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Annihilation of ecosystems by large asteroid impacts on the early Earth

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Large asteroid impacts produced globally lethal conditions by evaporating large volumes of ocean water on the early Earth. The Earth may have been continuously habitable by ecosystems that did not depend on photosynthesis as early as 4.44 Gyr BP (before present). Only a brief interval after 3.8 Gyr exists between the time when obligate photosynthetic organisms could continuously evolve and the time when the palaeontological record indicates highly evolved photosynthetic ecosystems.

THE early Earth was bombarded by large projectiles at the same time that large impact basins formed on the Moon. It is sometimes asserted (for example, ref. 1) that these impacts would have precluded the continued existence of life before 3.8 Gyr, but little has been done until recently to quantify the untoward effects of impacts on life. Maher and Stevenson[2] considered the possibility that frequent impacts could have prevented the origin of life, particularly for sites at the surface. They did so by comparing the time between impacts at a given location on the Earth's surface with the time required for biogenesis. In this paper, we concern ourselves with a related, yet inherently different problem, namely: When did the last planet-sterilizing impact occur? That is, we assume that life originated in some unspecified manner, and then consider what it would have taken to have wiped it out. This approach is better because the timescale for originating life is very poorly constrained, whereas early ecosystems with well defined organisms are broadly constrained by analogy with modern ones. In addition, extant organisms and the palaeontological record relate only to continuous evolution after the last impact to sterilize the planet.

We argue that the major way in which the planet can be sterilized is by evaporation of ocean water. This process affects some organisms more strongly than others, depending on how much water is evaporated and what type of ecosystem is envisioned. Certain classes of ecosystems may be able to withstand thermal perturbations lasting hundreds or thousands of years. We first discuss the flux of impacting objects, then the effects of impacts, and finally their effects on plausible early ecosystems.

Flux of impacting objects on the early Earth

The impact of a Mars-sized object, which is believed to have formed the Moon, would have either melted or vaporized much of the Earth[3]. Neither living cells nor complex organic molecules would have survived this blow. It is the flux of lesser later projectiles on the early Earth that is relevant to biology. No direct record of the earliest impacts exists on the Earth. By contrast, the lunar record is well preserved and well studied. Thus, the lunar crust is a reasonable starting point for our study. Our approach follows that in refs 4-7. We represent impact fluxes in terms of the equivalent thickness that the projectiles would add to a planet if all the material were retained. This measure is independent of planetary size so long as gravitational focusing and the statistics of the small number of largest objects can be neglected.

Geochemical estimates of flux. A minimum estimate of the material that hit the moon is obtained from the meteorite component retained in lunar samples. Rocks exposed in the lunar highlands, which are believed to be typical of the lunar upper crust, are breccias composed of pristine rocks, impact melts and still older breccias. It is generally agreed that Ir abundances indicate a meteoritic component of 1-2%. It is controversial whether Ni abundances indicate a comparable meteoritic component that lacks Ir (ref. 8) or an internal component that was reduced to metal in the crust[9]. Studies of lunar meteorites indicate that the excess nickel is local to the nearside landing sites and that the meteoritic component is between 1-2%[10]. The total thickness of lunar crust that is contaminated with siderophiles is not well constrained. Granulites believed to be deeply buried breccias as old as 4.26 Gyr (ref. 11) and 4-Gyr old near-surface breccias are both enriched in siderophiles, indicating that the meteoritic component is mixed in much of the upper crust. The megaregolith thickness is one estimate of the mixing depth. Spudis and Davis[12] give half the thickness of the lunar crust, 35 km. If the meteoritic component is between 1% and 4%, this implies a total meteoritic thickness between 0.35 km and 1.4 km, with 0.7 km being our preferred estimate.

The time of these impacts is also constrained. If a magma ocean existed on the Moon, the upper crust should have frozen from the top down while being stirred by impacts[7]. The date at which a mostly solid crust existed at any depth defines the start of the time interval for Ir and Ni retention in the lunar regolith. The oldest lunar rocks approach 4.5 Gyr and indicate that a crust existed early in the history of the planet[13]. The upper part of the lunar crust, ferroan anorthosite, solidified as early as 4.44 Gyr (ref. 14). The lunar crust was largely solid by 4.36 Gyr (the model age of KREEP (potassium rare-earth elements phosphorus) basalt), which is believed to be the date of the last solidification of the base of the crust[15]. Much of the accretion was over by 4.26 Gyr, the age of the oldest regolith sample.

Energy and mass of estimates of total projectiles. The lunar crust is so mildly churned that this fact by itself sets a useful upper limit on the flux of large projectiles. Large impacts were sufficiently rare that regional heterogeneities, primary stratification, and even local igneous bodies have not been obliterated[16,17]. Impacts similar to or larger than Imbrium seem to have been rare after the upper crust froze at 4.44 Gyr, as the crustal stratification remained intact over much of the lunar surface. In particular, impacts larger than Imbrium should have excavated mantle, but no lunar mantle samples have ever been found despite careful search. Similarly troctolite, an expected constituent of the lower crust which was excavated by Imbrium, is quite rare[18]. The total amount of material from impacts occurring since 4.44 Gyr should be less than the 1.88-km equivalent layer needed to saturate the surface with Imbrium-sized craters[4].

The size of the largest projectiles is relevant in this regard (Fig. 1). The energy of Orientale (diameter, 930 km), the youngest (3.8 Gyr) and best preserved of the basins, is obtained from the amount of buried heat by examining the thermal contraction of the centre of the basin[19]. If we assume that 25%

of the energy is deeply buried, the impact energy is between 4×10^{25} J and 3×10^{26} J with the middle of the range 1.2×10^{26} J being our preferred estimate. This corresponds to an object of mass 1.4×10^{18} with a diameter of 93 km (assuming a projectile density of 3,300 kg m^{-3} and a velocity of 13 km s^{-1}). The ejecta mass[20] of 2.4×10^{19} kg implies a ratio of ejected mass to projectile mass of 17. The largest confirmed basin, Imbrium, ejected 3.6×10^{19} kg (ref. 21) or 1% of the 35-km-thick megaregolith layer. The impactor mass was 2×10^{18} kg (ref. 6) Baldwin[4] gives an equivalent thickness of 0.2 km of accretion after 4.30 Gyr, equal to the mass of 11 Imbrium projectiles.

Extrapolating to the Earth. The Earth should have been hit with many more projectiles than the moon as it has more surface area and larger gravity. Statistically, the Earth is also likely to get hit with larger objects. For the impact velocities we assume below, the Moon is hit by $\frac{1}{24}$ of the objects and the Earth the remaining $\frac{23}{24}$. The probability that the moon is not hit by any of the 16 largest objects is over 0.5, because $(\frac{23}{24})^{16}>0.5$. Thus, if Imbrium is the largest lunar projectile after 4.30 Gyr, 16 larger objects are expected to have hit the Earth in the same period. The time and size of these impacts are difficult to constrain because of the small numbers involved. For example, the late impact of the Imbrium object on the Moon could be a statistical fluke. There is no convincing evidence for a 'spike' of late impacts between 3.8 Gyr and 3.9 Gyr and no obvious physical mechanism for producing them[6]. Conversely, it can be concluded that the number of huge objects that hit the Earth was not large.

Theoretical size-number distributions are one way to infer the size of the large objects that hit the Earth from the lunar record. For fragmentation, such mass distributions are power laws. For example, the cumulative number of objects with a mass greater than m is proportional to m^{1-q}. We can visualize the properties of such a distribution by 'binning' the objects in logarithmic intervals of mass. For $q=2$, there is equal mass in every bin. For $q=\frac{5}{3}$, there is equal surface area in each bin, and most of the mass is concentrated in the largest objects. The largest is expected to be $2-q$ of the total mass of the ensemble. With some algebra, it can be shown that the median fraction of the total mass hitting the moon is $(\frac{1}{23})^{1/1-q}$. The size distribution is maintained in a quasi-steady state by a fragmentation cascade. Theoretically, expected values of q range from $\frac{5}{3}$ for mild collisions to 2 for violent collisions[22].

The most relevant surviving class of objects for impacting the early Earth are asteroids with diameters between 130 and 260 km which are believed to be fragments of larger bodies. Such bodies would enter Earth-crossing orbits after the fragmentation event. The lower end of the range is comparable to the Imbrium object from which our extrapolation starts. For this size range, $q\approx2$ (refs 22, 23). If this distribution applies, the largest objects hitting the Earth need not have exceeded 260 km.

The larger gravity of the Earth increases the probability and energy of impacts[24], by a factor of 1.74 assuming an approach velocity of 13 km s^{-1}, which corresponds to an impact velocity of 13 km s^{-1} on the Moon and 17 km s^{-1} on the Earth. The approach velocity (weighted for probability of impact) of present Earth-crossing asteroids, 16.8 km s^{-1} (calculated from ref. 25, Table 8.5.1), is higher for large early asteroids by an unknown amount because the sample includes small asteroids that began as comets[26] and small asteroid fragments ejected by collisions[27].

Global sterilization by large impacts

We are concerned here with global effects of large impacts which would sterilize the entire planet. Boiling of the ocean by the heat of rock vapour produced by the impact is the most obvious such effect. Other global effects, including pressure changes in the ocean from large tsunami waves, fouling the ocean with meteorite debris and ejecta, and salinity changes as the ocean is boiled off and later as the water rains out have no obvious planet-sterilizing effects and are not considered further. The crater itself is of only regional importance. For example, according to the scaling relations compiled by Maher and Stevenson[2], an asteroid of 500-km diameter moving at 17 km s^{-1} would create a crater 1,500 km in diameter with 0.1-km-thick ejecta extending over a region 4,000 km in diameter. The ejecta at the rim of this crater would be 3 km thick.

Energy considerations. The estimates of the energy necessary to boil the ocean or the photic zone are insensitive to whether the final state is considered to be isothermal, adiabatic or on the vapour-saturation curve. For water that is initially at 0 °C, isobaric boiling requires 2.5×10^{6} J kg^{-1} at 1 bar and 2.1×10^{6} J kg^{-1} at the critical pressure. Thus, $\sim4\times10^{26}$ J of energy are needed to boil the 200-m-thick photic zone of the ocean, and 5×10^{27} J is needed to evaporate the entire ocean (1.4×10^{21} kg) and raise the temperature of the water vapour above the critical point. To further raise the surface temperature to the melting point of typical silicate rocks requires about half again as much energy.

An impact of an ocean-evaporating scale corresponds to a 440-km diameter (an object of mass 1.3×10^{20} kg)—roughly the size of the large asteroids Vesta and Pallas—hitting at 17 km s^{-1}. A 190-km-diameter (mass 1.1×10^{19} kg) object would evaporate the photic zone. We estimate that ~25% of the impactor energy goes into evaporating sea water, 25% is radiated to space, and 50% buried near the impact site. For impacts of this size, most of the material that leaves the crater is either melted or vaporized[24]. Rock vapour is produced directly by the impact and secondarily when fine particles of ejecta return to the atmosphere. Thus, the planet is quickly enveloped by about 100 bar of hot rock vapour and suspended droplets. The rock-vapour atmosphere radiates upwards to space and downwards onto the ocean with an effective temperature of 2,000 K. For an impact on an ocean-vaporizing scale, the radiative cooling time for rock vapour to condense is a few months. The precise radiating temperature is not too important, because roughly half of the rock vapour's energy is radiated to space, and roughly half is absorbed by the ocean.

To continue with the events following impact, the high opacity of sea water to infrared radiation concentrates boiling to a thin surface layer. Radiation from the rock vapour ablates the surface of the ocean, leaving the ocean depths cool. Convective mixing of the ocean is strongly inhibited by the temperature gradient. As water vapour is somewhat transparent to 2,000-K blackbody radiation, the oceans are not effectively shielded from the hot rock above, yet the water vapour is sufficiently opaque that it is quickly heated to the 2,000-K atmospheric temperature[28]. During this period the atmosphere is kept roughly isothermal around 2,000 K, the condensation temperature of silicates. Water clouds would not form under these conditions.

Once the rock vapour has condensed, a steady-state water-vapour atmosphere above a liquid ocean can radiate to space no faster than a certain threshold level that is only some 30% higher than Earth's present infrared flux[28]. This is precisely the 'runaway greenhouse' threshold often encountered in the comparative planetology of Earth and Venus[29]. This rate is determined by the transmission of infrared radiation through a moist, convective, H_2O-dominated atmosphere. The presence of clouds can only reduce the rate of thermal emission. Faster radiative cooling demands very high surface temperatures, ~1,500 K, to exploit the lower opacity of water vapour to visible and near-infrared radiation. The runaway greenhouse threshold also controls the maximum rate at which a water-vapour atmosphere can cool while water condenses.

For the minimal ocean-vaporizing impact, about half the ocean remains as liquid water at the time when the rock vapour has rained out. (About 300 m of rock raindrops would sit on the ocean floor.) At this point the atmosphere would consist of about half an ocean (140 bars) of hot (>1,500 K) steam. Shortly afterward, the top of the steam atmosphere cools sufficiently to form a moist, convective, optically thick upper layer. Thereafter, the planet radiates to space or below at the runaway greenhouse threshold, which is negligible compared with the rate at which energy is transferred to the ocean. In the minimal ocean-vaporizing impact, the last droplet of ocean is evaporated as the first droplet of rain reaches the ground. The surface temperature is then near the critical temperature, 647 K.

The lifetime of the steam atmosphere can be estimated by dividing the energy invested in steam by the effective cooling rate of 150 W m^{-2}, this being the difference between the runaway greenhouse threshold and total absorbed solar irradiation around 4 Gyr BP. The cooling time is on the order of 2,000 years for a minimal ocean-vaporizing impact. (The cooling rate is twice as fast if the clouds are sufficiently opaque that no solar radiation is absorbed; this difference is unimportant here.) An impact ~50% more energetic than the minimum ocean-vaporizing impact will leave a 1,500-K steam atmosphere when the ocean evaporates. This takes 3,000 years to cool. Cooling times for still larger impacts may not be greatly longer, because the runaway greenhouse threshold does not apply for very hot surfaces. But it is clear that extremely adverse conditions are expected for at least 3,000 years after any impact even modestly larger than the minimal ocean-vaporizing impact.

By contrast, adverse conditions in the atmosphere and the shallow ocean persist for about 300 years for an impact which evaporates only the photic zone. Bottom-dwelling photosynthetic organisms would be killed directly as the ocean boiled away above them. Floating photosynthetic organisms might, if they were lucky, be continually mixed downward as the ocean surface boiled. Survivors would, however, be killed when they were mixed back up to the surface and exposed to hot, fresh rainwater.

Discussion and uncertainties. Given that 25% of the impact energy is used to evaporate sea water, a total impact energy of 2×10^{28} J is needed to evaporate the entire ocean. An object of mass 1.3×10^{20} kg (440-km) hitting the Earth at 17 km s^{-1} would suffice. Ocean vaporization is plausible at much later times given a scale-invariant impactor distribution. Using the inference that the Imbrium projectile is $\frac{1}{11}$ of the mass colliding with the Moon after 4.3 Gyr to obtain $q = 1.91$, the largest object inferred to hit the Earth since 4.44 Gyr is expected to account for 9% of the total accreted mass, or 3×10^{20} kg. Between 3.8 and 4.3 Gyr, an object 31 times as massive as the Imbrium projectile, or 6×10^{19} kg, is also statistically expected (Fig. 1). Given the uncertainties inherent in such a scale-invariant distribution, the last ocean-vaporizing impact may have occurred as early as 4.44 Gyr or as late as 3.8 Gyr. Impact vaporization of the photic zone as late as 3.8 Gyr is probable, however, whatever the size distribution, as the larger fragmental asteroids would suffice. A long time interval between the last impact to vaporize the ocean and the last to have evaporated the photic zone is thus statistically likely.

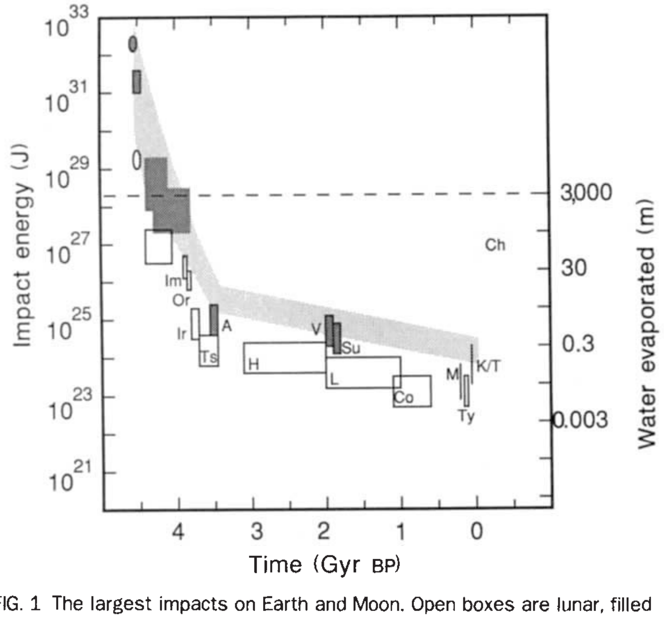

FIG. 1 The largest impacts on Earth and Moon. Open boxes are lunar, filled boxes terrestrial. Lunar craters are Tycho, Copernicus, Langrenus, Hausen, Tsiolkovski, Iridum, Orientale and Imbrium. Terrestrial events are the K/T impact, Manicougan, Sudbury, Vredevort and an impact energy corresponding to the thickness of Archaean spherule beds. Ovals are self energies of formation; the early box refers to a possible Moon-forming impact. Impact estimates between 3.8 and 4.4 Gyr are discussed in the text. The stippled region for Earth is inferred from these data. The depth of ocean vaporized by the impact is also given; the dashed line corresponds to an ocean-vaporizing impact. A possible but extremely unlikely collision with Chiron is placed safely in the future.

Survival of ecosystems

Biological systems require an energy source. The present biosphere receives its enthalpic buildup largely from solar energy; a second minor component in the deep sea is derived from the oxidation of reduced components at hydrothermal vents. The latter depends on oxidants being produced in the photic zone. It is likely that both these energy sources were available on the early Earth. The detailed nature of the earliest ecosystems is unknown because no sedimentary rocks older than 3.8 Gyr have been found. Life is unambiguously present at 3.56 Gyr in the Warrawoona Group in Australia; the organisms exhibited wall formation, some kind of motility, phototaxis and mat formation[30]. This complex and sophisticated set of properties should have required a substantial evolutionary history. In addition, carbon isotopes in the highly metamorphosed sediments of the 3.77-Gyr Isua Group of Greenland indicate that the ratio of organic carbon to carbonate in sediments was similar to the present implying that photosynthetic life was already abundant by Isua time[31].

Types of ecosystems. Even if individual organisms survive, it is necessary to reestablish a sustainable ecosystem immediately following the impact. The key issue becomes that of the primary producers.

Ecosystems where all the primary productivity is by obligate photosynthetic autotrophs have the primary producers in the top 200 m of the ocean where there is sufficient light. Impacts which sterilized this zone would destroy the global ecosystem and reset biogenesis.

Ecosystems where the primary producers are chemoautotrophs, especially those at hydrothermal vents, would have been much better protected. However, there is still a problem of supplying oxidized material. The present ecosystem at hydrothermal vents ultimately depends on photosynthesis for its oxidants[32]. A primitive system of that type would die out when oxidants were exhausted. An ecosystem where oxidants are produced inorganically, by oxidation of volcanic SO_2 (ref. 33), or by photolysis of dissolved ferrous iron[34], would survive.

A system with photosynthetic primary producers that could act as faculative anaerobic heterotrophs using organic matter and oxidants is also difficult to sterilize. Such reserves are easily exhausted in a well mixed system, such as modern oil spills or the mid-depths of the ancient ocean stirred up by an impact. The oxidant and reductant reservoirs are heterogeneously distributed on the modern Earth, however, and were probably also heterogeneously distributed on the ancient Earth. The structure of the ancient ecosystem would depend on whether sulphates (or sulphites) were present in the ocean.

If significant sulphate (or sulphite) existed, the situation in the global ocean was basically similar to modern closed basins in that excess organic matter existed in the sediments and excess mobile oxidant, the sulphate, existed in the sea water. Heterotrophs could then exist on the interface between the two reservoirs, that is on the sea floor and in very shallow sediments. In the absence of sulphate or sulphite, the main oxidant would be ferric iron, which is less mobile than organic matter and tends to sink to the sea floor. Excess ferric oxide would accumulate on the sea floor beneath regions of high organic productivity[35]. In either case, two reservoirs are not easily exhausted as they are utilized by only a small mass of organisms living at the sea-floor interface.

Additional interfaces within the sedimentary column might escape the surficial heating. The thermal diffusion distance, $\sqrt{\text{diffusivity time}}$, is about a few hundred metres. Deep habitable interfaces, for example, might exist between sedimentary calcium sulphate deposits (or sediments with dissolved sulphate) and sediments with organic matter. By analogy, modern oil-field bacteria exist at the interface of oil and sulphate (or oxygen)-rich water[36]. Although such deep ecosystems could persist for millions of years, some survivors would need to evolve photosynthesis to sustain life.

Implications. The window in time for the emergence of extant life is defined by a palaeontological considerations on the lower end and by geophysical and planetological considerations on the upper. The necessity for a long evolutionary pre-history to evolve the Warrawoona fossils and the Isua carbon isotopes suggests the existence of life as early as 3.8 Gyr. Here we have been concerned with defining the upper limit. This limit depends on what type of early ecosystem is envisioned and, hence, what one believes is the earliest life form. Did photosynthetic prokaryotes give rise to organisms that could utilize redox gradients or vice versa? Is it easier for life to originate in surficial environments or the deep ocean? Rather than attempting to resolve these biological issues, we examine the broader implications of our physical inferences.

If one postulates primitive photosynthetic biota as the sole primary producers, then annihilating events in the photic zone probably occurred as late as 3.8 Gyr, and our estimate for the time of the origin of continuous life converges on this date. The situation is less clear if one postulates early deep-marine life. (This situation is also less probable, as no evidence for such organisms exists in either the fossil record or in the evolutionary systematics of advanced life.) In this case, the upper limit may be as great as 4.44 Gyr if the maximum size of impactors was similar to fragmental asteroids or if the few largest objects in a scale-invariant distribution happened to hit the Earth early on. However, size and frequency of impacts should statistically increase as one moves further back in time. Thus, early organisms and ecosystems would have had to be increasingly sophisticated to survive. As this is counter to the usual course of evolution, the upper limit for continuous evolution would be around 4.0 Gyr, with considerable uncertainty. Conversely, a long continuity of life before 3.8 Gyr would imply occupation of the deep sea. □

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Intermediate states in the movement of transfer RNA in the ribosome

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Direct chemical 'footprinting' shows that translocation of transfer RNA occurs in two discrete steps. During the first step, which occurs spontaneously after the formation of the peptide bond, the acceptor end of tRNA moves relative to the large ribosomal subunit resulting in 'hybrid states' of binding. During the second step, which is promoted by elongation factor EF-G, the anticodon end of tRNA, along with the messenger RNA, moves relative to the small ribosomal subunit.

IN its simplest conceptual form, the mechanism of translation[1] must account for three fundamental events: codon-anticodon recognition, peptide-bond formation, and movement of tRNA and mRNA relative to the ribosome. The classical model[2] proposed that the ribosome contains two binding sites for tRNA, usually called the A (aminoacyl, or acceptor) site and the P (peptidyl, or donor) site. The A site binds the incoming aminoacyl tRNA (delivered to the ribosome as a tRNA–EF-Tu-GTP ternary complex), whose anticodon must read the codon in the mRNA presented in the A site. A peptide bond is then formed between the C-terminal carbonyl group of the nascent polypeptide chain, attached to a P-site-bound tRNA, and the α-amino group of the A-site-bound aminoacyl tRNA. This results in growth, by one amino-acid residue, of the polypeptide