basin size is sufficient to bring the asthenosphere to the surface, basin size has little effect upon the maximum initial temperatures. For a 1000 km-sized basin the initial maximum increase in temperature ΔT ~ 1000°C and is located at the surface in the center of the basin. With time this temperature decreases and is located at greater depth. The main rate controlling step in basin cooling is the Rayleigh Number, Ra, which controls the rate of convective

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overturn. For example, in our model basin, 0.5 ΔT and 0.25 ΔT are reached in ~ 10 and 27.5 m.y., respectively, for Ra - 10^4, compared with ~ 5.0 and 9.0 m.y., respectively, for Ra - 10^7. Ra may have been in the range 10^8-10^6 in the first 800 m.y. of earth history (9), which suggest that convection was turbulent. Initial heat loss is rapid and the lithosphere regains its original thickness in 5 m.y. (Ra = 10^7) to 25 m.y. (Ra - 10^7). Once the heat loss is mainly by conduction, the rate decreases and the time for ΔT to reach zero is on the order of 10^8y, regardless of the initial Rayleigh Number and basin size. The total heat loss from the earth in our model basin is orders of magnitude less than the addition of exogenic energy for the impact event (10^{26}-10^{27} J). Thus, although these large events will serve to speed up internal cooling of the early earth, they still represent a net gain of energy to the system. The relatively rapid (10^7 y) loss of the bulk of the heat from model basins suggests that, if they are to be the sites for the production of highly differential lithologies, then rapid subsidence of basin-fill products is required.

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QUANTITATIVE MODELING OF THE EARLY INTENSE BOMBARDMENT

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Many recent studies have increased awareness of the dominance of impact processes in early crustal evolution, origin and early evolution of the atmosphere and hydrosphere, and early climatic evolution (1-4). However, the magnitude and time-dependence of the intense early bombardment of the Earth are poorly known quantities. We are attempting to refine knowledge of this important function through the application of dynamical models and whatever observational constraints exist. The formation interval as determined through isotopic studies constrains the formation of the Earth to a few tens of My. This allows the derivation of a peak flux which was necessary to accrete the Earth on this timescale. This accretionary flux is estimated to be 2×10^9 times the current terrestrial mass influx, at time $t = -4.5$ Gy (5). Crater counts of dated surfaces on the Moon reveal a flux at $t = -4.0$ Gy of approximately 10^3 times the current influx. The transition between these regimes is being modeled as the consequence of a combination of accretional remnant planetesimals with a range of dynamical lifetimes and collision probabilities. Note that the mean half-life for the decline during the above transition is about 20-30 My.

Combining these results, we investigate the possibility that this dramatically declining flux represents a relatively smooth sweepup of planetesimals left over after Earth's formation. Wetherill (6) studied the dynamical properties of leftover planetesimals in an effort to explain a hypothetical terminal catastrophe, or short, sudden bombardment that was once hypothesized to have occurred 4 Gy ago. Although Wetherill's models were aimed at explaining the supposed cataclysm, one of the models showed the plausibility of a relatively smooth decline with an early half-life of 20-30 My, just as we observe. The analysis suggests that the half-life would slowly increase, as bodies are pumped up into more inclined and eccentric orbits due to close encounters with Earth. This matches the observations.

Based on these consistencies between the observed record and the dynamical analysis of planetesimal calculations, we attempt to reconstruct a plausible impact rate for primordial Earth as a function of time during its first few hundred My.

An entirely different second model for the early bombardment came from early analyses of lunar rocks. The relative absence of rocks older than 4 Gy led some researchers (7) to conclude that a short-lived, sudden cataclysm of intense cratering about 3.8-4.0 Gy ago destroyed earlier rocks. In this view, the cataclysm was much more intense than earlier cratering, and obliterated earlier surface records. Recent presentations of this model (8) stress the lack of impact melts from the pre 4.0 GY period, and even suggest that there was virtually no cratering in the first 800 My.

Unfortunately, many workers discussing early bombardments do not clearly distinguish between these two models. Phrases such as "terminal cataclysm," "late bombardment," and "early bombardment" have been widely used ambiguously. We urge researchers to distinguish between these models, and we also believe that further research on planetesimal dynamics, cratering rates, and lunar rock fragment ages can test between these two models.

We argue that this second model, at least in its extreme form, is inconsistent with the accretion of planets and the subsequent scattering of the leftover planetesimals, which would have sweepup timescales of tens of My.

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Grinspoon (9) affirmed models by Hartung (10), showing that our model of high primordial cratering rate and sharp early decline would lead to destruction of rocks earlier than 4.0 Gy, and preponderance of rocks formed around 4.0 Gy, as observed. This comes from simple models assuming that rock fragment production rate at time T (and/or age resetting) is proportional to cratering rate at time T, while destruction of rocks formed at time T is proportional to total integrated cratering after T.

We note that in our model the half-life of impact flux decline rate starts out short (a few My) and grows as gravitational scattering extends the range of a, e, and i among the planetesimals, in accord with Safronov theory.

Improved quantitative models of the early intense bombardment are crucial in understanding a host of planetary effects, including origin of moons, formation of heterogeneity of initial crusts, formation and retention of initial atmospheres, and possible impact frustration of the origin of life.

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COMET DUST AT THE K/T BOUNDARY:
IMPLICATIONS FOR THE YOUNG EARTH

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Zhao and Bada (1) have recently reported the detection of large amounts of apparently extraterrestrial amino acids in sediments at the Cretaceous/Tertiary (K/T) boundary. These amino acids were found immediately above and below the K/T boundary, but not in the boundary clay itself. Zhao and Bada suggested diffusion from the boundary clay of organics contained in the impactor as the source. Alternatively, we suggest that the amino acids provide a record of the deposition history of debris from the K/T comet swept up by Earth and collected gently and non–destructively as interplanetary dust both before and after the impact (Figure 1). This model agrees with the observed deposition time scale, can be shown to be reasonably probable from cometary orbital statistics, and has good potential to supply the observed quantity of extraterrestrial amino acids. This observation has important implications for the arrival of cometary organics on the young Earth. It is much more efficient to collect some dust from all the comets than just the unvaporized fraction of those few comets that survive impact with their organic cargo intact. Dynamical models which predict massive early comet fluxes in the inner solar system therefore suggest large influxes of unshocked prebiotic organic molecules to the young Earth.

1) Z. Zhao and J. Bada [*Nature* **339**, p463, 1989]

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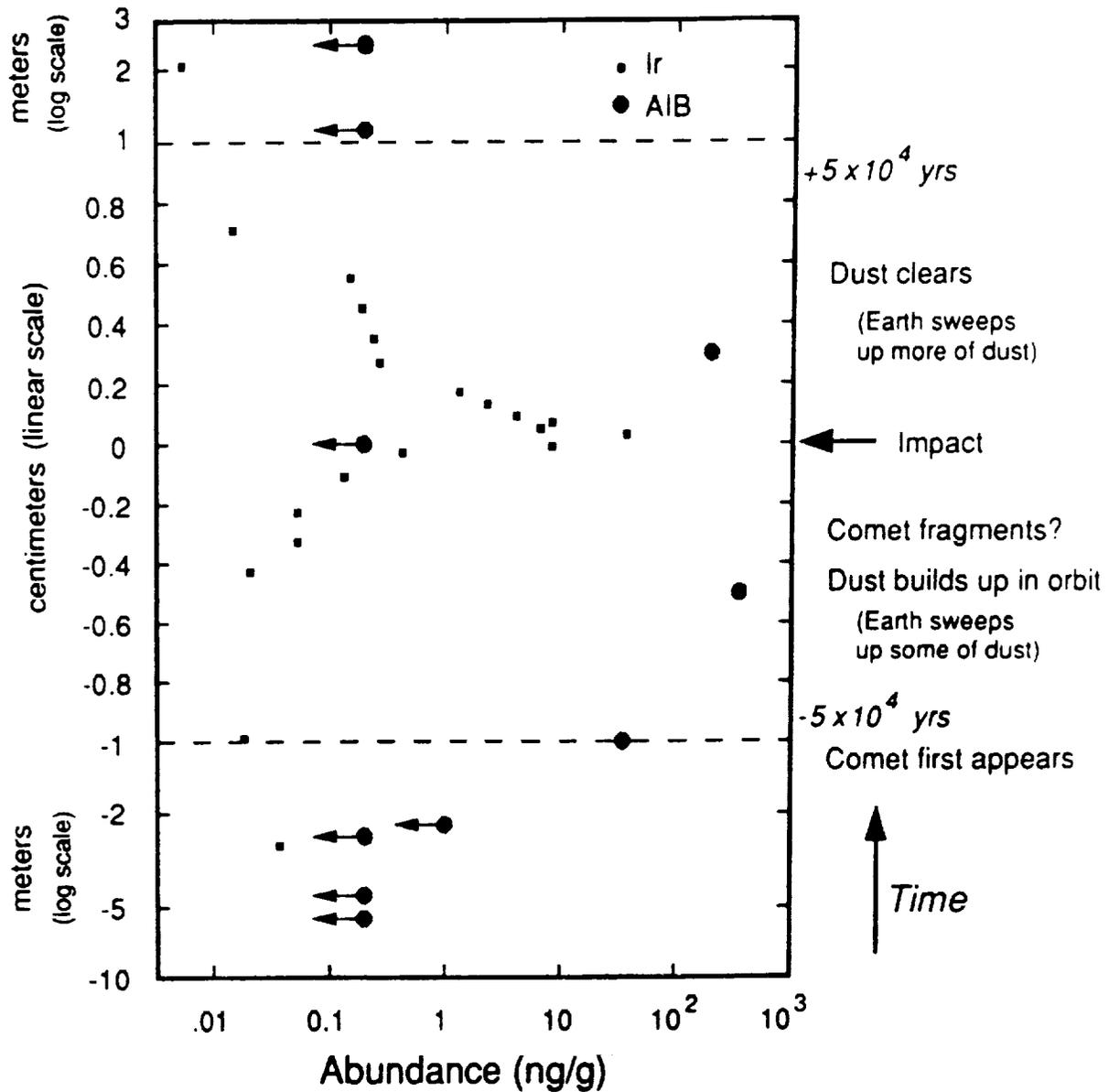
One possible interpretation of abnormal events at the Cretaceous/Tertiary boundary. Iridium [Kastner *et al, Science* 226, 137-143, 1984] and α-aminoisobutyric acid (AIB) [Zhao & Bada, *Nature* 339, 463-465, 1989] abundances near the K/T boundary at Stevns Klint, Denmark, are plotted as a function of depth. The time-scale shown assumes a deposition rate of 1.9 cm/kyr [Preisinger *et al, Nature* 322, 794-799, 1986]; the actual rate could have been faster by a factor of several. AIB is an apparently extraterrestrially-derived amino acid [Zhao & Bada, *op. cit.*]. Arrows denote upper limits. We suggest that the AIB profile is naturally explained by the deposition history of small, organic-rich dust particles evolved from a large comet that evaporated in the inner solar system.

THE NATURE AND CONSEQUENCES OF EARTH'S EARLY INTENSE BOMBARDMENT

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Empirical evidence, even before Apollo, suggested that primordial bombardment rates were much more intense than recent rates. For example, Kuiper proposed this in 1954 (1). More quantitatively, in 1966, I demonstrated from crater counts that the pre-mare cratering rate in the first 1/7 of lunar time _averaged_ some 200 x higher than the post-mare rate--a phenomenon I labeled the "early intense bombardment" or "EIB" (2).

Profound questions persist about the nature of the EIB:

1.	Was the EIB part of a rapid sweep-up of debris following planet accretion ("ACCRETION MODEL") or was it primarily a short-lived catastrophic event about 600 My later ("LATE CATACLYSM MODEL")? In particular, how is the EIB related to the lack of lunar rocks and impact melts older than about 3.9 Gy?

2.	Were the largest bodies in the EIB large enough to cause stochastic catastrophic effects on planets, adding to their diversity? Most importantly, did a giant impact eject Earth mantle material to form our moon, and if so, was this a "normal" part of the EIB?

3.	Did the sources of impactors change with time, and did this cause significant changes in the mean composition or size distribution of impactors over time? Most importantly, what was the role of outer-solar-system bodies?

4.	Did the EIB have important effects on evolution of planetary crusts and environments, in particular, Earth's continental crusts, atmosphere, and origin of life? Are other impact catastrophes, such as the putative K-T boundary event, an extension of this phenomenon?

Some notes on these questions:

1.	At "T=0," 4.55 Gy ago, Earth accreted in an estimated 67 +/- 20 My (3), giving a mean impact mass flux some 10^9 times the present flux (4). Crater counts from the oldest lunar sites indicate that at T = 500 My, 4.0 Gy ago, the bombardment rate was roughly 10^3 times the present (4). Models of planet accretion indicate a gradual sweep-up of interplanetary bodies with gradually lengthening "half-life," which averaged around 20 My (5). These data favor the accretionary EIB model. However, when Apollo and Luna missions revealed virtually no "Genesis rocks" older than 4.0 Gy, Wasserberg and co-workers proposed the late cataclysm model, with a cataclysm about 3.9 Gy ago that destroyed the older rocks (6). I questioned the need for this model (7). **The language distinguishing these two scenarios has since become confused.** Various speakers and writers have used terms such as "early intense bombardment," "late heavy bombardment," and terminal cataclysm" to refer to these events without distinguishing these two radically different models. An intermediate model, with a declining flux curve and stochastic spikes, is also possible. Recently, Ryder (8) defended the late cataclysm model, arguing that there was "only light bombardment in the first 600 My, and then an intense cataclysmic bombardment that produced virtually all of the visible highland landforms." In my view, this is hard to

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reconcile with planetary accretion processes, but Ryder argues that absence of primordial impact melts can't be explained by the accretion model. Alternatively, Hartung calculated that intense early impact processes destroyed the record of early rocks, causing the cutoff of rock ages around 4 Gy (9). Ryder argues that such a calculation cannot explain the lack of impact melts (8). Ryder's work shows the need for a reexamination of the lunar petrologic data in the context of these questions.

2. Hartmann and Davis proposed lunar origin by a giant impact as part of the EIB sweep-up of large planetesimals, and that similar processes caused distinctive features of some other planets (10, 11). Computer simulations support this, but implications and questions about Earth's mantle chemistry point toward further work (11). Confirmation of a primordial giant impact would support the accretionary EIB model.

3. Proto-Earth initially accreted planetesimals of Earth-like composition from its own zone. At some point, accretion of Jupiter's 300 Earth-mass atmosphere allowed Jupiter to perturb planetesimals from its zone throughout the solar system. This may have caused a strong flux of volatile-rich carbonaceous objects that affected other planets (12). In support of a solar-system wide accretionary EIB, Rb-Sr ages of carbonaceous clasts in the Kapoeta meteorite range 4.5 to 3.7 Gy, showing that the period of intense asteroidal collisions and carbonaceous scattering matches that inferred for the EIB (13). Meteorite ages do not show evidence of a late cataclysm (8).

4. Various workers have presented a host of striking, but so far unconfirmed, possible consequences of the EIB. Among them are addition of some atmospheric volatiles, climate modification of early Earth, crustal heterogeneity of early Earth and Mars, and impact frustration of the origin of life. As guaranteed by observed power law size distributions among interplanetary debris, smaller sporadic catastrophic impacts continued through history and may have affected evolution of species. Such effects show the importance of understanding the EIB and subsequent cratering, processes that are dramatically revising our understanding of Earth's history.

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A POSSIBLE ROLE OF HEAVY BOMBARDMENT IN CONTINENTAL GROWTH PATTERN; T. Matsui and K. Kuramoto, Geophysical Institute, Faculty of Science, Univ of Tokyo, Tokyo 113, Japan.

Background   Many geological featurs seem to suggest that tectonic environment have changed at around 3.8 and 2.5 Ga ago (1). For example, no rocks have been found before 3.8 Ga ago, and significant changes in geologic activities such as granitic rock type, tectonic style of metamorphic belts and surface oxidation state are noticed at around 2.5 Ga ago. These changes may be ascribed to change in growth pattern of continents, that is, many micro continents to a few giant continents. The internal process such as vigor of mantle convection has been considered to play a key role in evolution of such tectonic activity. However we propose here that the external process such as huge meteorite impacts play also a significant role in continental growth pattern.

Model   Impact flux is much higher than today's one before 3.0 Ga ago (2). Specifically the highest impact frequency before~ 3.9 Ga ago has been known as the heavy bombardment era. Impact of a giant meteorite on micro continents may cause a disintegration of the continent because formation of a lunar mare-like crater on continent results in change in internal stress field and also initiation of volcanic activity. Then impacts of giant meteorites may play a role in interrupting growth of micro continents through their mutual collisions. We consider that coagulation of continents due to plate motion and disintegration of continents due to giant meteorite impacts compete between 3.9 to 3.0 Ga ago and coagulation becomes dominant with decreasing impact flux after the 3.0 Ga ago. To test this idea we studied the following coagulation equation:

$$\frac{\partial n(s,t)}{\partial t} = -n(s,t) \int_{s_{min}}^{s_{max}} p(s,s')n(s',t)ds'$$
$$+\frac{1}{2}\int_{s_{min}}^{s} p(s,s-s')n(s',t)n(s-s',t)ds'$$
$$-T(s,t) + G(s,t) + D(s,t)$$

where $n(s,t)$ is the number of continent with the surfce area s to $s + ds$, $p(s,s')$ is the collision probability of continents with the surface area s and s', and T, G and D represent the terms of decrease in number of continents due to impacts, generation rate of continents by impacts and increase in number of continents due to disruption of larger continents by impacts, respectively. p is assumed to be proportional to the relative velocity and diameter of continents.

Numerical results   The coagulation equation is numerically solved. When we solved numerically this type of equation, we took into account the numerical effect of choice in size bin interval. Impact flux is assumed to be expressed by an inverse power law with index of -1.8. We considered three models in which we changed the condition for disintegration of continents: model A is no disruption by impact, model B is assumed that continent is disrupted by impact wich crater diameter larger than 100km, and model C is assumed that continent is disrupted by impact with crater diameter larger than 0.1×diameter of continent. Numerical results are shown in Fig. 1 . Evolutions of size distribution of continents are shown in Fig. 1a and temporal variations of total number of continents are shown in Fig. 1b. We can see changes in growth pattern of continents at around 2.5 Ga ago for model C. This suggests that impact flux of meteorites affects the continental growth pattern.

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IMPACT AND CONTINENTAL GROWTH PATTERN: Matsui T. and Kuramoto K.

Fig. 1a
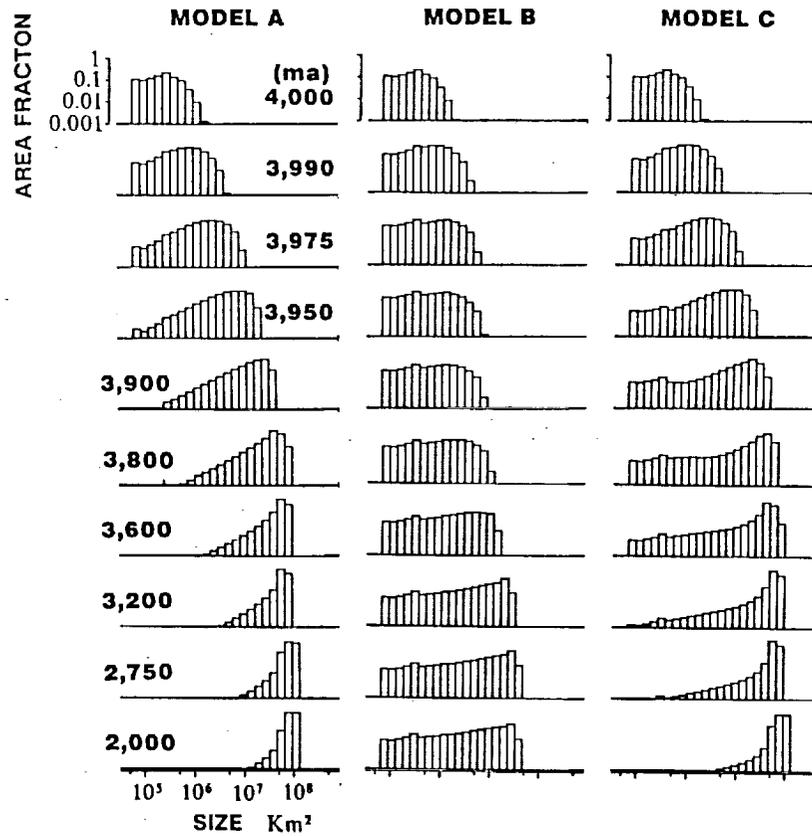
Fig. 1b
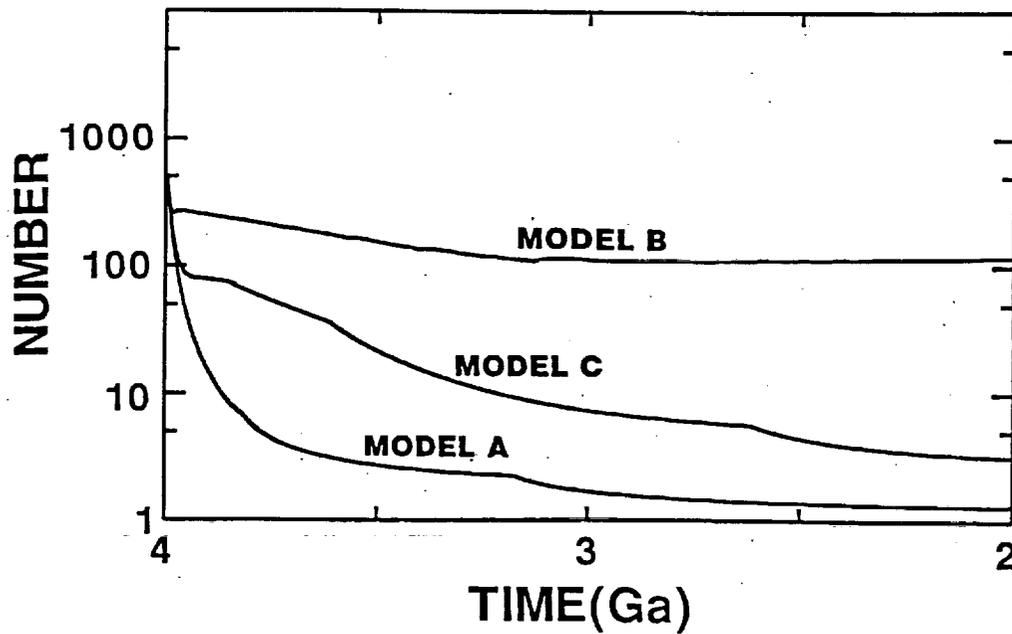
LARGE IMPACTS AND CLIMATIC CATASTROPHES ON THE EARLY EARTH; H. J. Melosh, Lunar and Planetary Laboratory, University of Arizona, Tucson, AZ 85721.

Radiometric dates of cratered lunar surfaces suggest that the cratering rate on the ancient Moon was substantially larger than the present rate before about 3.2 Gyr. A fit to this data suggests that the cumulative flux $N_{cum}(m)$ of impactors of mass m can be adequately represented by the expression $N_{cum}(m) = a[1+Be^{-\lambda(t+4.6)}]m^{-b}$, where a = 1.55 x 10^{-23} $kg^{b}m^{-2}sec^{-1}$, b = 0.47, B = 2300 and λ = 4.53 Gyr^{-1}. Since the cratering rate was higher than present on the Moon, it seems likely that it was similarly higher on the Earth. It is thus gratifying that Lowe and Byerly (1) have recently reported the occurrence of beds of spherules up to 2 m thick in 3.2 to 3.5 Gyr old Archean rocks. These spherule beds closely resemble the 3 mm thick spherule beds associated with the K/T boundary (including elevated iridium abundances), widely believed to have been deposited in association with the impact of a 10 km diameter comet or asteroid.

Until recently it was believed that the spherules at the K/T boundary were transported worldwide as windblown dust. However, it is clear from the 0.1 to 1 mm diameter reported for the bulk of these spherules that their atmospheric residence time is very short, only a few hours, leaving ballistic transport as the only viable means for their global dispersal. We have argued previously (2) that when ballistically transported spherules and other debris reenter the atmosphere approximately 1/6 of their total energy is converted to thermal radiation on the Earth's surface, whereas about 1/3 is deposited in the atmosphere itself, absorbed by water and CO_2. In the case of the K/T impactor the energy irradiating the Earth's surface was about 10 kW/m^2, and the adsorbed energy was capable of raising the average temperature of the lower atmosphere by about 10°C. This amount of thermal radiation is just capable of causing spontaneous ignition of the Cretaceous forests, thus explaining the soot and charcoal found in the boundary clays (3), but the temperature rise of the lower atmosphere is not sufficient to alter its overall stability (the potential temperature difference between the surface and the stratosphere is between 100° and 140°, depending on latitude(4)). On the other hand, the reentering debris would have been relatively efficient at producing NO. Using an estimated efficiency of NO production of one molecule/ 40eV of energy deposited, or 7 x 10^{-9} kg of NO/Joule (5), the ejecta from the K/T impact may have produced 1-3.5 kg of NO/m^2, or NO concentrations of 100-350 ppm. The low Ph caused by such an increase in NO has been suggested (6) as potentially responsible for major oceanic as well as terrestrial extinctions.

The spherule beds in the Archean rocks, however, suggest still greater climatic perturbations. Since thermal energy generation scales directly as the mass deposited, a 10 cm thick spherule bed, if deposited ballistically over the entire Earth, implies thermal irradiation powers of roughly 300 kW/m^2 for periods of time of about an hour (the time scale for deposition is the same for large and small events), temperature rises in the lower atmosphere approaching 300°C, and NO production approaching 1% of the total atmospheric mass (assuming that the ancient Earth's atmosphere was similar in density and structure to the present atmosphere). Surface temperatures on rocks or soil would have approached 1000 to 1700°C, the temperature of the radiating ejecta in the upper atmosphere. It seems unlikely that any life could have survived the thermal pulse on the surface, although oceanic life would have been protected by the evaporation of a few tens of cm of water. The rise in overall atmospheric temperature would have been sufficient to overturn the atmosphere, mixing the suddenly heated troposphere into the stratosphere on a time scale of a few hours. After this sudden event further climatic perturbations may be expected to have continued for some time, perhaps years. Using the equation for the impact cratering flux on the Moon given above, an impact of this magnitude should occur roughly once every 150 Myr on the Earth. The 2 m thick spherule beds imply corresponding greater, although rarer, catastrophes.

The early Earth thus appears to have been a violent and rather inhospitable place: The recent detailed study of the K/T impact has shown that the climatic perturbations of large impacts appear to be more profound than previously estimated. Although ideas similar to this have

been previously suggested (7) for very large impacts, we argue here that even the smaller events recorded by ejecta layers in Archean rocks probably played an important role in shaping the environment of the early Earth, and thus the environment in which life arose.

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ESTIMATION OF IMPACT SITES FROM MAFIC CONTENTS OF PLAGIOCLASE IN METEORITIC CRATER; Y. Miura, Dept. Min. Sci. & Geol., Fac. of Sci., Yamaguchi University, Yamaguchi, 753, Japan.

1. Introduction

Chemical composition (esp. mafic contents of FeO+MgO) and physical properties (esp. shocked lamellar texture and density change) are very important to estimate the degree of shocked metamorphism and source (i.e. host-rock) and distance from the center of the impact site in the meteoritic impact crater and impact fragments of meteorites and lunar rocks [1-8].

2. Mafic contents of terrestrial, meteoritic and lunar plagioclases:

In order to discuss the mafic contents of plagioclase in meteoritic craters, the comparative data of terrestrial, lunar and meteoritic plagioclase (-like) plagioclases are summarized as follows (Fig. 1):

1) In terrestrial volcanic and plutonic plagioclases, mafic contents are less than 0.2 (wt.%), and crystals in volcanic rock have about 3 times the mafic contents (i.e. 0.6 wt.%).

2) In lunar plagioclases, the mafic contents are almost similar to the terrestrial plagioclases, though the brecciated plagioclases of 77515, 67435 and 14066 have about 5 times the mafic contents.

3) Crystalline plagioclases of the Juvinas and Zagami achondrites show ca.0.5(wt.%) of mafic contents which are similar to the mare basalts and lunar anorthosites but about 3 times than the terrestrial basalts and anorthosites.

4) Crystalline plagioclases of type 6 chondrites show various mafic contents of 0.3 (in L6), 1.0 (in LL6) and 1.6 wt.% (in H6).

5) Diaplectic plagioclase(-like) phases of type 3 chondrites show also much more MgO contents of 1.1 (in H3), 1.9 (in L3) and 3.4 wt.%(in LL3).

6) The compositions of the mafic contents are also different between crystalline and diaplectic (cf. maskelynite) plagioclases of meteorites.

7) plagioclase(-like) compositions of Y-691 (EH3) chondrite show three major groups of mafic contents; that is, 1.1 (optically and X-ray crystalline An_{74}), 4.7 (optically crystalline An_{58}) and 2.9 wt.% (glassy An_{38}).

3. Mafic contents of plagioclases in impact crater(Fig. 1):

The mafic contents of plagioclase in impact crater are summarized as follows:

1) In shocked-plagioclases of the Manicouagan impact crater, the mafic contents are similar between crystals and diaplectic plagioclases.

2) The mafic contents among central peaks, intermediate and marginal rocks of the Manicouagan impact crater are clearly different; that is about 20 times the mafic contents in the marginal melt rocks.

Therefore, the mafic contents of plagioclase(-like) phases can be used as indicator of source host-rock, shocked metamorphism of impact sites.

IMPACT SITES FROM MAFIC CONTENTS
Y. Miura

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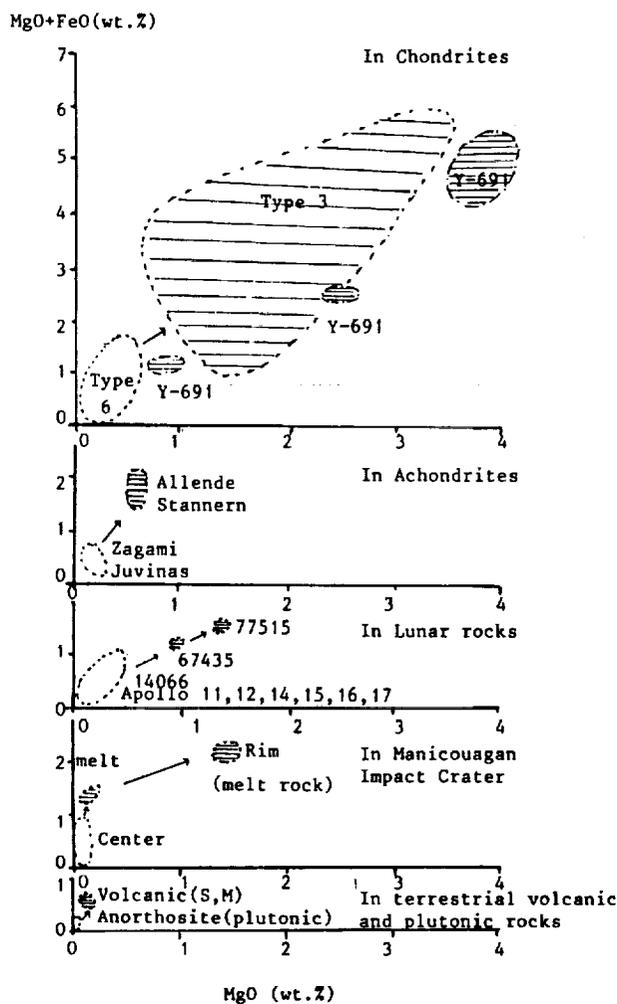
Fig. 1. Relation between mafic content (i.e. MgO+FeO) and MgO (wt.%) in chondritic, achondritic, lunar, Manicouagan impact crater, and terrestrial volcanic and plutonic plagioclases.

DENSITY DEVIATION OF SHOCKED QUARTZ IN THE CRETACEOUS-TERTIARY
BOUNDARY; Y. MIURA and M. IMAI, Dept. of Min. Sci. and Geol., Fac.
of Sci. Yamaguchi University, Yamaguchi, 753 Japan.

Introduction:
    The detailed mineralogical data of quartz silica indicate pressure-
or temperature- (P/T) dependent formation processes of the host rock.
    The purpose of the present paper is to discuss the P/T formation
processes of the Cretaceous-Tertiary (K/T) boundary samples by using
the density-deviation of shocked-quartz minerals [1-3].
Ir content and shocked quartz:
    There are three types of the K/T quartz with and without Ir anomaly:
    1) Type I shocked quartz with abrupt Ir anomaly which is observed
in the sea-sedimental layers of the Italian, Danish, Austrian, Tunisian
K/T boundary samples [3].
    2) Type I shocked quartz without Ir which is observed in the Japanese
Hokkaido K/T boundary sample showing strong tectonic process with
washing out the Ir content [3].
    3) Type II shocked quartz with multiple lamellae with abrupt Ir
anomaly which is observed in the K/T boundary sample from the Clear
Creek North (CCN), Colorado, U.S.A. [1-4].
Density and its deviation:
    Density and its deviation of anomalous K/T peaks with Ir anomaly
in the six K/T boundary samples of the CCN, SK, GI, JKT, ELK and AG
series are listed in Table 1. Almost all K/T boundary quartz crystals
are classified in the high-pressure metamorphic type I (in GI and JKT
series) and the high-pressure type II quartz (in AG, JKT, CCN and SK
series) which indicate the higher density and pressure-formation
processes of the K/T boundary sample as in average data [3].
Formation processes of the CCN K/T boundary samples:
    The CCN K/T boundary quartz samples are considered to be mixed with
the various types of geological and impact processes. Quartz crystal
grains obtained in this study are mixed also as unshocked normal quartz
(type Q, ca. 20%), high-pressure type shocked quartz (type I, ca. 20%),
the Barringer impact crater type shocked quartz with multiple lamellae
(type II, ca. 20%), and tectonic type shocked quartz (type III, ca. 40%)
[3].
    Thus, the probable formation processes of the CCN K-T boundary
samples are summarized as follows:
a)Normal unshocked quartz similar to rock-crystal (Type Q; 20 vol.%).
b)Impact process of the Barringer Crater (Type II quartz; ca.20%),
c)impact and pressure-type metamorphic processes also found in the
    other K/T boundary samples (Type I quartz; ca. 20 vol.%), and
d)impact melt (similar to melt-rock in the Manicouagan Crater) with
    temperature-type metamorphic process (similar to acid-rain quartz;
    Type III quartz; ca. 40 vol.%).

K/T BOUNDARY QUARTZ
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Even in the larger shocked quartz grains of the CCN K/T boundary
sample, the major quartz grains (ca. 80%) have relict data of normal
terrestrial geologic processes (including acid-rain type events), which
have been produced previously the various interpretations of the K/T
boundary problem. In discussion of the K/T boundary problem, it is
more important to identify at first the types (Q, or I to IV) of quartz
samples before final conclusion.

Acknowledgements:

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Table 1. Density and density-deviation of quartz minerals and formation
          condition type (P/T) obtained by calculated density variation.

Sample	Density	$\Delta \rho$ (%)	Formation condition
K-T boundary samples:			
Denmark (SK-2-3)	2.656	+ 0.26	P
Italy (GB-IJ)	2.652	+ 0.06	P
Japan (JKT 3-3)	2.658	+ 0.38	P
Austria(AG3)	2.653	+ 0.15	P
Tunisia(ELK 6)	2.660	+ 0.42	P
Colorado(CCN-SQ2)	2.667	+ 0.68	P
Impact crater samples:			
Barringer (B-3W)	2.664	+ 0.57	P (white sandstone)
Manicouagan (0)	2.646	- 0.11	T (melt rock)
Terrestrial igneous and metamorphic samples:			
Nagato tectonic orthogneiss (393Ma):			
(G-14)	2.655	+ 0.23	P (coarse)
Sangun metamorphic rocks - Crystalline schist (264Ma):			
(S-18)	2.650	+ 0.04	P
Mesozoic volcanic rocks - Rhyolite(92Ma), Rhyolitic tuff(101Ma):			
(KK-15)	2.654	+ 0.19	P (rhyolitic tuff)
Mesozoic plutonic rocks - Granodioririte, Granite (102Ma):			
(A-1)	2.649	+ 0.01	P
Acid rain experiment (172 hours by 15% HF+HCl soln.):			
(AR-5)	2.642	- 0.11	T

ESTIMATION OF SHOCKED PRESSURE FROM DENSITY DEVIATION OF SHOCKED QUARTZ IN IMPACT CRATER AND THE CRETACEOUS-TERTIARY BOUNDARY; Y. MIURA, T. ASHIDA, K. OKAMOTO, Dept. of Min. Sci. and Geol., Fac. of Sci. Yamaguchi Univ., Yamaguchi, 753 Japan;

INTRODUCTION:

Shocked pressure has been estimated from the mineral texture and the size of impact crater [cf.1,2]. But it is desired to make an useful and convenient method of the estimation of the shocked magnitute.

Miura (1990)[3,4] reported that from higher precision cell-parameters and calculated density of quartz crystal, the various quartz crystals hold the structural memory of pressure- and temperature-dependent formation processes. This is mainly because high-pressure silica forms of coesite and stishovite are easily to become too glassy to obtain the the crystal data [3,4].

The purposes of the present study are to investigate shocked pressure of the well-studied impact craters of the Charlevoix [5,6], and (2) to compare the estimated shocked-pressure with the previous impact crater, volcanic tuff, and the Cretaceous-Tertiary (K/T) boundary samples.

ESTIMATED SHOCKED-PRESSURE:

In order to estimate the maximum shocked pressure at the center of the Charlevoix impact crater, quartz sample Nos.11 (MBP-25-67B; 140 kb) and 12 (MBP-35-67; 200 kb) have been considered to be maximum density deviations, as shown in Fig. 1. The other granitic samples show average values of temperature-type negative deviations.

By using this relation, impact craters of the Barringer (B-3W) and the Sudbury are plotted, and six K/T boundary samples of Colorado(CCN), Tunisia(ELK6), Japan (JKT3-3), Denmark(SK2-3) Austria(AG3) and Italy (GB-IJ) are also plotted in Fig. 2 to estimate the proposed shocked-pressure, which are summarized as follows:

1) Land-sediment of the CCN Colorado K/T quartz shows the maximum shocked-pressure of ca. 400 kb of all the K/T boundary samples.

2) The Barringer impact crater holds the maximum shocked-pressure of 300 kb in this study.

3) Volcanic tuff (KK-15) shows also pressure-type quartz in this study which is one third or fourth less than the Barringer Crater or Colorado K/T (CCN), respectively.

4) The estimated maximum shocked-pressure of unknown rock samples from quartz crystals will be discussed by Fig. 2.

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ESTIMATED SHOCKED-PRESSURE
Miura et al.
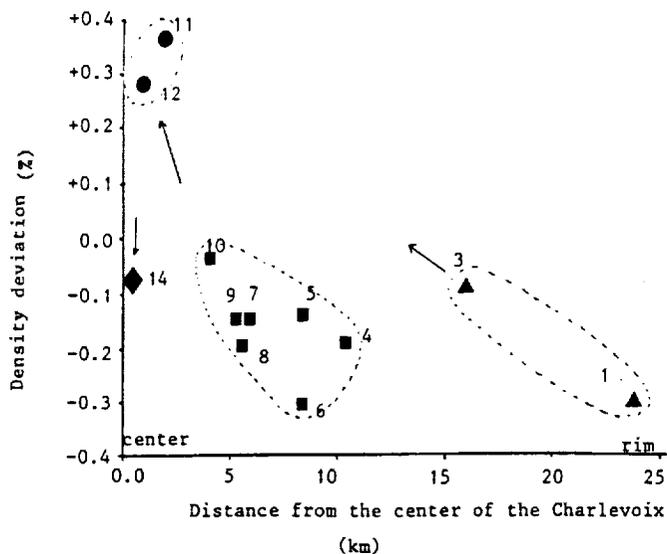
Fig. 1. Plot of density deviation ($\Delta\rho$,%) vs. distance from the center
of the Charlevoix crater[4, 5].
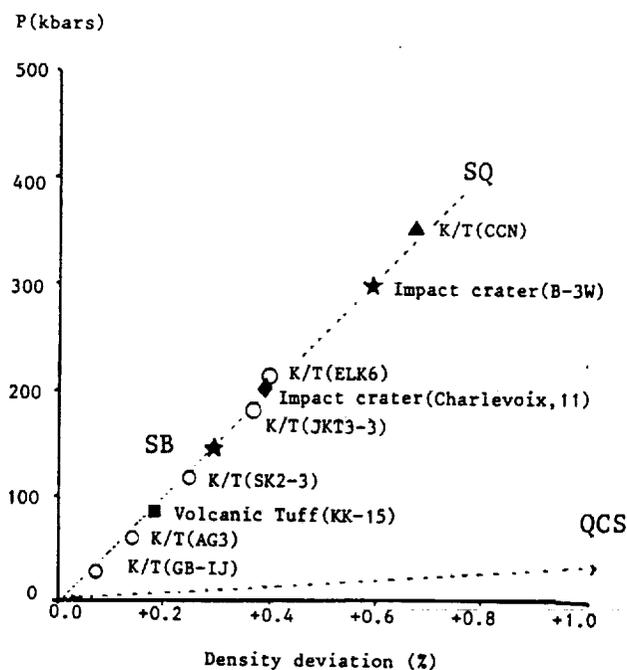
Fig. 2. Plot of estimated shocked-pressure (kb) vs. density deviation.

THE ECONOMIC SIGNIFICANCE OF IMPACT PROCESSES. W.U. Reimold[1] and B.O. Dressler[2]. [1]Economic Geology Research Unit, Dept. of Geology, University of the Witwatersrand, P O Wits 2050, Johannesburg, RSA. [2]Ontario Geological Survey, 77 Grenville Street, Toronto, Ontario, Canada M7A 1W4.

Investigations of terrestrial impact craters and impact cratering processes, and the study of craters on the moons and planets of our solar system appear to be subjects of pure academic research on planetary evolution. They are hardly ever considered as having any economic significance. Mining operations in terrestrial impact structures are rare. There are, however, a few small and medium-sized structures, where explorationists are believed to have viable economic targets. In addition, one of the largest structures of possible impact origin on Earth, the Sudbury Structure, hosts the largest concentration of base metals on Earth.

Amongst the small and medium-sized impact structures on Earth is the Carswell Structure in Saskatchewan. It contains some of the richest uranium deposits on our planet. Their formation is unrelated to the impact itself, but the structural modifications of the deposit and their host rocks due to the impact have significance in terms of exploration. In another impact structure, the Meteor Crater in Arizona, D. Barringer attempted unsuccessfully to mine the Canyon Diablo meteorite. In the Siljan structure in Sweden a deep-drilling project recently tested for hydrocarbons in the igneous target rocks, based on the assumption that the impact-fractured crust at Siljan might have facilitated the intrusion of hydrocarbons from deep crustal or mantle sources. The crater-filling sedimentary rocks of the Nördlinger Ries in Germany were unsuccessfully explored for lignite and the small Roter Kamm crater in Namibia attracted short-lived attention by explorationists looking for diamonds. Several other crater structures are in fact oil producers; titanium was mined in the Charlevoix structure, and the Boltysch crater structure in the USSR hosts some 10^9 tons of oil shale. Much valuable insight into cratering processes has been obtained from exploration and drilling for salt domes and other potentially oil-producing structures in the midwestern USA, Canada, and the USSR. A few of these exploration targets, such as the Marquez Dome in Texas, are now considered to be impact structures.

Two of the largest structures on Earth, considered by many, but not all, researchers to have formed by meteorite impact, are the Sudbury Structure in Ontario and the Vredefort Structure in South Africa. These structures, especially Sudbury, as well as the Simpson Desert depression in Australia, (which has recently entered the scientific debate as a potential impact basin), are of substantial economic importance. Impact-related deposits, such as the lunar regolith on the moons and planets, may prove to become of similar significance. They are presently being considered as sources of commodities required for space station-based planetary exploration.

The Sudbury Structure in Ontario is widely regarded as an impact structure. Earth's largest nickel-copper deposits are associated with it and almost one billion tons of ore have been mined since 1883. The ore deposits occur within the Sublayer at the base of the Sudbury Igneous Complex. Within the impact scenario, the Igneous Complex was either generated by an impact- triggered mechanism from deep crustal or upper mantle sources, or constitutes entirely an impact melt body. Assimilation of silica-rich country rocks by the rocks of the Igneous Complex triggered precipitation of sulphides. These are believed to have accumulated within a still unexplored basal portion of the Igneous Complex, from where, at a later stage, sublayer magma pulses transported the ores into their present position at the lower contact of

ECONOMIC SIGNIFICANCE OF IMPACT REIMOLD, W.U. and DRESSLER, B.O.

the Igneous Complex and into "Offset Dykes". Shock brecciation, shock melting and the post-impact residual temperature in the country rocks facilitated bulk assimilation. Studies of post-Sudbury event deformation of the southern part of the Sudbury Structure as well as further structural studies and testing of the impact model are needed and have high potential regarding the discovery of economic mineralization at presently unexplored depths.

The Vredefort dome in South Africa by itself has not been a very important terrane with regard to available resources. Between 1887 and the 1930's numerous attempts were made to find gold mainly in the conglomerates of the Kimberley formation and with less success of the Johannesburg subgroup in the collar around the granitic core of the structure. Gold exploration also extended to the greenstone terrane in the south-eastern sector of the basement. However, with the exception of several mines in the northwest collar, most of these prospects were quickly abandoned. It was also thought that the positive gravity anomaly in the centre of the dome could be caused by a mafic complex possibly containing base metal deposits. Preliminary assay of mafic rocks from this area did not, however, yield favourable results. Diamonds have been found in gravel terraces associated with the Vaal River beds on the northern part of the dome, but these workings have been terminated as well. Present-day mining activity in the structure is restricted to bentonite workings at the margin of the greenstone occurrence in the southeast and to several dimension stone quarries in the Outer Granite Gneiss. However, the economic importance of the Vredefort structure lies in its setting near the centre of the structurally preserved Witwatersrand Basin - the world's foremost gold and uranium province. The basement of the structure is being studied and compared with other Archean granite-greenstone terranes in the hinterland of the Witwatersrand basin, in an attempt to better define the nature of the source area(s). For the same reason, sedimentary studies relating to the paleo-transport directions for Witwatersrand sediments are being carried out around the dome. There can also be no doubt that a major catastrophic event at Vredefort at ca 2 Ga ago would have seriously affected the structural preservation of the whole basin. With respect to the general understanding of the tectonic evolution of the basin it is vital to continue research into the Vredefort structure in order to be able to separate Vredefort effects (structure, fluidization, etc.) from other earlier tectonic events. One of the problems is whether the enormous volumes of fault rocks (pseudotachylite, mylonite, cataclasite) associated with major faults in the basin are related to the Vredefort deformation phenomena. The enigmatic Vredefort Discontinuity, a mid-crustal zone rich in pseudotachylite and charnockitic rocks, and which displays the highest degree of deformation within the basement (between Outer Granite Gneiss and Inlandsee Leucogranofels), needs to be further investigated and its origin ascertained with respect to the secular evolution of the dome. Recently it has been suggested that this zone could be the extension of a major décollement in the northern portion of the Witwatersrand basin, known as the Black Reef décollement. A parallel discontinuity just to the south of the centre of the structure and equally trending SW-NE was proposed as the extension of the Master Bedding Fault, also tangential to the dome in the northwest. Regional geophysical studies have shown that the Vredefort structure lies on a NW-SE trending basement high, that also forms the axis of symmetry to the Witwatersrand basin. These arguments clearly demonstrate that further study of the Vredefort structure should yield important new insights into the origin and evolution of the structure itself, and the regional effects of the structure on the preservation of the Witwatersrand province.

THE VREDEFORT STRUCTURE - NEW RESULTS, WITH A FOCUS ON STRUCTURAL ASPECTS OF THE VREDEFORT DOME, AND SURROUNDING AREAS OF THE WITWATERSRAND BASIN. W.U. Reimold[1], P. Fletcher[2], C.A.M. Ferreira[3], and W.P. Colliston[4]. [1]EGRU, Univ. of the Witwatersrand, P.O. Wits 2050, Johannesburg, [2]14 Vrede Str., Fochville 2515; [3]Gold Fields of S. Afr., P.O. Box 53, Krugersdorp; [4]Dept. of Geol., Univ. of the OFS, P.O. Box 339, Bloemfontein, 9300, RSA.

For several decades it has been - sometimes quite hotly - debated whether the Vredefort structure near the centre of the Witwatersrand basin represents the deeply eroded remnant of a Proterozoic (2 Ga), 140 km diameter impact structure or was generated by a catastrophic internal gas explosion. In recent years two additional hypotheses have been promoted: (i) that this enigmatic structure could be the result of regional tectonism (e.g. refs. 1-4) and (ii) that the present structure represents an impact structure that was significantly modified by post-2 Ga tectonics (5). It was for the following reasons that a single catastrophic origin has been questioned: a) evidence for multiple pseudotachylite-forming events on the Vredefort dome and in the northern Witwatersrand basin, b) anomalous nature and distribution of microdeformations in quartz from the basement core of the structure, c) large-scale development of major faults and décollements in the Witwatersrand strata, which have been impregnated with massive pseudotachylite and mylonite, d) the asymmmetric structure of the dome, with up- or overturned stratigraphy in the N and NW, but low-angle stratigraphy in the S, e) structural field data that show a lack of pervasive structural deformation in the core that could be linked to a catastrophic event at ca. 2 Ga ago (1,6). All structures observed in basement granite predate the deposition of the Precambrian cover rocks. This implies that any crypto-explosion hypothesis could not possibly be employed to explain the subvertical structures predating the Dominion Group and younger Precambrian rock sequences.

These collective observations led to the development of two working hypotheses, firstly the subhorizontal shear model of (1,6), and secondly the suggestion by (2) that the structure could be the result of interaction between gravity sliding on major bedding-parallel décollements (Master Bedding Fault (MBF) and Black Reef Décollement Zone (BRDZ)) in the northern Witwatersrand basin that were extended onto the dome as thrusts, a compressional component directed northwesterly (Vredefort Axis, cf. (7)), and accelerated uplift at Bushveld times. In addition to these observations we now add new remote sensing and field results.

The interpretation of Landsat MSS over the Witwatersrand basin (8) shows the presence of major linear and circular structures, reflecting the tectonic evolution of the area. Around the Vredefort dome a number of major lineaments cut across the structure, those trending northwesterly being more prominent. A marked lineament trending NE-SW cuts the southern end of the dome, coinciding with the here introduced Mara Décollement Zone (MDZ, Fig. 1). A detailed air photo interpretation has shown a structural discontinuity across this zone, evident by a change of direction of prominent fractures: northerly in the N, and NW in the S. A southwesterly running present-day stream (Rietspruit) essentially follows the trend of the Décollement Zone; this stream also roughly delimits the northernboundary of continuous Karroo sedimentary cover. A number of scattered outcrops

1

P Parys
V Vredefort
I.S. Inlandsee Pan
Exp. Gran. Gneiss
Karroo cover
Exp. West Rand Gr.

N
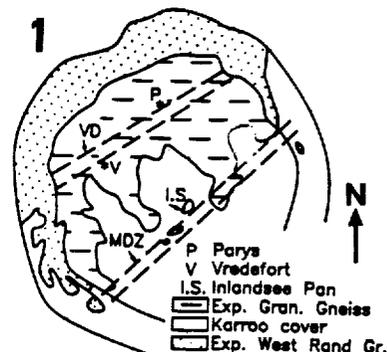
VREDEFORT - STRUCTURAL ASPECTS: Reimold, W.U. et al.

along this proposed décollement have been studied. Major pseudotachylite occurrences are associated with basement outcrop on farms Rietgat 264, Wonderheuvel 173, Rondeheuvel 266, Samaria 484, and Mara 1084. Minor occurrences are found on Bethesda 1087 and Welgerust 172. Many of these veins or network breccias trend parallel to the orientation of the décollement. The basement granite within the zone is strongly sheared and often displays a well-developed foliation dipping at gentle angles (< 15°) towards the N/NE. It appears significant that at most outcrops along this zone mafic intrusives into heavily sheared granite were observed. No outcrops to the E/NE of the Inlandsee along this trend could be studied due to extensive Karroo cover. However, it is noted that on strike the arc of collar strata in the E is abruptly terminated against an extrapolated extension of the MDZ.

Fletcher and Reimold (2) on the basis of stratigraphic and wavelength-mechanical arguments suggested that the SW-NE trending Vredefort Discontinuity (VD) through the northern portion of the dome along the transition from Outer Granite Gneiss to Inlandsee Leucogranofels could be linked with the BRDZ and postulated the existence of a second major décollement zone sympathetic to the BRDZ and running through the centre of the dome. They related this feature to the MBF. The here introduced MDZ coincides with the hypothetical feature of (2). Movements along the MBF and BRDZ are thought to have occurred between ca. 2.3 and 1.95 Ga ago. Times of formation of and movements along both the VD and MDZ still need to be determined.

Conclusion: An explanation of the Vredefort structure necessitates firstly, knowledge of structures present in both the Archaean rocks (basement gneiss and greenstone complexes) and the Precambrian rocks, and secondly, it is required to differentiate between structures formed pre and post the deposition of the Dominion and Witwatersrand Sequences, and also to separate these from those structures which may be related to the generation of the Vredefort structure. Thirdly, an understanding of the causative tectonic processes that gave rise to the structures in both the Archaean and Precambrian rocks relative to the process(es) that formed the Vredefort structure is necessary. Since 1927 (9) only small individual areas of the Vredefort structure have been mapped in detail. Recently several major structural features (VD, MDZ) have been discovered. In the light of the contrasting genetic hypotheses for the Vredefort structure it is vital to create a basis of structural and chronological data for the whole area of the dome. Only after this has been established will it be possible to separate the distinct events in the history of this region and compare them with phases in the evolution of the entire southern Kaapvaal craton. It is suggested that a first major collaborative effort should be made to collect structural data from critical areas on the dome and to generate a base of chronological data by attempting to determine radiometric ages for mafic intrusives, so abundant in both the Vredefort structure and Witwatersrand environs. A grid of magmatic ages could then be used to place structural events into relation. Direct dating of pseudo- tachylite from Vredefort and Witwatersrand basin could also further contribute to our understanding of the evolution of this important region.

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HOW DEEP ARE ANCIENT ASTROBLEMES ERODED?

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The morphology of lunar craters degrades only slowly by formation of new impacts and their ejecta. It is not the same on earth because of weathering and tectonism. Generally the typical crater topography has disappeard and the term impact structure or astrobleme is applied (1). As on the Moon, two classes of craters are recognized with a transition from simple-to-complex, sharp and inversely proportional to the local gravity (3 km for 10 m/sec^2, 30 km for 1 m/sec^2) (2).

The small, simple craters, because they are generally recent, retain some of their morphology. The best studied craters present an apparent depth/diameter ratio of 0,14 ± 10% and a true depth/diameter ratio of 0,32 ± 20% and about 40% of allochthonous breccia inside the crater.

Using the measurements of nine astroblemes (3) a scaling model is presented for two types of readjusted craters: a well readjusted model based on the data from Charlevoix, Canada (4) and a partly readjusted model based on the data from the Ries, Germany (5). Both possess a central uplift and a ring-graben. The mechanism of readjustment is suggested after the measurements of Charlevoix where the relief is more than one thousand meters and there is a good cross-section, from rim to rim, along the St-Lawrence fault near the middle of the structure.

The model supposes the existence of a transient excavated crater, similar to the present day small craters, with a conical shape, slightly less in depth but more open, and the formation of inverse listric slip surfaces, behind the shock wave front, during the descent of the impactor. The readjustment is only a reequilibration by gravity along the same gliding surfaces and rebound and collapse of the border are believed to be only minor phenomena. The spreading out of 60% of the excavated material on the surface of the readjusted crater obscures the ring-graben of the larger craters on the Moon, but lunar crater Taruntius displays an annular depression which can be due to packing down of soft allochthonous breccia in the ring-graben. The scaling model of readjustment (figure 1) fits well with the astroblemes studied, except for the Ries and Sudbury, Canada, which are only partly readjusted with a central basin filled with 20% of excavated breccia.

The erosion of small craters down to the bottom of the breccia lens, r equivalent to one-third of the diameter of the crater, will leave only few traces of the impact. On the contrary, the allochthonous breccia of the astrobleme will disappear when the erosion attains only two-hundredths of the diameter. But to a

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depth equivalent to one-fifth of the diameter, well below the impactite layer, it is possible to recognize displaced blocks of the readjusted crater with mylolisthenite or pseudotachylite. In fact, in the astroblemes studied, the slip surfaces and their specific breccia would be visible even to 4 to 28 km below sea level, depending of their size.

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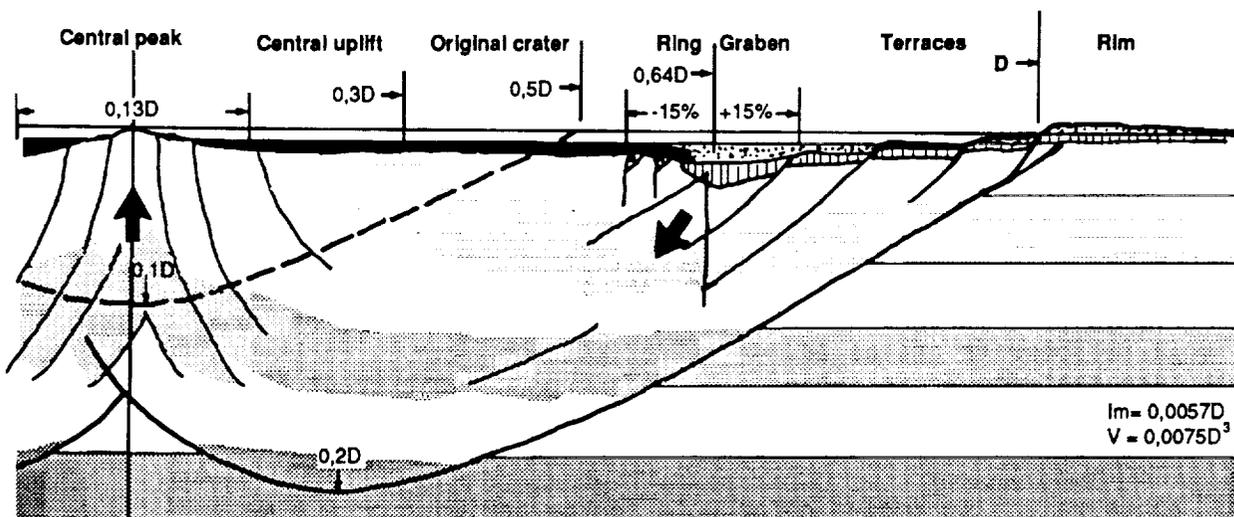
Figure 1: Scaling model for astroblemes more than 20 km in diameter, based on Charlevoix measurements, D = diameter of the astrobleme; 0,5 D = diameter of the excavated crater; black = impactite and breccia; V = volume of excavated crater; Im= thickness of impactite and breccia; dotted = allochthonous breccia; vertical lines = Ordovician cover in Charlevoix; stripes indicate hypothetical horizontal layers.

ACCRETION AND BOMBARDMENT IN THE EARLY EARTH-MOON SYSTEM: THE LUNAR RECORD. Graham Ryder, Lunar and Planetary Institute, 3303 NASA Rd. 1, Houston, TX 77058, USA.

The Earth and Moon have been in mutual orbit for more than 4.5 Ga, hence have been subjected to the same population of potentially impacting bodies. The direct record for collisions with the Earth has been severely modified by internal processes, and that before 3.8 Ga has been entirely obliterated. The Moon is the place to establish as fully as possible the cratering record for the Earth-Moon system: the lunar highlands is an intensely-cratered terrain, and the mare plains are less-intensely but still clearly cratered. Samples collected by the Apollo missions demonstrate that the record of cratering visible on the Moon extends back beyond 3.8 Ga. In this abstract I emphasize impacting in the period prior to 3.8 Ga, i.e. that recorded in the densely cratered highlands that includes the larger basins.

The canonical view for the last decade has been that the bombardment was one of declining accretion, with possibly small, essentially random spikes superimposed [1], and was intense enough to create a megaregolith tens of kilometers deep. Several authors have recognized the problems that the late large basins, particularly Imbrium and Orientale, create for such a paradigm, but have managed to find mechanisms that allow for such impactors in a declining, heliocentric, accretional model [2]. However, recent advances in both lunar sample science and planetary accretion models demand a new look at the assumptions on which these bombardment views are based. From this reinspection [3,4,5,6], I conclude that

a) because the Moon accreted very fast and from material in geocentric orbit there was little lingering helicoentric accretion. In the model of formation of the Moon from material injected into Earth orbit following a giant collision [7], the Moon's accretion is later than heliocentric accretion (the large impactor is one of the latest events, and the Earth had formed and differentiated). Accretion of a Moon from geocentric orbit is rapid. The Moon was in place fully formed, had differentiated, and produced a solid crust by 4.45 Ga, as shown by the ferroan anorthosite suite and other plutonic rocks.

b) there was no impacting of note in the period from 4.4 Ga to 4.0 Ga. No lunar impact melt older than 4.0 Ga has been found, yet continued resetting in a heavy bombardment is not the cause: most ejecta is cold, only impact melts are reset, and most of the Apollo samples are NOT impact melts. Old mare basalts (as old as 4.3 Ga) and plutonic rocks (as old as 4.45 Ga) show that resetting was not a major factor. The absence of impact melts of this period must result from the paucity of impacting. The crust retains lateral and vertical heterogeneities on scales of a few tens of kilometers that would tend to have been obliterated under the heavy bombardments commonly postulated (e.g. [8]). The lunar basin impact melts of about 3.85 GA contain abundant pristine material such as norites that show they are excavating from depths not previously plumbed.

c) at about 3.9-3.8 Ga there was a cataclysmic bombardment of the Moon that created most of the visible landforms in the highlands, including all the basins. All lunar impact melts and nearly all granulites have ages between 3.8 and 3.95 Ga. These melts have a wide variety of ages and compositions, and represent many distinct events: at least 11 at the Apollo 14 site. They

include samples most credibly associated with the Imbrium, Serenitatis, and Nectaris basins, but also numerous smaller events. The varieties cannot be ascribed to a single event, such as Imbrium. New data strengthens this picture of 3.8-3.95 Ga ages of melts [5]. Older ages are derived from igneous rocks and cannot be shown to have any relationship with impacts, except for one granulite.

d) the total amount of material that was added to the Moon in the period 4.4 to 3.8 Ga was only a few tens of meters equivalent, not the kilometers often claimed. The abundance of siderophile elements in the upper lunar crust as represented by feldspathic breccias represents less than 0.25% meteoritic material. Even if this represents a 100 km crust it would correspond with only 250 meters of meteoritic material, but the abundance must decline with depth; if it represents say 15 km then the total meteoritic material is only 40 meters equivalent [6]. Other siderophiles were added in the impact melts of the late cataclysm, but the total abundance remains low.

The population of late impactors probably originated in the Earth-Moon system, because its existence was so brief and intense. It may have been the result of the collision of two other (perhaps considerably smaller) Moons. The siderophiles in the projectiles that were part of the cataclysm are unlike known meteorite groups and are chemically somewhat complementary to the Moon [9]. The same population would have impacted the Earth. The impacting population subsequent to 3.8 Ga was chondritic according to lunar regolith data, hence of heliocentric origin.

The bombardment inferred is summarized in Fig. 1, with the more intense, mega-regolith-producing curve of [1] shown as a dashed curve for comparison. The peak at 3.9-3.8 represents the late cataclysm (geocentric?) following a much lower accretion of heliocentric material in the 4.4 to 3.9 Ga period.

Figure 1.
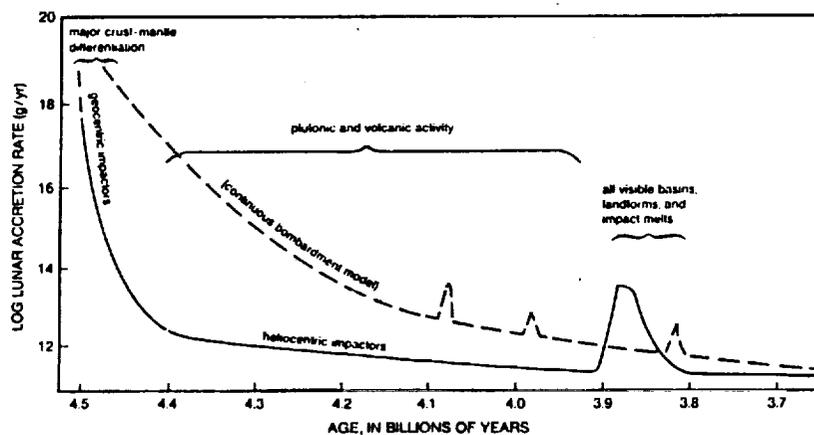
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LARGE PRECAMBRIAN IMPACT SITES LACKING THE USUAL CRITERIA FOR SHOCK METAMORPHISM. John M. Saul, ORYX, 3 rue Bourdaloue, 75009 Paris, France

It is frequently argued that gradual erosional effects caused by the Earth's atmosphere and by H_2O in its liquid state, along with the plate-tectonic regime which characterizes our planet, have led to the total disappearance of all terrestrial vestiges of the Late Heavy Bombardment (LHB) of 3900-4000 million years ago. There is, however, no compelling theoretical reason to assume that such erosional processes have necessarily accomplished the task of erasing each and every vestige of the craters formed during the LHB. There is only the observation that populations of ancient impact craters such as those known from the surfaces of the Moon, Mercury and Mars are not present on our planet.

Since smaller LHB-craters will have inevitably been eroded out of existence long ago, a search for terrestrial vestiges of the LHB might logically begin by looking for extremely large circular scars. For "large" does not only imply wide, deep and complex but also resistant and long-lived.

Such a strategy leads to the discovery of remnants of several very large terrestrial circles, three which are now proposed. Each can be discerned on published geological and tectonic maps but the choice of map projection and coloring scheme play substantial roles in their observation and many maps are not usable.

1) A North American circle some 3700 kms in diameter enclosing most of Canada east of the Rocky Mountains. The Black Hills and the Adirondacks, two areas of "anomalous uplift", help define the southern rim. (For this circle, sheets 2-3 of the UNESCO 1:10,000,000 Geological World Atlas are useful.)

2) A European circle 3400 kms or more in diameter. On the west, the Adriatic is part of a moat and the Italian Peninsula part of a rim while, to the northeast and east, the Kanin Kamen and Timan Ridges and eastern portions of the Urals define a partial rim. (Discernable, for example, on the International Geological Map of Europe and the Mediterranean Region, 1:5,000,000 (UNESCO, 1971).)

3) A yet larger Asian circle some 5500 kms in diameter with the southern Himalayas marking a southern rim and western parts of the Urals defining a western rim. The arcuate islands of Novoya Zemlya seem to be part of this structure, perhaps as vestiges of an outer rim or as a portion of the Uralian rim which has been tectonically displaced. (Other possibilities can be envisaged if the Kara Sea Basin between Novoya Zemlya and the mainland is interpreted as the vestige of yet another LHB impact.)

Once the nature of such scars has been perceived and accepted, many more can be observed. In fact, it turns out to be convenient - perhaps even prudent - to treat all substantial geographic/geological arcs as being "guilty until proved innocent", an approach consistent with the observation that impact craters produced by the Late Heavy Bombardment constitute the most common landform in the inner Solar System.

In addition to their curvature, rim-zones and moats alike are often characterized by perennial and deep-seated weakness and/or high permeability. This may be expressed by vulcanism, trenches, drainage, hydrothermal mineral deposits, hot-springs, faulting and so on almost

LARGE PRECAMBRIAN IMPACT SITES LACKING SHOCK CRITERIA: J.M. Saul.

indefinitely. Indeed, later geological manifestations may be thought of as brush-marks painted (and repainted) on an irregularly damaged canvas whose punctured texture was similar to that of the surface of Mercury or the Moon. Such imagery highlights an additional point, that the present-day fractures are inherited features, in other words, that the cracks are older than the rocks in which they are found (1). Hence the usual criteria for shock metamorphism are absent. This means that the vestigal circular scars described here do not belong to any of the recognized categories of impact craters, terrestrial or extra-terrestrial, and must be treated as a distinct phenomenon worthy of a new name.

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New Developments in Sudbury Geology

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The Sudbury Structure (SS) straddles the present boundary of the Archean Superior Province with the Proterozoic Southern Province of the Canadian Shield in Ontario. The 1850 Ma old SS (1) consists of: brecciated and shock metamorphosed footwall rocks; the norites, gabbros, granophyres, and quartz dioritic Sublayer of the Sudbury Igneous Complex (SIC) (2); and the rocks of the Whitewater Group within the Sudbury Basin, i.e. heterolithic breccias and melt rocks of the Onaping Formation (3), mudstones of the Onwatin Formation and wackes of the Chelmsford Formation (4). The origin of the SS is contentious. Proponents of an exogenic origin (5,6,7) point to shock metamorphism and strong brecciation of footwall rocks as evidence for hypervelocity impact, whereas advocates of an endogenic origin (8,9) maintain that the SS lies at a location in the Canadian Shield of repeated tectonic and magmatic activity and that the SS never was circular. Most impact structures are circular or subcircular.

This abstract reports on some new developments which address some of these contentious issues. The new investigations include a thorough structural analysis and field and geochemical studies on gabbroic rocks occurring within the Sudbury. The SS acquired its present oval surface shape during NW-directed thrusting which culminated in a major reverse shear zone, the Fairbank-Whitson Lakes Zone (FWLZ). Estimates of minimum continuous net displacement across the FWLZ exceed 8 km assuming heterogeneous simple shear. Carbonate concretions in the Chelmsford wackes, used as strain gauges, suggest approximately 36% layer shortening strain. The development of slaty cleavage in the mudstones of the Onwatin Formation may represent 30-80% shortening. Shortening due to buckle folding of the wackes of the Chelmsford Formation occurred after NW-SE strata shortening and can account for about 10% shortening of the central part of the SS. The magnitude of net displacement on several intrabasin strike parallel faults is not known. The faults, however, belong to the NW-directed thrusting event and enhanced the shortening of the NW-SE axis of the SS. All these considerations suggest that the non-eroded SS could have been circular at the onset of tectonic deformation.

Recent geochemical and planetological considerations suggest that the SIC in its entirety represents an impact melt system (10,11). Field observations, however, appear to indicate that some phases of the lower SIC are younger than the upper SIC and that at least one of several types of gabbroic rocks within the Sudbury Basin may be related to the Sublayer of the SIC. These gabbros intrude all formations of the Whitewater Group. They and the lower phases of the SIC, therefore, are unlikely to be impact melts. The granophyres of the SIC, based on these considerations, may still be impact melts, as has been suggested previously (12).

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PROTEROZOIC IMPACT RECORD OF AUSTRALIA; E.M. Shoemaker and C.S. Shoemaker, U.S. Geological Survey, Flagstaff, AZ 86001.

The most complete record of Precambrian impact structures recognized to date has been found on the Australian craton. This circumstance is due to two factors. First, continental basins of Proterozoic age that contain gently to moderately deformed, relatively unmetamorphosed sedimentary rocks are extensively exposed in Australia; structural deformation by impact is fairly readily recognized in these basins. Secondly, much of the Australian craton has been exceptionally stable, not only during the Phanerozoic but also during a significant part of Proterozoic time. Hence many large impact structures have been neither too deeply eroded nor too deeply buried to render them undetectable. We present here a summary of seven structures of known or probable Proterozoic age and the implications that may be drawn from these structures for the cratering rate in Proterozoic time.

Lawn Hill, Queensland.--The Lawn Hill structure (lat. 18°40'S, long. 138°39'E) is about 20 km in diameter. It has been formed in clastic sedimentary rocks of the Lawn Hill Formation of the McNamara Group [1]. Beds of this group are roughly correlative with the Mount Isa Group dated at about 1670 m.y. [2]. Impact postdates an episode of fairly tight folding of the McNamara Group. The central uplift consists of a complex northeast-plunging anticline; shatter cones are well developed and fairly numerous in the uplift. A structural moat surrounding the central uplift is largely concealed beneath limestone and limestone breccia of Middle Cambrian age which rests on a surface of low relief that cuts the folded Precambrian; the impact structure was planed off prior to deposition of Cambrian beds, although a very shallow ring depression is present at the unconformity where it crosses the moat. A local remnant of Cambrian filling this ring drew attention of geologists to the structure.

Kelly West, Northern Territory.--The Kelly West structure (lat. 19°56'S, long. 133°57'E) was first recognized by Tonkin [3] from the presence of shatter cones. The feature mapped by Tonkin is part of the central uplift, which we remapped in detail in 1989. Impact was centered about on the axis of a synclinal inlier of Hatches Creek Group quartzite of Early to Middle Proterozoic age that rests with strong angular unconformity on the Warramunga Group of Early Proterozoic age. Uplift at the center of the structure partly unfolded the preexisting syncline of Hatches Creek beds. Extremely limited exposure of the less resistant Warramunga beds surrounding the central uplift has prevented secure determination of the size of the structure from observations of outcrops. On the basis of the well exposed part of the central uplift, we estimate the diameter of the whole structure to be not less than about 10 km and probably not greater than 20 km. In the Early Cambrian, the resistant Hatches Creek quartzite of the central uplift formed a monadnock that was later buried by cherty limestone of early Middle Cambrian age. The Hatches Creek Group rests on terrain intruded by granites with estimated ages of about 1800 m.y., but the precise age of the Hatches Creek is not yet established. Near Kelly West, more than 5 km of Hatches Creek beds was deposited, folded, and then largely eroded away in Precambrian time. The impact almost certainly occurred after this episode of deep erosion was largely completed, perhaps during the Late Proterozoic.

Strangways, Northern Territory.--A Precambrian age for the 26 km diameter Strangways structure (lat. 15°2'S, long. 133°35'E), identified as impact in origin by Guppy et al. [4], is slightly in doubt. Rocks that are unequivocally deformed by impact include clastic sedimentary rocks of the Roper Group, of Middle to Late Proterozoic age and the granite basement on which they rest. Unpublished detailed geologic mapping by D.J. Milton strongly indicates that the Tindall limestone of Middle Cambrian age overlaps the structural moat on the southwest side and rests on a surface of low relief that truncates the impact structure. However, P.R. Brett and M.R. Dence (personal communication) think that the limestone may have been deformed by impact. Bottomley [5] obtained a rather poorly defined $^{40}Ar/^{39}Ar$ age of ~470 m.y. from impact melt rocks preserved near the center of the structure. Probably the Strangways structure is of Late Proterozoic age.

Lake Acraman, South Australia.--The Lake Acraman structure (lat. 32°01'S, long. 135°27'E) is the largest known Precambrian impact structure in Australia and also the best dated. Williams [6] suggested that the impact structure might be about 160 km in diameter; from our own field observations, we think the crater probably was not much more than about 35 km in diameter, although some faulting or renewed displacement on old faults may have occurred at greater distance, as suggested by Williams. The crater was formed in 1590 m.y. old volcanic rocks, chiefly dacite. Distant fallout from the Lake Acraman impact was discovered by Gostin et al. [7] in the upper part of a ~10 km thick sequence of sedimentary rocks of Late Proterozoic age in the Adelaide geosyncline. The fallout layer occurs about 1100 m below strata known to contain an Ediacaran fossil assemblage and about 1500 m. below Cambrian beds. Impact occurred very late in the Proterozoic, probably near 600 m.y. B.P. [7]

Spider, Western Australia.--The Spider impact structure (lat. 16°44'S, long. 126°05'E), discovered by J.E. Harmes [8], was mapped in detail by us in 1984 and 1985 [9]. It is about 13 km long and 10 km wide and was formed apparently by low oblique impact into nearly flat-lying beds of the Kimberley Group of Early Proterozoic age. The present land surface at the Spider structure is not far below an ancient erosion surface, that is capped nearby by glacial deposits of Late Proterozoic age. We suspect that a substantial section of Lower Proterozoic rocks was present at the time of impact.

Teague Ring, Western Australia.--The Teague Ring (lat. 25°52'S, long. 120°53'E) is a 30 km diameter impact structure first reported by Hadyn Butler [10] and mapped in greater detail by Bunting et al. [11]. Beds of the Eraheedy Group of Early Proterozoic age and granitoid rocks of Archean age were deformed by the impact. On the basis of Sr/Rb isotopic studies, Bunting et al. [12] suggested that a quartz syenite within the central uplift may have been emplaced at

1630 m.y. coincident with the deformation. We interpret the syenite as belonging to the Archean complex; strontium isotopes evidently were reset at about 1630, which probably reflects a pulse of shock metamorphism.

Fiery Creek Dome, Queensland.--The Fiery Creek Dome, (lat.19°13'S, long. 139°13'E) is a circular structure 30 km in diameter consisting of a central uplift surrounded by a structural moat, which has been modified by a younger regional network of close-spaced faults [13]. We interpreted this feature as a possible impact structure [14] on the basis of a reconnaissance examination in 1987. The structure is developed chiefly in clastic sedimentary rocks of the Haslingden Group of Middle Proterozoic age. These rocks are thought to be about 1700-1800 m.y. old [13]. On the north flank of the structure these rocks are truncated and overlain unconformably by the Surprise Creek Formation and locally by the Fiery Creek Volcanics, both of Middle Proterozoic age. Rocks that probably are correlative with the Fiery Creek Volcanics have been dated at 1678 m.y. Development of the dome and structural moat predates the Fiery Creek Volcanics and probably occurred close to 1700 m.y. Further work is planned prior to the conference to test the impact origin of the dome.

The cumulative size-frequency distribution for the 7 known and probable impact structures of Proterozoic age is illustrated with the solid line with one sigma error bars in Fig.1. If the Fiery Creek dome proves not to be of impact origin, the cumulative frequency would follow the dashed line. From analysis of the size-frequency distribution of craters of Copernican and Eratosthenian age on the Moon mapped by Wilhelms [15], and observations of Earth-crossing asteroids and comets [16], we infer that the expected size-frequency distribution for Proterozoic impact structures on Earth, in the diameter range of 10 to 30 km, has a form represented approximately by the dotted line in Fig. 1 ($N_D \propto D^{-2.27}$, where N_D is cumulative frequency and D is crater diameter in km). A curve of this form that passes through $N_D = 5$ at D = 20 km represents the observed frequency between 20 and 35 km diameter satisfactorily (i.e. well within the one sigma errors); the sum of the squared deviations of log frequency are close to a minimum for this fit. This curve also lies close to the cumulative frequency in the case that Fiery Creek Dome is dropped from the count. Discovery of exposed impact structures equal to or larger than 20 km diameter formed in stratified Proterozoic of Australia may be complete or close to complete. Extrapolation of the dotted curve suggests that about 2/3 of the exposed Proterozoic impact structures larger than 10 km diameter remain to be found.

The area of exposed stratified Proterozoic rocks in Australia is about $1.15 \times 10^6 km^2$. (Exposure here is taken to mean the mapped geologic unit at scales of 1:250,000 and 1:5,000,000.) We include only stratified rocks, as experience to date indicates that impact structures are extremely difficult to recognize in complex metamorphic or plutonic terrains. Roughly comparable areas are exposed of Early Proterozoic strata (2500 m.y. to 1800 m.y.) and Middle Proterozoic strata (1800 m.y. to 1000 m.y.); Late Proterozoic strata (1000 m.y. to 570 m.y.) are exposed over an extensive but somewhat subordinate area. The mean age of the exposed Proterozoic strata evidently is fairly close to the midpoint of Proterozoic time, about 1540 m.y. Assuming that most of the exposed Proterozoic strata have seldom been so deeply buried as not to record the structure associated with ≥20 km craters, the mean time of exposure of these rocks to the production of impact structures during the Proterozoic is about 1540 m.y. - 570 m..y. = 970 m.y. The estimated mean Proterozoic crater production rate for craters ≥20 km diameter, then, is $(5 \pm 2.2) / 1.15 \times 10^6 km^2 / 0.97 \times 10^9 yr = (4.5 \pm 2.0) \times 10^{-15} km^{-2}yr^{-1}$. This rate is close to the estimated present rate of crater production by asteroid and comet impact, $(4.9 \pm 2.9) \times 10^{-15} km^{-2}yr^{-1}$ [16] and to the rate $(5.4 \pm 2.7) \times 10^{-15} km^{-2}yr^{-1}$ estimated by Grieve [17] from the terrestrial geologic record of impact for the last 120 m.y.

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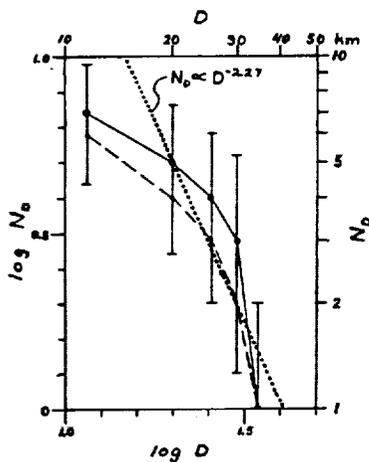
Figure 1

IMPACT SIGNATURES IN THE GEOLOGICAL RECORD; D. Stöffler,
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On the Earth's surface some 120 impact craters have been re-
corded so far (1). The criteria for recognizing an impact crater
vary with its size and state of erosion. Small bowl-shaped cra-
ters become complex flat craters with central uplifts and ring
structures if they exceed diameters of about 5 km (2). Since al-
most all pre-Phanerozoic craters are large complex structures
which survived erosion, this review will be focussed on such
craters.

Diagnostic signatures of impact craters can be defined on
megascopic, macroscopic, and microscopic scales. The megascopic
signatures include mainly morpological and structural
characteristics of craters and the geological setting of breccia
formations. Macroscopic signatures relate to the petrographic
and chemical properties of breccias. Microscopic signatures are
specific shock effects in minerals, the degree of shock metamor-
phism of target rocks and the textural-mineralogical properties
of monomict and polymict breccias. In dealing with this variety
of diagnostic features it is useful to distinguish between the
outer impact formations (rim structure and ejecta blanket) and
the inner impact formations (breccia lens and crater basement)
(3). Independent of the geological setting of these formations,
they belong texturally to either one of the following macro-
scopically recognizable groups of impact-metamorphosed target
rocks: (a) monomict breccias, (b) polymict breccias (either with
clastic or with melt matrix) and (c) shocked rocks ranging from
mechanically deformed to molten. These breccias and shocked
rocks reflect a shock pressure range between about 1 to 10 GPa
and about 80 to 100 GPa (1 GPa = 10 kbar).

In most cases the outer impact formations of terrestrial
complex craters are not preserved and all known characteristics
are exclusively based on the Ries crater (4) whose target stra-
tigraphy is probably atypical for Precambrian terranes. There-
fore, the following discussion will be restricted to the inner
impact formations of complex craters because they are diagnos-
tically more important for the recognition of old craters (1, 5,
6). They can be subdivided into the following depth zones from
top to bottom: (a) allochthonous polymict breccias (fragmental
breccias and melt breccias), (b) parautochthonous crater floor
zone, (c) shocked and brecciated autochthonous crater basement.
A model section of the central part of a complex crater can be
summarized as follows (3, 5, 6): 1.) A layer of a polymict, melt-
bearing (suevitic) fallback breccia forms the top of the alloch-
thonous breccia lens. It is underlain by an impact melt sheet
(clast-bearing crystalline matrix breccia) which, with increa-
sing radial range, rests upon the brecciated autochthonous to
parautochthonous uplifted basement, on a layer of suevitic brec-
cia, and on a layer of fragmental breccia without melt particles.
In larger complex craters with ring structures the melt sheet

forms an annulus around the central uplift. At some radial range the breccia lens contains a complete layered sequence of melt breccia, suevitic breccia, and fragmental breccia going from the top to the base of the lens. 2.) In a transition zone between this allochthonous breccia lens and the parautochthonous crater basement "injection" dikes of impact melt and of suevitic and fragmental breccias, mostly decimeters to meters thick, occur as indicators of the floor zone of the transient crater. The brecciated basement rocks of this zone display shock effects up to 30 GPa as recorded by tectosilicates and mafic minerals. 3.) Below the crater floor zone the basement is affected by monomict brecciation, in-situ displacement of megablocks, and by a variety of breccia dikes cutting or separating the megablocks: pseudotachylite dikes formed during shock compression and dikes of fragmental breccias (containing clasts of pseudotachylite dikes). Shock-induced planar elements in quartz persist to some depth and radial range in this zone which also displays shatter cones. Brecciation and clastic dike breccias continue to even greater depth and radial range.

As observed in subhorizontal erosional sections of complex craters the described impact metamorphic signatures vary with radial range causing circular zones of impact effects (5, 7, 8, 9). In an upper stratigraphic section this zoning is related to various types of breccias which extend to increasing radial ranges in the following sequence: impact melt breccias, suevitic breccias, fragmental breccias, and dike breccias. At a deeper stratigraphic level the zoning involves: isobars of decreasing shock pressure, radial ranges of shatter coning and of various types of dike and monomict breccias, as well as subvertical fault and shear zones including the peripheral fault system near the structural crater rim which causes the annular depression zones observed in deeply eroded craters. The exact relation of such zones of impact-induced phenomena to the geometrical parameters of a crater such as the radius of the transient or of the apparent crater is not absolutely clear. A tentative synthesis of current observations and interpretations has been given by (5).

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IMPACTS AND ATMOSPHERIC EROSION ON THE EARLY EARTH; A. M. Vickery, Lunar and Planetary Laboratory, The University of Arizona, Tucson, AZ, 85715, U.S.A.

Until recently, models for the origin and evolution of the atmospheres of terrestrial planets ignored the effects of accretionary impacts. In the 1970's, however, it was suggested that heating and/or vaporization of accreting carbonaceous-chondrite-type planetesimals could result in the release of their volatile components (1,2). Modeling of this process (e.g., 3,4) strongly suggests that substantial atmospheres/hydrospheres could develop this way. During most of the accretionary process, impact velocities generally differed little from the escape velocity of the growing proto-planet because most of the collisions were between bodies in nearly matching orbits. Toward the end of accretion, however, collisions were rarer but much more energetic, involving large planetesimals and higher impact velocities (5). It has been postulated that such impacts result in a net loss of atmosphere from a planet, and that the cumulative effect impacts during the period of heavy bombardment might have dramatically depleted the original atmospheres (6,7).

Walker (8) showed that shock heating and compression of the atmosphere by the projectile during entry can eject at most a few times the mass of the air traversed, which is generally a negligible fraction of the total atmospheric mass. The solid ejecta are also unable to eject more than a few times the mass of the air traversed by the projectile (9). The vapor plume produced by a sufficiently energetic impact is, however, capable of ejecting the entire atmospheric mass lying above a plane tangent to the planet at the point of impact (10). Models developed to study atmospheric erosion by impacts on Mars and the interaction of the vapor plume produced by the KT impactor on earth (11,12) are here applied to the case of the evolution of earth's atmosphere.

The simplest model involves estimating the minimum impact velocity and impactor mass required to eject the atmospheric mass above the tangent plane (M_{tp}) and concatenating this information with estimates of the impact flux. A model for vapor plume expansion (13) gives the mean expansion velocity as $[2(\epsilon - \Delta H)]^{1/2}$, where ϵ is the initial internal energy of the vapor and ΔH is the vaporization energy. The internal energy of shocked material is $u^2/2$, where u is the particle velocity; for projectile and target of similar materials, the peak particle velocity is roughly half the impact velocity. By requiring that the mean expansion velocity exceed escape velocity, and using $\Delta H = 13$ MJ/kg for silicats, the minimum impact velocity for atmospheric blow-off on earth is ~25 km/sec. Simple momentum balance suggests that the minimum impactor mass for blow-off is $m^* = M_{tp}$. The evolution of atmospheric mass with time is then given by

$$\frac{dM_{atm}}{dt} = -N_{cum}(m^*,t)4\pi R^2 M_{tp}$$

where $N_{cum}(m,t)$ is the cumulative number of impactors with masses greater than or equal to m (per unit area and per unit time) and R is the radius of the target planet. Using the approximation that $M_{tp} = H/2R$, where H is the scale height of the atmosphere, allows this equation to be integrated to find $M/M_0 = P/P_0$, that is, the ratio of the atmospheric mass (or pressure) at any time to its current value. Using this equation, $P(-4.5$ Gyr$) \cong 5 \times P_0$ for the earth (Figure 1). This contrasts sharply with the results of similar calculations for Mars, for which $P(-4.5$ Gyr$) \cong 100 \times P_0$. For both planets, however, the atmospheric loss rate is greatest during heavy bombardment and has been negligible since the end of heavy bombardment.

These calculations implicitly assume that the atmosphere is 'distributed homogeneously with respect to zenith angle, but the atmosphere is in reality concentrated near the horizon. More detailed numerical work, which takes this inhomogeneity into account, suggests that $m^* = 5$ to $10 \times M_{tp}$. This makes stmospheric erosion by impacts less efficient. Other factors tend to make atmospheric blow-off more efficient than these models indicate. First, the latent heat of vaporization will be added back to the internal

energy of the vapor plume as the material begins to condense. Second, when the acceleration due to pressure gradients with the plume becomes comparable to the acceleration due to gravity, the plume will descend below the tangent plane and so a single impact may blow off more than Mtp. Third, these calculations ignore partial blow-off, that is, loss of less than Mtp; because partial loss may occur for smaller but more numerous impactors, the net effect may be significant. Fourth, the effect of obliquity of impact has been neglected. Experiments suggest that oblique impacts produce more vapor than normal impacts with the same impactor mass and speed (14). Furthermore, this vapor has a velocity component downrange, which means that it is directed toward the highest atmospheric mass concentration. Oblique impacts may thus be much more efficient at ejecting atmosphere than normal impacts.

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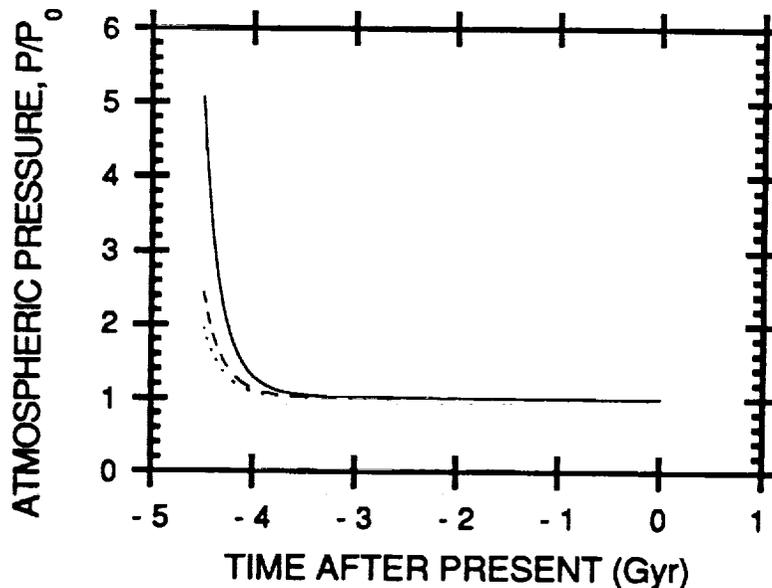
LARGE IMPACT STRUCTURES ERODED BEYOND THE IMPACT MELT SHEET: SIGNIFICANCE OF THE IMPACT MELT DYKES INJECTED IN THE FUNDAMENT

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A comparison of the prevailing size of known impact structures on the Earth (D=100 m to 100 km) with variable erosion rates of the continental crust segments during time confirms that a high percentage of impact structures was erased by erosion. Consequently, there must be also a significant number of as yet unrecognized large impact structures which are exposed as deep erosional sections. Specialists in the field of impact structures concentrate preferentially on well preserved structures as these examples carry the richest information on the impact process.

It is suggested that strongly eroded structures should receive more attention in view of (1) a significant number of structures at these exposure conditions, (2) better understanding of the effects of large impacts on the middle and lower levels of crust, (3) significance of this data for the cratering record, (4) improving the criteria for recognition of impact structures from endogenic circular structures.

Distensional deformational flow of crustal masses accompanying excavation of transient craters (1) results in a brief opening of fractures and their filling by breccia and impact melt. Deep and quick penetration of the injection dyke material is facilitated by vacuum pumping. Injection dykes belong to the group of deep subsurface products accompanying impact crater formation. Similar to the melts injected into the crater floor breccia (2), the injection dykes are likely to contain pivotal portions of melt which is more contaminated by projectile.

There are but few examples of well studied probable impact melt dyke rocks from deeply eroded structures (3). This contribution presents data for two types of dyke rocks (quartz monzodiorite and microgranodiorite) from the Sevetin structure, D=46 km, (4,5), both showing a similar increase in Ni,Cr,Co, and Mg relative the potential crustal parent rocks (i.e. mafic granulite and sillimanite-biotite paragneiss respectively). Data for 53 elements and the concentrations of Ni,Cr,Co,Ir, and Mg show that monzodiorite and microgranodiorite could be derived from the above crustal rocks plus 5-7 wt. % of a component similar to primitive mantle. However, since monzodiorite has a higher REE content and a higher La/Yb compared to mafic granulite and microgranodiorite has a lower HREE content and a higher La/Yb compared to paragneisses, and since Sr and Nd/Sm isotopic work is not completed, these dyke rocks have no better status than probable impact melt rocks.

Lechatelierite, very fine-grained skeletal zircon, Cr-bearing ilmenite, and the presence of four pyroxenes (monzodiorite) point to very high temperatures during the early stages of dyke solidification and to high cooling rates in the range of high temperatures. Chlorite spheroids, distinct from local vesicles, persistent through all known localities of monzodiorite, are interpreted as former drops of an immiscible melt (2). These features support interpretation of monzodiorite and microgranodiorite as impact melt rocks.

It is suggested that search for insoluble residue minerals in holocrystalline impact melt rocks should be undertaken by specialists on insoluble minerals in meteorites. These informations may provide additional contraints on the origin of rocks potentially belonging to this category.

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CALCULATION OF MASS AND VELOCITY DISTRIBUTIONS OF TERRESTRIAL AND LUNAR IMPACTORS BY USE OF THEORY OF PLANETARY ACCUMULATION.

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This is a preliminary report of an overly-ambitious project of calculating the complete size and velocity distribution of the population of objects in the inner Solar System from the beginning of planetary accumulation up until about 3.5 b.y. ago, by which time the Solar System had approached its present steady-state, as indicated by the near-constancy of the lunar and terrestrial impact flux since that time. Although such a complete synthesis is undoubtedly premature, the present work is expected to serve several purposes:

(1) To identify more clearly what problems must be solved before a more mature theory of this kind can be developed, and to distinguish between problems that are real difficulties and those that represent details, that although poorly understood, are likely to be of minor importance.

(2) To quantify presently qualitative "scenarios" of the sequence of events during this time in Earth history, and to understand the non-linear and often counter-intuitive interaction of collisions, gravitational perturbations, growth, fragmentation, and radial migration.

(3) To begin to compare theory with observation of the cratering and meteoritic record, and to at least provide predictions of the 4.55–4.1 b.y. terrestrial and lunar impact flux that are more constrained than those obtained by pure assumption of, e.g., a power law with an assumed exponential index.

It is found from these calculations that in the terrestrial planet region the accumulation and impact history divides itself naturally into several subsequent stages characterized by distinct size and velocity distributions:

(1) An early stage in which a population of initial ~ 10 km diameter planetesimals rapidly (10^4 to 10^5 years) merge to form about twenty $\sim 5 \times 10^{26}$ g "planetary embryos" in nearly circular orbits and with very low inclinations. Throughout this stage all velocities remain low, $\lesssim 100$ m/sec. This is a consequence of gravitational interaction forces that tend to equipartition the energy of the random component of motion of the growing planetesimals ("dynamical friction"). In the presence of nebular gas, expected to be present during this early stage, this has the effect of requiring the larger bodies to lose energy by continually and vainly trying to "pump up" the eccentricity and inclination of the smaller bodies. This is prevented by gas and collisional dissipative drag on the smaller bodies, causing the velocities of all the bodies to remain low. These effects are not influenced in a major way by deviations from the two-body formalism at very low relative velocities. The size distribution during this stage is trimodal: about 20 rapidly growing runaway bodies, isolated from significant interaction with one another; a large number ($\sim 10^7$) 10 to 100 km bodies, merging with one another and being swept up by the large runaway bodies; a collisionally-produced "tail" of < 1 km bodies, containing $< 1\%$ of the total mass. Toward the end of this stage of growth the characteristic time scale for significant growth increases rapidly from 10^4 to 10^5 years to 10^6 years.

MASS AND VELOCITY DISTRIBUTIONS OF IMPACTORS: Wetherill, G.W.

(2) The next stage consists primarily of the merger of the embryos with one another to form the present terrestrial planets, on a time scale of 10^6–10^8 years. The transition between the first and second stages has not yet been calculated in a "hands off" manner. There is some possibility that this transition is spontaneously triggered by the loss of gas and by the sweeping up of almost all the bodies smaller than $\sim 10^{25}$g, eliminating the ingredients for the gas drag-dynamical friction feedback mechanism that maintains the entire planetesimal ensemble at low velocities during the first stage. This possibility must be considered speculative at present, however. In any case, an assemblage of simply $\sim$ twenty 5×10^{26}g embryos in nearly circular non-crossing orbits will be unstable with respect to a transition to a high relative velocity state, characterized by a high degree of orbital crossing and acceleration to velocities of 1 to 10 km/sec, even though the mechanism that induces this transition is not known. During this stage the terrestrial planets grow to near their present mass primarily as a result of "giant impacts" between growing embryos. A new, and larger ($\sim 10^{27}$g) population of collision debris is produced during this stage by mutual collisions between embryos, and cratering of embryos by these collision products of embryo-embryo collisions. The steady-state size distribution of these collisionally produced bodies has not yet been completely calculated, but preliminary results suggest it contains an excess (over conventional power-law distributions) of basin-forming 10 to 500 km diameter impactors. Toward the end of this stage ($\sim 10^8$ years), perturbation of material by Earth and Venus into hyperbolic solar-system escape orbits becomes the dominant process for loss of residual material.

(3) A final stage of diminishing bombardment of the terrestrial planets, and the moon, by this residual population of basin-forming and smaller bodies. After $\sim 10^9$ years, the contribution to Earth-cratering of this residual terrestrial planet material becomes considerably smaller than that obtained from other sources — comets and asteroids, although some of the present "asteroidal" flux may have originally been terrestrial planet debris transferred into quasi-stable asteroidal orbits during the second stage of growth. It is hoped that further results regarding the size and velocity distribution expected during this final stage can be presented at the Workshop.

IMPLICATIONS OF EARLY CRATER-CENTERED VOLCANISM AND TECTONISM AT THE SUDBURY STRUCTURE, ONTARIO.

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Introduction: The Sudbury structure in Ontario, Canada is one of the oldest and largest impact structures recognized in the terrestrial geologic record (1). It is also one of the most extensively deformed and volcanically modified impact structures on Earth (2,3,4). Numerous impact craters on the Moon have been volcanically and tectonically modified (5) and provide possible analogs for deciphering the initial size, form and structure of Sudbury. In this study, we compare patterns of deformation at Sudbury to possible lunar analogs and derive estimates of the initial Sudbury crater rim and floor diameters with implications for impact patterns in the Archean.

Structure Descriptions: Two patterns of deformation can be distinguished at Sudbury: 1) an early sequence centered on the impact delineated by the Sudbury Igneous Complex (6) and 2) several later episodes of regional deformation which cut basin-controlled features, ie. are insensitive to the impact structure. The Main Igneous Complex presently defines an elliptical ring about 60 km long and 27 km across. This norite/micropegmatite layered intrusion has a crystallization age of -1850 Ma, an age commonly assigned to the time of the impact (7). It also feeds an extensive sequence of offset dikes in the surrounding basement rocks radial and concentric to the undeformed structure (6). The radial dikes are the most evenly distributed and, although disrupted by later deformation, they can extend up to 30 km from the edge of Main Igneous Complex. The less extensive concentric dikes mostly occur south of the structure where they are typically about 3-10 km from the Main Complex (6). Lastly, basin-centered lineaments, apparently not linked to the Igneous Complex, can be identified in satellite images 20-30 km north and west of the structure (8). Unfortunately, deformation along the Grenville front to the south and at the Wanapitae impact to the east mask any similar trends elsewhere around the basin.

Haldane crater on the Moon also exhibits an extensive volcanic and tectonic modification. Haldane is a multi-ringed structure with an outer rim diameter of 40 km and an uplifted central "floor plate" separated from this rim by a wide (-5 km) moat structure (9). Although classified as a floor-fractured crater, it exhibits an extreme stage of modification with structural and volcanic processes extending beyond the crater rim (5). The central floor region has been uplifted -3 km and retains a central peak complex, some of which rise above the outer rim. Concentric fractures occur near the edge of the uplifted floor plate, in the moat or in a well-defined annulus 9-16 km beyond the crater rim. Radial fractures are restricted to the central plate and moat regions. Although these radial fractures typically do not continue past the crater rim, one set east of the crater extends over 10 km past the crater rim with additional concentric fracturing between the crater rim and the outer fracture ring (9). More generally, radial fracture patterns in less modified craters (eg., Petavius, Humboldt, Schrodinger) never extend beyond the crater rim (5). Crater counts indicate that both the crater and the superposed fracture systems formed at nearly the same time, and volcanism in Haldane appears to be coeval with other basalt units in Mare Smythii. Lavas appear to flow out of the moat to the northwest from inside the crater rim (9). Mare lavas also may flow southeast from a section of the outer fracture ring; possible pyroclastic deposits occur elsewhere along these fractures (9).

Discussion: The most prominent fracture pattern at Sudbury comprises the radial offset dikes outside the Main Igneous Complex. At Haldane, radial surface fractures are best developed in the central floor plate and moat region with a few extending past the crater rim. Two interpretations for the radial dikes are therefore possible. If the Main Igneous Complex marks the location of the crater rim, then the exterior radial dikes might correspond to the fractures extending beyond the eastern rim of Haldane. The even distribution around Sudbury, however, requires a uniformly tensile stress field during deformation. Since dike formation appears to coincide with the Penokean Orogeny and regional compression (10), such a stress field seems unlikely. Alternatively, the Sudbury dikes may be analogous to the radial fractures restricted to the floor of Haldane and other lunar craters. In this case, an episode of floor uplift could produce the uniformly tensile stress field provided that the crater rim decouples the floor plate from external stresses. Consequently, the radial extent of the dikes would be limited to the interior of the Sudbury crater.

One observation complicates this simple comparison. At Haldane the only place radial fractures extend out from a major concentric feature comparable to the Main Igneous Complex is in the moat where radial fractures intersect the outer edge of the uplifted floor plate. In direct contrast to Haldane where the central peaks were uplifted to the elevation of the crater rim, the central Sudbury feature does not exhibit the positive relief expected for an uplifted central peak but is a basin containing -3 km of ejecta and sedimentary fill (11,12). This central basin at Sudbury indicates that any uplifted region inside of the Main Igneous Complex foundered (2). Although rare, central peak loss is also seen on the Moon, where a few modified craters (eg. Camoens) have central depressions rather than uplifted central peak complexes (5). Moreover, other lunar craters (eg. Schluter) exhibit fractures encircling the central peak complex. The simple interpretation for

EARLY DEFORMATION PATTERN AT SUDBURY: Wichman, R.W. and Schultz, P.II.

these features is that the central peak complex became detached from the floor plate during uplift. If applied to Sudbury, the Main Igneous Complex then marks the outer edge of a down-dropped block containing the central peak complex; the radial dikes extend away from this central detachment across an uplifted floor plate. Ring fractures identified by (8) to the north and west appear to lie beyond these radial dikes and could correspond to deep-seated faults at the edge of the floor plate or faulting associated with wall terraces. The original crater rim. then. would lie outside of these features.

With this analogy, the patterns of deformation at Sudbury can provide estimates for the original crater floor and rim diameters. Since radial fractures are confined to the crater interior, a minimum value for the crater size becomes ~100-120 km diameter. If the Sudbury basin represents a down-dropped central peak complex, however, an alternative estimate can be obtained. Strain analyses of deformation in the Sudbury basin indicate that it was originally ~60 km by ~40 km (3). If the minimum (40 km) value corresponds to the basal diameter of the central peak complex, then the morphometric relations derived by (13) for unmodified lunar craters would indicate a floor diameter of at least ~100 km with a corresponding rim-crest diameter of ~150 km. Since moats in lunar craters form near the edge of the original crater floor (5), the model diameter of the outer-most concentric lineaments (~100 km) matches the predicted floor diameter very well. The inferred rim-crest diameter is also similar in magnitude to the recent estimate of 180-200 km made on the basis of preserved shock features around Sudbury (14). That estimated range of rim-crest diameters translates to basal central peak diameters of 44-49 km.

As an alternative, Sudbury may represent only the central core of a larger multi-ring structure and has no central peak. In this analogy, the Sudbury basin corresponds to the central depression inside an inner ring with radial fracturing confined to the interior of the crater, as in the multi-ringed lunar crater Schrodinger. The inner ring of two-ring basins on the Moon and other planets corresponds to a certain fraction of the rim diameter (15). If applied to the undeformed diameter of the Main Igneous Complex (40 km), then the original Sudbury rim-crest diameter should not exceed 140 km.

Implications: For Sudbury on the Earth, the development of impact-centered deformation apparently occurs soon after the time of impact. Indeed, the times of dike and fracture formation are currently indistinguishable from the time of impact. Since most terrestrial craters do not exhibit comparable degrees of deformation, the post-impact conditions at Sudbury must have been significantly different from those associated with other craters on the Earth. One possibility is that Sudbury, Haldane and Schrodinger formed in regions with anomolously high thermal gradients and locally thin lithospheres with respect to the size of the impactor. For Sudbury, enhanced heat flows and lithospheric thinning associated with the Penokean Orogeny apparently coincide with the crater-forming event (10). For Haldane, lithospheric thickness at the time of deformation appears to have been less than 10 km (16), possibly due to the effects of local mare volcanism. The large size, time of formation and placement of Schrodinger is also consistent with this perspective. If the style of deformation at Sudbury requires locally thin lithospheres at the time of impact. then structures such as Haldane (and lesser modified craters such as Schrodinger) could have typified craters during the Archean when stable cratons were just being established and tectonically active regions of thin lithosphere may have been more widespread. As a result, recognition of ancient impact sites and proper assignments of crater size could be difficult without reference to preserved analogs on other planetary surfaces.

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IMPLICATIONS OF IMPACT INTO VISCOUS MANTLE RHEOLOGIES FOR CRATER PRESERVATION AND MANTLE EVOLUTION; R. W. Wichman and P. H. Schultz. Dept of Geological Sciences, Brown University, Providence, R.I. 02912.

Introduction: In the standard model of crater excavation, the transient cavity represents target material displaced through a combination of ballistic ejection and downward displacement of shock-compressed materials (e.g., 1, 2, 3, 4, 5). In large craters, the final crater form results from collapse of this transient cavity with uplift and inward flow of the displaced material comprising the crater floor (6). Although this idealized model of crater formation fits most well-preserved planetary impact structures, it implicitly assumes an elastic halfspace beneath the target surface, an assumption that may not be appropriate for large terrestrial impact events 3 to 4 billion years ago. For such impacts, the depth of the transient cavity may exceed the lithospheric thickness, and interaction with an underlying viscous mantle regime during both cavity formation and cavity collapse becomes possible (7,8). In this study, we consider the implications such viscous responses might have for the preserved crater signatures and subsequent mantle evolution (irrespectively).

Effects on crater form: Studies of multi-ring basins on Mars and Callisto (9,10,11) indicate that the nature of impact-induced fracturing may change as the transient cavity depth approaches the lithospheric thickness. Transient cavities penetrating an elastic layer into an underlying viscous region can induce concentric lithospheric fracture to great distances beyond the crater rim (11). Such a process has been proposed for the formation of both the extensive ring scarp system around Valhalla on Callisto (10,11) and concentric canyons on Mars over 600 km outside the outer, Cordillera-equivalent scarp of the Hellas basin (9). For extremely low underlying viscosities, complete disruption of the lithosphere is predicted (10), thereby providing a mechanism for effectively masking or even destroying the largest terrestrial impact structures (8). Impact structures smaller than some critical size determined by the elastic layer thickness, however, should be unaffected by such modification. If a 5:1 transient cavity diameter/depth ratio is assumed, typical oceanic and continental lithospheres on Earth result in transitional crater diameters of 150-250 km and -300 km, respectively (8): over twice the size of the largest observed impacts on Earth. If the elastic layer instead reflects crustal thicknesses, then a maximum preserved diameter of 150-200 km (continental) and -50 km (oceanic) would be expected. These values would place the Sudbury and Vredefort structures near the limit of typical elastic behavior. The failure to identify larger craters in the terrestrial record might reflect ambiguous preserved signatures due to the disruption of these structures by extended lithospheric failure.

Reduction in elastic layer thickness to a value near that of the crustal thickness may occur in several ways. First, reduced lithospheric thicknesses would be expected in the archean due to increased radiogenic and convective heat fluxes (12). Second, regional tectonic activity like that of the Basin and Range Province in the western U.S. can sharply reduce lithospheric thickness. Finally, the elastic layer in the impact environment, due to shock effects, may not correspond directly to a flexurally or thermally defined lithosphere (10) and the compositional discontinuity beneath the crust might decouple the mantle response from a more elastic crustal response. Regardless, elastic behavior would be less likely in regions of high heat flow and thus the preferential identification of large craters in cratonic regions could reflect the reduced heat flows typical of these regions.

Effects on the mantle: Although material flow fields during transient cavity formation result in ballistic trajectories for most near-surface regions, a full target section is driven down at the center of the impact (2,7). In the case of an elastic half space, this disrupted section spreads across the base of the transient cavity (2). If viscous flow occurs beneath the impact, however, downward displacement of this section is partly accomodated by lateral flow in the mantle. This process transfers impact deformation from the lithospheric section to the mantle and may place shocked crustal material in the upper mantle or asthenosphere.

Viscous deformation of the mantle during the impact event, however, is necesary for such interaction of crustal material with the mantle. Such deformation is expected if the duration of impactor penetration exceeds the Maxwell time of the mantle (7). The Maxwell time is defined as the time required for viscous creep under stress to equal elastic strain (13); consequently, viscous behavior occurs when deformation times are greater than the Maxwell time. If impact shear stresses exceed 10 kilobars (10^3 MPa), strain rates in an olivine mantle range from 10^{-2} to 10^{-4} /s for mantle temperatures of 800-1000°C (14) and the mantle Maxwell time is on the order of 1-100 seconds (7).

Only large impactors which have such penetration times (eg. a 200 km diameter object impacting at 15 km/s has a penetration time of at least 15 s), therefore, are likely to induce such a viscous mantle response on the Earth.

The extent of viscous deformation beneath an impact also depends on the impact angle. Most impacts do not occur vertically, and systematic changes in the crater shape, energy partitioning and projectile fate have been documented as functions of impact angle (15,16). In particular, near-vertical impactors (>65° from the horizontal) can survive in the crater floor at 6 km/s, whereas oblique impactors will ricochet downrange due to shock and spallation effects (17). Penetration depths of the projectile and down-driven target section decrease significantly with decreasing impact angle; hence, near-vertical impacts are the most likely to place crustal/projectile compositions at mantle depths. While rare, such impacts are not improbable and should compose a distinct sub-set of the terrestrial impact record.

The requirement of long duration impact events for viscous deformation limits the extent to which this process will affect terrestrial history. Impacts of sufficient size (greater than some 200 km diameter) are mostly restricted to times before -3 Ga. Impact velocity, however, imposes a strong constraint on the consequences of such deformation. We have previously proposed (7) that such collisions on the Moon and Mars, where impact velocities range down to 5-6 km/s (18), may be able to bury coherent crustal sections at depth in the mantle. The range of impact velocities on Earth of 15-40 km/s, however, restricts such "subduction" events to larger bodies and enhances disruption of the displaced crust. The increased extent of vaporization and melting beneath the impactor should preclude preservation of a coherent crustal section and violent decompression and high shock vapor pressures could disperse this section (18). Nevertheless, the impact process still drives highly shocked crustal materials into the mantle with consequent crustal contamination of the mantle possible.

Such mantle contamination by impact can potentially influence mantle evolution in two ways. First, injection of crustal radiogenic elements into the mantle could locally affect the long-term thermal history; second, the introduction of crustal materials could affect mantle melting temperatures by injecting minor elements (eg. water, carbon, sulfur) into the mantle. If a disseminated crustal component is incorporated into mantle compositions after impact, such mixing (for mantle-crust ratios of 100:1) can double the local abundance of heat producing elements in the mantle. These added heat sources eventually can induce mantle partial melts some 100-500 Ma after the subduction event (7).

Conclusions: Very large terrestrial impacts (diameters greater than -200-300 km) will have transient cavity depths comparable in size to elastic plate thicknesses. The collapse of these transient cavities can produce extensive lithospheric fracture and may disrupt the lithosphere sufficiently to enhance degradation of the impact structure. Low-velocity, near-vertical impacts of this size range also may significantly contaminate the mantle with shocked crustal compositions. Unlike the more continuous subduction of oceanic plates observed on Earth, impacts are randomly located and episodic on a global scale. Hence only random and isolated regions of a planetary mantle can be modified by this process, which primarily affects radiogenic isotope concentrations and mantle melting.

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THE ACRAMAN IMPACT STRUCTURE, SOUTH AUSTRALIA; G.E. Williams, Department of Geology & Geophysics, University of Adelaide, GPO Box 498, Adelaide, South Australia 5001, Australia

The Acraman structure (1,2), located in the middle Proterozoic (1590 Ma) Gawler Range Volcanics on the Gawler Craton, South Australia (Fig. 1), is one of the largest known terrestrial impact structures. It is notable also as the likely source of a unique ejecta horizon of shock-deformed volcanic fragments within the late Proterozoic (~600 Ma) Bunyeroo Formation of the Adelaide Geosyncline located ~300 km east of the impact site (1,3-5) and within the correlative Rodda Beds of the Officer Basin ~450 km northwest of the impact site (6).

The structure comprises a central uplift area ~10-12 km across of intensely shattered and shock deformed Yardea Dacite, within an inner topographic depression 30-35 km in diameter that contains the Lake Acraman salina. Rocks of the central uplift area exhibit shatter cones and multiple sets of shock lamellae in quartz grains that indicate shock pressures of ~150 kbar. A near-circular "ring" of structurally controlled valleys and low-lying country occurs at 85-90 km diameter, and arcuate topographic features occur near Lake Gairdner to the east at ~150 km diameter. The area within the inner depression is a declared Geological Monument of the Geological Society of Australia.

Fig. 1. Gawler Ranges region, South Australia. 1, Lake Acraman within the inner depression; 2, Lake Gairdner; 3, 85-90 km ring structure.

The study of satellite images played a key role in my discovery of the Acraman structure. In 1979 I first thought that the near-circular shapes of Lake Acraman and the surrounding depression, as shown by Landsat images, might reflect a major impact structure. In 1980 I located intensely shattered and shock-deformed rocks of the Gawler Range Volcanics on low islands within Lake Acraman; the presence of shatter cones and shock lamellae in quartz grains in the shattered rocks strongly supported an impact origin. NOAA-AVHRR images of South Australia, available to me in the mid 1980s, revealed the full structure as a near circular inner feature with a conspicuous ring at 85-90 km diameter and arcuate features at ~150 km diameter.

A negative Bouguer gravity anomaly of 10-15 mgals and ~60-70 km across covers the inner depression. In addition, the inner depression has a subdued aeromagnetic signature, and a shallow, dipolar magnetic anomaly is located below the central uplift area.

The only known outcrop of meltrock in the Acraman structure is a variably altered apparent dyke <1 m thick that is flat-lying and exposed over a few square metres in the central uplift area. Palaeomagnetic study of the meltrock by Dr P.W. Schmidt (CSIRO Division of Exploration Geoscience, North Ryde, NSW) indicates a stable magnetisation and palaeomagnetic poles that are in close agreement with palaeomagnetic poles determined for the Bunyeroo Formation and which are quite distinct from Ordovician (Delamerian) overprint directions in South Australia. The palaeomagnetic data thus provide strong support for the correlation of the Acraman impact event and the ejecta horizon in the Bunyeroo Formation.

Apatite fission track analyses for shattered rocks from the central uplift area by Dr I.R. Duddy (Geotrack International, Melbourne, Australia) and estimated rates of erosion suggest
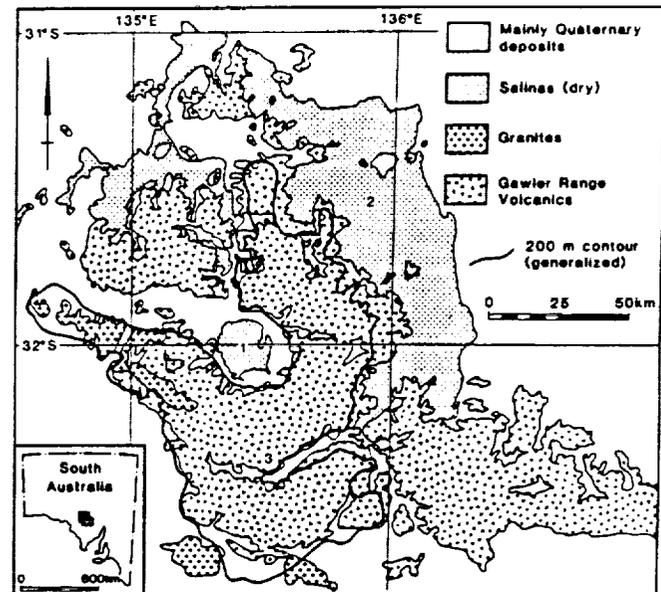
THE ACRAMAN IMPACT STRUCTURE, SOUTH AUSTRALIA: G.E. Williams

that at least 2-2.5 km thickness of overburden, and perhaps up to twice that thickness, has been eroded from the Gawler Craton since the impact event probably at ~600 Ma. The transient cavity therefore may have had a diameter of ~35-40 km, some 15% greater than that of the present inner topographic depression. The limits of the final collapse crater may be marked by the structural and topographic ring feature at 85-90 km diameter. The estimated dimensions of the uneroded Acraman structure (diameters of central uplift, "inner ring", and final collapse crater of ~10-12 km, 35-40 km, and 85-90 km, respectively) are comparable to those of numerous protobasins (large peak plus inner-ring basins) on Mercury and Mars (7).

Petrography, XRD and electron microprobe analyses show that the meltrock dyke consists primarily of slender, skeletal laths of albite up to ~350 x 20 μm and locally arranged in radial quench textures, set in a matrix of cloudy K-feldspar and finely intergrown quartz; numerous grains of iron oxides and a few small, partly resorbed xenoliths consisting mainly of quartz and K-feldspar also are present. Microprobe analyses indicate that both the Na- and K-feldspar phases have virtually pure, end-member compositions. XRD indicates that the matrix K-feldspar, which constitutes about 40% of the meltrock, is related to orthoclase. The matrix is interpreted as a devitrified glass. The Yardea Dacite from the impact site contains ~4-5% K_2O and the meltrock ~7.3% K_2O; similar enrichment of potassium in meltrock relative to basement rocks has been observed in other impact structures. The lack of anomalies for meteoritic and cosmogenic elements such as Ni, Cr, Co, Ir and other PGEs in the meltrock and shattered bedrock can be explained by the deep level of erosion of the Acraman structure.

The virtual end-member compositions of the feldspars in the meltrock indicate that they are not equilibrium products from a melt derived from the Gawler Range Volcanics, but are low-temperature (~25-150°C), authigenic phases that formed by diagenetic alteration of the meltrock, including albitisation and devitrification. As the meltrock evidently contains no primary K-bearing phase (feldspar or glass) that formed upon cooling immediately after the impact, it is unlikely to provide reliable K/Ar or $^{40}Ar/^{39}Ar$ ages of the impact event, but rather minimum ages that may record the time of diagenetic alteration of the K-bearing phases. Indeed, meltrock from very old impact structures may prove difficult to date reliably by the K/Ar and $^{40}Ar/^{39}Ar$ methods because the material dated would be from the exhumed, deeper levels of the structures (depths >2 km) and would probably have undergone diagenetic changes such as authigenic replacement of feldspars and devitrification of glass.

In summary, the Acraman structure displays numerous criteria for the identification of impact structures, namely: circular plan; ring structure; central uplift structure; negative gravity anomaly; subdued magnetic signature; intense brecciation of bedrock; presence of meltrock; shattercones in bedrock; and multiple sets of shock lamellae in quartz grains. Although meteorite fragments are not of course preserved, the Acraman structure is unique in evidently having part of its distal ejecta blanket preserved; this ejecta horizon displays distinct Ir-PGE anomalies (5). The continued study of the Acraman structure and the Bunyeroo ejecta horizon therefore provides a rare opportunity to ascertain the structural, geochemical and dynamic effects of a major terrestrial impact.

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ANNIHILATION OF LIFE BY VERY LARGE IMPACTS ON EARTH; K. Zahnle, NASA-Ames Research Center, and N. H. Sleep, Stanford University

The abnormal events recorded at the Cretaceous/Tertiary (K/T) boundary show that, even today, biospheric cratering can be important on Earth. Impacts were very much larger and more numerous on early Earth. Very large collisions, releasing 100 - 1000 times as much energy as the K/T impact, were taking place on the Moon at 3.8 Ga, forming basins such as Imbrium and Orientale. The Moon suffered at least ten such collisions between 3.8 and ~ 4.1 Ga[1]. Owing to its greater collisional cross-section, Earth would have accreted some 200 similar objects over the same period, and in all likelihood tens of objects much larger still. If the size of the largest impactors were commensurate with the larger number of smaller impactors, impacts on Earth at least 30 and probably more than 100 times greater than Imbrium are expected. To order of magnitude, the energy released in such an impact is enough to vaporize the oceans. In all likelihood impacts posed the greatest challenge to the survival of early life, with only the most protective niches, e.g, the mid-ocean ridge hydrothermal vents or deep aquifers being in any sense continuously habitable[2].

It takes some $\sim 5 \times 10^{34}$ ergs to evaporate the oceans. Assuming that something like a quarter of the impact's energy is spent evaporating water, this corresponds to the impact of a $\sim 10^{23}$g body, roughly the size of the large asteroids Vesta and Pallas. Something like half the energy of an ocean-vaporizing impact would initially be in the ejecta. At least a few impactor masses of ejecta are expected. Much or most of the ejecta would be very hot rock vapors created on impact; the rest would be melt droplets and grains that would soon be vaporized by later events, such as mixing with hotter vapors or atmospheric re-entry heating[3]. The resulting ~ 100 bars of rock vapor would displace any pre-existing atmosphere. It is through the rock vapor that much of the impact's energy is globally distributed. Fig. 1 begins at this point. The rock vapor atmosphere radiates upward to space and downward onto the ocean with an effective temperature of order $\sim 2000°K$. It takes a few months to radiate away the energy initially present in the rock vapor. In all, roughly half the energy should get absorbed by the ocean.

The last planet-sterilizing impact presents what is effectively a biological "event horizon", since none of the biochemical events that may have occurred before this impact bears on the origin of life as it exists now. Although no doubt dangerous, an ocean-vaporizing impact would not obviously wipe out life on Earth today. Frontier outposts several hundred meters below the surface in aquifers and oil fields would probably survive a minimal ocean-vaporizing impact. It is less clear that these ecosystems could repopulate the Earth, or that comparably evolved ecosystems even existed on early Earth. Cool, potentially survivable regions could exist within rock or sediment if (i) they were deep enough to be shielded from the heat at the surface, and (ii) the local geothermal temperature gradient were shallow enough. Ancient organisms may have hidden in marine sediments or deep aquifers, or been buried alive by tsunamis and gigantic landslides triggered by the impact.

[1]Taylor, S.R. (1982) *Planetary Science: A Lunar Perspective*, LPI
[2]Maher, K. and Stevenson, D.J. (1988) *Nature 331* 612-614; Sleep, N.H. *et al* (1989) *Nature 342* 139-142
[3]Sleep *et al, op. cit.*; Melosh, H.J. *et al* (1990) *Nature 343* 251-254
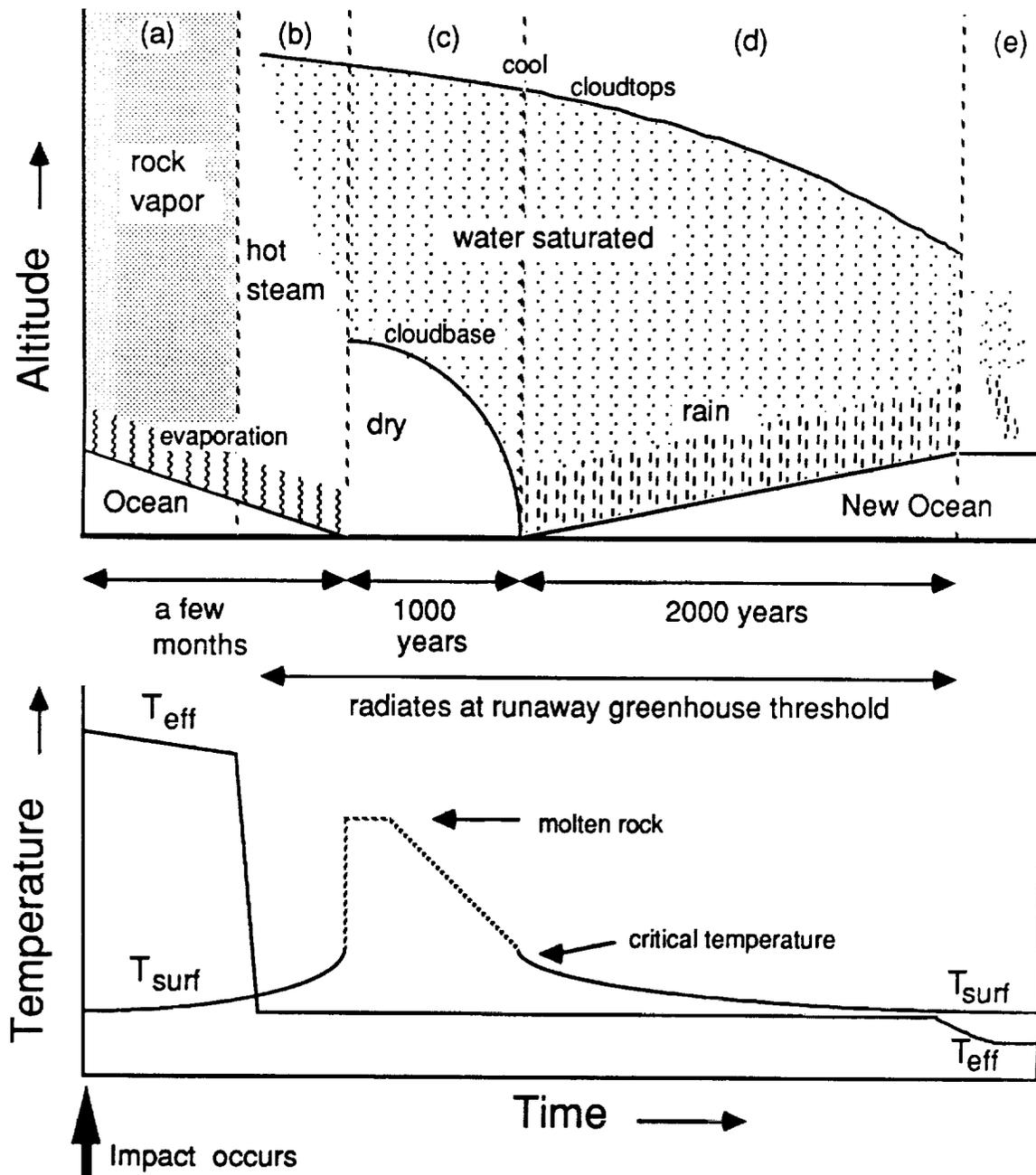
Fig. 1. An ocean-vaporizing impact. An impact on this scale produces about 100 atmospheres of rock vapor. (a) Thermal radiation from the rock vapor boils off the surface of the ocean. (b) Once the rock vapor has condensed the steam cools and forms clouds. (c,d) Thereafter cool cloudtops ensure that Earth cools slowly, with an effective radiating temperature of order 300K. In the minimal ocean-vaporizing impact the last brine pools are evaporated as the first rains fall. (e) Normalcy returns after some 3000 years.