Impact delivery and erosion of planetary oceans in the early inner Solar System

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The terrestrial planets may have acquired oceans of water (and other surface volatiles) as a late-accreting veneer from impacts of comets and carbonaceous asteroids during the period of heavy bombardment 4.5 to 3.5 Gyr ago. On any given body, the efficiency of this mechanism depended on a competition between impact delivery of new volatiles and impact erosion of those already present. For the larger worlds of the inner Solar System, this competition strongly favoured the net accumulation of planetary oceans.

BOTH carbonaceous asteroids[1] and comets[2] may have been important sources for Earth's surface volatile inventory, and in particular for its oceans. A long-standing objection to this hypothesis has been that large impacts might erode as much or more of the terrestrial volatile inventory as they deliver[3]. Analytical fits to the lunar cratering record, appropriately scaled to Earth's larger gravitational cross-section, allow estimates of the terrestrial bombardment history. By convolving these estimates with quantitative models for impact erosion of planetary atmospheres[4] and condensed oceans, I demonstrate here that Earth probably accreted a net (after impact erosion) amount of water corresponding to $\gtrsim$0.2–0.7 ocean masses during the period of heavy bombardment 4.5 to 3.5 Gyr BP (before present) (where the terrestrial ocean mass M_{oceans} is 1.4×10^{21} kg H_2O). Venus should have collected approximately as much water as Earth, whereas Mars, more subject to impact erosion, should have accreted an ocean equivalent to a layer ~10–100 metres deep distributed over the planet.

Ironically, the greatest uncertainty in these results arises from the part of the calculation that is based most strongly on empirical data. Differing lunar crater counts, as well as disparate acceptable fits to even the same lunar cratering data sets, lead to cumulative bombardment history models that differ by $\sim10^3$ when extrapolated back to 4.5 Gyr BP. (Lunar geochemical data do allow a firm lower limit to be placed on total incident impactor mass, however, which yields the 0.2–0.7 M_{oceans} cited above.) I also conclude that calculations of the impact frustration of the origins of life[5,6] that rely on such bombardment models may suffer from correspondingly great uncertainties.

Uncertainties in the lunar cratering record

Attempts to estimate the impact environment of early Earth typically begin with an analytical fit to the lunar cratering record. One such fit is that found by Melosh and Vickery[4] in their recent work on impact erosion of atmospheres. They give

$$N(>4\text{ km}, t) = 2.68\times10^{-5}[t + 4.57\times10^{-7}(e^{t/\tau_a} - 1)]\quad \text{km}^{-2} \quad (1a)$$

as an analytical fit to empirical data (from the Basaltic Volcanism Study Project (BVSP, ref. 7, Table 8.4.2)) relating surface density of lunar craters with diameters $D > 4$ km, $N(>4\text{ km})$, to the radiogenic age t of the cratered surface. Melosh and Vickery chose the time constant $\tau_a = 220$ Myr. The BVSP data, and the fit given by equation (1a), are shown in Fig. 1.

It is not clear that $\tau_a = 220$ Myr is the correct time constant for the decay of the early impactor flux. Other authors have made different choices. For example, Maher and Stevenson[5] modelled the impactor flux with $\tau = 70$ Myr, whereas Neukum[8] favours $\tau = 144$ Myr (corresponding to a 100-Myr half-life). Fundamentally, imprecision is inevitable because, as has been emphasized[9-11], the impactor flux cannot actually have decayed at a constant rate. Rather, the flux must initially have been dominated by objects in nearly circular Earth-like orbits (for which $\tau \sim 20$ Myr); as this population was swiftly swept up, different, more slowly-decaying populations (with $\tau \sim 100$–300 Myr, such as comets or Mars-exchanged asteroids) would have become predominant. Thus fitting a decay 'constant' to the lunar data is a crude approximation at best. Nevertheless, the actual data (~15 points) may not justify a more sophisticated treatment (see, however, ref. 11, for such an attempt).

To demonstrate the range of decay rates permitted by the BVSP data, I have fit these data, using two-dimensional χ^2 minimization, with an equation of the form (1a) using $\tau =$ 144 Myr. I have chosen this decay constant because a 100-Myr half-life has been demonstrated for the decrease of the primordial comet flux through the inner Solar System by independent numerical simulations of the formation of Uranus and Neptune[12,13]. My fit, shown in Fig. 1 as the solid curve, is given by

$$N(>4\text{ km}, t) = 3.5\times10^{-5}[t + 2.3\times10^{-11}(e^{t/\tau_b} - 1)]\quad \text{km}^{-2} \quad (1b)$$

where $\tau_b = 144$ Myr. Both equations (1a) and (1b) provide moderately good fits to the BVSP data, with χ^2 values of 8.06 and 10.6, respectively (for $15 - 2 = 13$ degrees of freedom). Equation (1a) is formally a better fit by ~25%, but this comparison is of little significance. First, as just discussed, fitting any single decay 'constant' to the cratering flux is a procrustean exercise. Second, the largest discrepancy between the two fits occurs for the point at ~4.4 Gyr in Fig. 1. There is, however, no true radiogenic data for this point; rather its age is simply an estimate[7]. Furthermore, if the lunar uplands are saturation cratered (see below), the crater density plotted for this point (which is an uplands value) could be a considerable underestimate.

Recent work on the impact frustration of the origins of life[5,6] uses a cumulative lunar crater density given by

$$N(>4\text{ km}, t) = 1.4\times10^{-5}[t + 5.6\times10^{-23}(e^{t/\tau_c} - 1)]\quad \text{km}^{-2} \quad (1c)$$

with $\tau_c = 70$ Myr. Equation (1c) provides a reasonable fit to the lunar cratering data (for $D > 20$ km) given by Wilhelms (ref. 14, Fig. 6.49). Maher and Stevenson[5] adopt a number-diameter scaling law $N \propto D^{-1.8}$ for lunar cratering, which they consider to hold at least to diameters as small as $D = 1$ km. In equation (1c), I have used this scaling law to convert Wilhelm's data (and their fit) to $D = 4$ km so that it may be compared directly (Fig. 1, dotted line) with the BVSP results. (Melosh and Vickery[4] implicitly adopt a -1.8 scaling as well, in their conversion from cumulative crater density to heavy-bombardment flux.) This conversion is of course perfectly consistent with Maher and Stevenson's model, but not quite with Wilhelm's data, which are based on combining crater counts at $D > 20$ km and $D >$ 1 km using an empirical law (derived from Imbrium basin counts) that is not consistent with -1.8 scaling. The $D > 1$ km counts are drawn from Neukum[8,14], whose small-crater counting methodology is incompatible with that adopted by BVSP[7]. It is not surprising that the results given by equation (1c) differ so

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markedly from those of equation (1*a*,*b*). Each model gives similar results for ages ~3.8 Gyr because the data at these dates are in rough agreement. (It is the requirement that the models approximately agree at ~3.8 Gyr, despite differing time constants, that leads to the vastly different coefficients, 10^{-7}–10^{-23}, in equation (1*a*)–(1*c*).) At dates ≳3.8 Gyr, however, the models diverge: extrapolation back to 4.5 Gyr gives cumulative crater densities that differ by $\sim 10^3$.

Another question that arises in comparing the BVSP and Wilhelms data sets is the role of secondary craters. BVSP generally include only craters with $D > 2.8$ km, to minimize contamination by secondaries. Basin secondaries as large as $D \sim 20$–30 km may, however, substantially contaminate crater counts in the lunar highlands[7,15]. Nevertheless, the slopes of the crater diameter distributions used by BVSP for the oldest lunar surfaces show no steepening at small *D*, indicating either that few secondaries contribute to these counts or that these surfaces are saturation cratered.

Early terrestrial impact environment

Given equation (1*a*, *b*, *c*) the total mass incident on Earth during heavy bombardment can be calculated. *N* in equation (1) scales as $D^{-1.8}$; this crater diameter scaling can be converted to impactor mass scaling using a mass-diameter equation. Following Melosh and Vickery[4], I use Melosh's[16] version of the Schmidt-Housen[17] relation, assuming an average impact angle of 45° and a crater collapse factor c_f of 30% (ref. 18)

$$m_p = (0.11)\rho_t^{19/15}\rho_p^{-1/4}g^{21/25}v_p^{-5/3} \times (1.3/c_f)^{19/5}(\sin 45°/\sin\theta)^{5/3}D^{19/5} \quad (2)$$

Here m_p is the mass of the projectile in kg, $\rho_t = 2{,}900$ kg m^{-3} is the lunar crustal density[19], $\rho_p = 2{,}500$ kg m^{-3} is the density of a typical impacting asteroid[20], $g = 1.61$ m s^{-2} is the gravitational acceleration at the lunar surface, $v_p = 1.3\times10^4$ m s^{-1} is a typical impact velocity of asteroids with the Moon (discussed below), *D* is measured in metres and θ is the impact angle. Melosh and Vickery[4] chose $c_f = 1.25$ which would increase m_p by ~16% over the values I obtain. Inserting equation (2) into the implicit $(4\text{ km}/D)^{-1.8}$ scaling in equation (1) yields three equations for the cumulative number, $n(>m, t)$, of impactors with mass $>m$ that have been incident on a lunar surface of age *t*. The total mass, $M(t)$, incident in impactors with masses in the range m_1 to m_2 is then

$$M(t) = \int_{m_2}^{m_1} m[\partial n(>m, t)/\partial m]\,dm \quad (3)$$

which yields

$$M(t) = 0.76[t + 4.57\times10^{-7}(e^{t/\tau_a} - 1)]m_2^{1-b} \quad \text{kg}^b\text{ km}^{-2} \quad (4a)$$

$$M(t) = 0.99[t + 2.3\times10^{-11}(e^{t/\tau_b} - 1)]m_2^{1-b} \quad \text{kg}^b\text{ km}^{-2} \quad (4b)$$

$$M(t) = 0.40[t + 5.6\times10^{-23}(e^{t/\tau_c} - 1)]m_2^{1-b} \quad \text{kg}^b\text{ km}^{-2} \quad (4c)$$

for equation (1*a*)–(1*c*), where $b = (1.8/3.8) \approx 0.47$, *t* is in Gyr and I assume $m_2 \gg m_1$.

What is the appropriate value in equation (4) for m_2? An *N*–*D* power law for lunar cratering is well obeyed in both the lunar frontside highlands and heavily cratered uplands (that is, the oldest lunar surfaces) for histogram bins in *D* up to $D = 512\sqrt{2}$ km = 724 km (ref. 7, Figs 23, 26, section 8.11). *D* = 724 km in equation (2) corresponds to $m_p = 1.5\times10^{18}$ kg, which I use for m_2 in equation (4). It is true that larger basins exist on the Moon (see, for example, ref. 14, Table 6.4), but the statistics of the formation of such basins as a function of time are problematic. Thus the calculation of $M(t)$ in equation (4) is a conservative underestimate. Because $M(t) \sim D^2$, the effect of a different choice for the largest basin included in equation (4) is easily deduced.

To calculate the total mass incident on Earth since $t = 4.5$ Gyr BP, I multiply $M(t)$ by the lunar surface area and scale to the larger gravitational cross-section of Earth. (4.5 Gyr has been chosen because the possible Moon-forming impact of a Mars-sized object with Earth is typically dated[16] at ~4.5 Gyr and volatiles previously accreted by the outer layers of Earth would probably be lost in this event. The oldest lunar rocks have ages consistent with crystallization 4.5–4.6 Gyr BP, ref. 21.) A body's gravitational cross-section σ is $\pi r^2[1 + v_{esc}^2/v_\infty^2]$ where *r* is the body's physical radius, v_{esc} is its escape velocity and v_∞ is the relative velocity at infinity of the approaching impactor. As discussed below, I use 17 km s^{-1} as a typical asteroidal impact velocity with Earth, corresponding to $v_\infty \approx 13$ km s$^{-1} \approx v_p$. For $v_\infty = 13$ km s^{-1} the gravitational cross-section of the Earth is ~23 times bigger than that of the Moon. But, ~4.5 Gyr BP, the Moon was much closer to Earth than now. In this case, gravitational focusing by Earth leads to both an increased lunar gravitational cross-section (v_{esc}^2 is replaced with the sum of the squares of the lunar escape velocity and the escape velocity of Earth at the Moon's orbital radius, which decreases Σ, the terrestrial/lunar gravitational cross-section ratio) and a larger average lunar impact velocity (v_p in equation (2)). Dynamical calculations of the evolution of the lunar inclination may exclude accretion of the Moon in Earth's equatorial plane within $\sim 10R_\oplus$ (Earth radii)[22]. At $10R_\oplus$, the effects of terrestrial gravitational focusing on both Σ and v_p are <10%, negligible compared to other uncertainties in the problem. If the Moon did in fact form near the terrestrial Roche limit, $\sim 3R_\oplus$, its orbital evolution out to $\sim 10R_\oplus$ was almost certainly so rapid ($<10^{-5}$ of the time required to evolve from 10 to $60R_\oplus$, in a model with constant terrestrial specific tidal dissipation *Q*—see, for example, ref. 23, equation (8)) that $t \approx 4.5$ Gyr may be used as the time when the Moon was at $10R_\oplus$.

I find that the total mass incident on Earth since 4.5 Gyr BP is $M_{tot} = 9.9\times10^{20}$ kg, 3.2×10^{21} kg or 6.9×10^{23} kg, respectively,

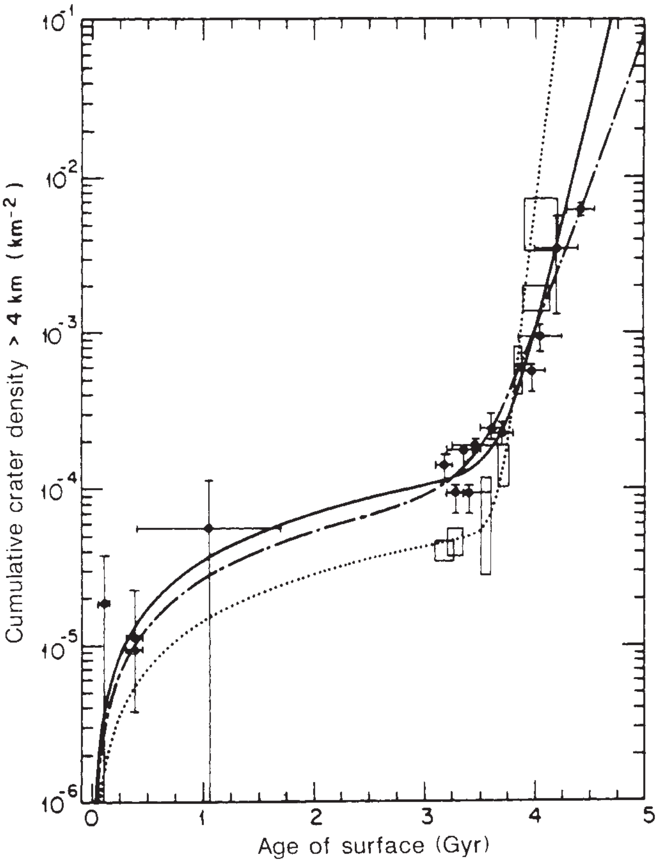

FIG. 1 Cumulative lunar crater density as a function of surface age, for the data in ref. 14 (boxes) and ref. 7 (crosses), fit by analytical formulae with 70 Myr (dotted line; ref. 5), 144 Myr (solid line) and 220 Myr (dot-dashed line, ref. 4), decay constants.

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depending on the version of equation (4) used. This range is similar to the lower limit ($7.6\times10^{20}-2.3\times10^{23}$ kg) found by summing the masses implied by all extant lunar craters and basins[2]. By comparison, Earth's mass $M_{\oplus}$ is 5.98×10^{24} kg, and the terrestrial ocean mass M_{oceans} is 1.4×10^{21} kg H_2O.

These results may also be compared with a recent determination by Sleep *et al.*[24]. They use iridium and nickel abundances of lunar highland rocks, together with estimates of the mixing depth of the meteoritic component of the lunar crust, to calculate a lower limit for the total amount of material striking the Moon since the solidification of the lunar crust ~4.4 Gyr BP. They conclude that the Moon has accreted an equivalent meteoritic thickness of 0.7 km, with an estimated uncertainty of a factor of two. Scaling to Earth with the parameters used here, and multiplying by a factor ~exp[(4.5–4.4) Gyr/100 Myr], I obtain $M_{tot}=4.8\times10^{21}$ kg as a lower limit for the total mass incident on Earth since 4.5 Gyr BP. Therefore the value $M_{tot}=1.0\times10^{21}$ kg, derived from the 220-Myr fit to the lunar cratering record, is evidently an underestimate of the total impact flux. This conclusion is even stronger when one considers impact models in which the Moon would have lost much material in the ejecta[25] and vapour plumes[4] resulting from high-velocity impacts. Although some of this material would have been trapped in Earth orbit and re-accreted by the Moon, much may have been lost from the Earth–Moon system altogether, because even as close as $10R_{\oplus}$, Earth's escape velocity is only 3.5 km s^{-1}, comparable to the Moon's (2.4 km s^{-1}).

Frustration of the origin of life

Certain conclusions of recent work on the impact frustration of the origin of life may be undermined by the uncertainties in the lunar (and hence, terrestrial) early impact environments described above. Maher and Stevenson[5] have compared proposed timescales (10^5–10^7 yr) for the origin of life to timescales for globally sterilizing impacts, and argued that life could not have originated before 4.2 Gyr BP. Oberbeck and Fogleman[6] corrected an error in the calculation and found that life could first have originated between 3.7 and 4.0 Gyr BP. Both these investigations, however, rely on the cratering time constant $\tau_c=70$ Myr. Figure 1 indicates that their conclusions may depend strongly on this choice of τ.

$^{12}C/^{13}C$ isotope ratios in 3.8-Gyr Isua metasediments[26] suggest that life originated before 3.8 Gyr BP. It is clear from Fig. 1 that, for times $\leq$3.9 Gyr BP, all three fits to the lunar cratering record provide approximately equal fluxes and hence must give about the same results for impact frustration (V. Oberbeck, personal communication). For times earlier than ~3.9 Gyr, however, conclusions drawn from the different models dramatically diverge. For example, at 4 Gyr, equation (1*c*) gives a cumulative cratering flux N_c which is about five times greater than that given by equation (1*a*), N_a. Timescales for devastating impacts will be proportional not to cumulative fluxes, but to differential fluxes. The relevant ratio is therefore $(\partial N_c/\partial t)/(\partial N_a/\partial t)\approx 5(\tau_a/\tau_c)\approx 16$. That is, the frequency of impacts at 4 Gyr for a $\tau_c=70$ Myr model is about sixteen times greater than that for a $\tau_a=220$ Myr model. The frequency of impacts sufficient to frustrate the origin of life at 4 Gyr in the τ_c model is only achieved in the τ_a model at a time Δt before 4 Gyr, where Δt is given by $\exp(\Delta t/220\text{ Myr})\approx16$, or $\Delta t\approx610$ Myr—which is before terrestrial accretion occurred. That is, frustration of life at 4 Gyr in the τ_c model implies there is no time in the τ_a model at which frustration occurs. These uncertainties emphasize the advantage of impact-frustration models that calculate the annihilation of assumed-extant ecosystems[24], or give as a result (rather than assume) a timescale for the origin of life[27].

Impact erosion of the hydrosphere

Melosh and Vickery[4] find that large impacts may cause atmosphere erosion when two criteria are met. The first is that the impactor must strike the planet at a velocity high enough for a vapour plume to form and expand at a speed $>v_{esc}$. Second, the mass of the plume must exceed the air mass above the plane tangent to the impact. In effect, these criteria are a momentum balance requirement. Here I consider an impactor that satisfies these two criteria to be entirely lost as an escaping vapour plume from the target planet. Not only will that impactor's volatiles not contribute to the target world, but atmospheric mass will be lost as well. As shown below for the case of the oceans, those volatiles present in a condensed state before an eroding impact will be largely immune to such erosion, resulting in the removal of some planetary volatiles (such as noble gases, nitrogen and carbon as carbon dioxide) and leaving others relatively unaffected.

Melosh and Vickery[4] found that the threshold impact velocity v_{min} for most of a vapour plume to exceed v_{esc} was $v_{min}=\sqrt{4(v_{esc}^2+2H_{vap})}$, where H_{vap} is the vaporization energy, assumed to be 13 MJ kg^{-1} for silicates and 3 MJ kg^{-1} for ice. For the Moon, $v_{min}\approx11$ km s^{-1} for silicate projectiles (asteroids) and $v_{min}\approx7$ km s^{-1} for ice impactors (comets), so that virtually all vapour plumes resulting from cometary impacts (for comets of any mass, as there was no atmosphere to be overcome), and most resulting from asteroidal impacts, would be lost. This helps to explain why the Moon, although battered by carbonaceous asteroids and comets throughout its history, remains so volatile poor. For Earth, however, $v_{min}\approx25$ km s^{-1} for asteroids, and $v_{min}\approx23$ km s^{-1} for comets. These threshold values are analytical approximations to numerical work in preparation by Melosh and Vickery (personal communication); my results, however, do not depend strongly on the exact values chosen.

Short period (SP) comets have $v_{rms}=28.9$ km s^{-1} (ref. 28), which has led to the suggestion[3] that 'the idea that the Earth owes its oceans to cometary bombardment long ago may have to be abandoned'. Root-mean-square calculations of most-probable impact velocities assign greatest weight to those objects with exceptionally high values of v_{∞}, however, skewing the velocity distribution towards high velocities. In fact, many SP comets will hit Earth with velocities <23 km s^{-1}. I have used Weissman's[28] compilation of SP-comet velocities and collision probabilities to calculate the percentage of comet–Earth collisions that occur at a given velocity (Fig. 2). The statistics are poor, but Fig. 2 indicates that a small number of high-inclination and retrograde comets pull v_{rms} towards high values. In fact,

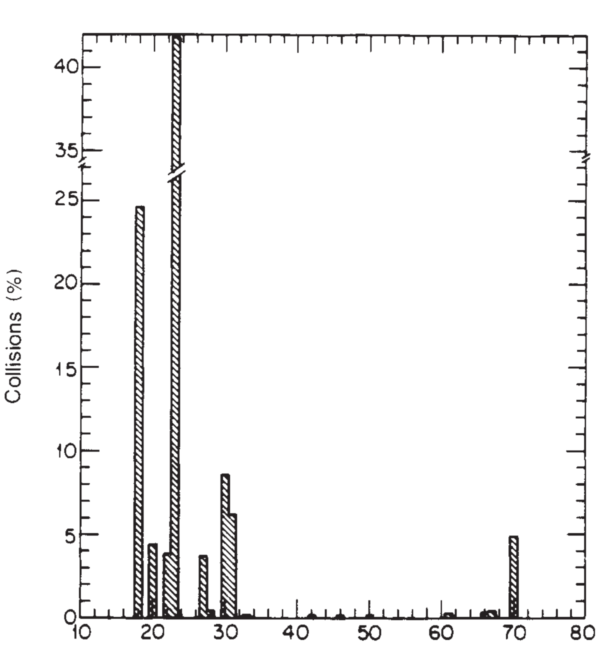

FIG. 2 The percentage of short-period-comet collisions with Earth as a function of impact velocity. Data from ref. 28.

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~30% of comet–Earth collisions occur at or below 20 km s^{-1} and ~45% occur just at ~22–23 km s^{-1} (the critical velocity for escape of the resulting vapour plume). Assuming that half of the latter collisions cause atmosphere erosion, there remains a full ~50% of SP-comet collisions with Earth that do not create vapour plumes with sufficient velocity to reach escape speed. Because models of the early cometary bombardment of the inner Solar System[12,13] indicate that the flux of comets scattered directly from the Uranus–Neptune region (that is, following SP-like orbits) dominated (by several orders of magnitude) the flux of those (long period) comets first scattered out to the Oort cloud, I consider that ~50% of the early comet flux incident on Earth had velocities low enough for their volatiles to have been retained. An analogous calculation using data for Earth-crossing asteroids (ref. 7, Table 8.5.1) shows that ~90% of asteroidal collisions with Earth occur at velocities <25 km s^{-1}. Yet these data also give a comet-like r.m.s. impact velocity for asteroids of ~25 km s^{-1}. Again, a small number of high-inclination objects skew the r.m.s. value. In fact, >50% of the terrestrial collisions of the Earth-crossing asteroids in this compilation will occur at velocities <17 km s^{-1}, the value chosen as 'typical' in the calculations above.

To determine the total mass of water accreted by Earth because of asteroidal and cometary impacts, I combine the calculation of M_{tot} above with an estimate of the division of M_{tot} into comet and asteroid fractions, with an estimate of the average water content of these bodies, and with the results of impact erosion in cometary and asteroidal collisions. The fraction α of the total ancient incident flux of comets is ultimately an unknown, free parameter. I have discussed elsewhere the variety of data pertaining to this question and argued that none is conclusive[2]. It seems likely that $\alpha \lesssim 10$–20%, but even this conclusion may be questioned. In the following discussion, I let α range from 10^{-3} to 1, and find that my results vary by less than a factor of four over this range.

Comets are ~50% water ice[29]. It has been estimated[20], on the basis of photometric observations of Earth-crossing asteroids, that the asteroidal flux at Earth is an equal mix of S and C types. Only types C1 and C2 carbonaceous chondrite meteorites match the spectra of C-type asteroids[20]; I assume C-type asteroids to be an equal mix of these. C1 and C2 meteorites are ~20% and 13% water by mass, respectively[30]. I assume that S-type asteroids contain no water.

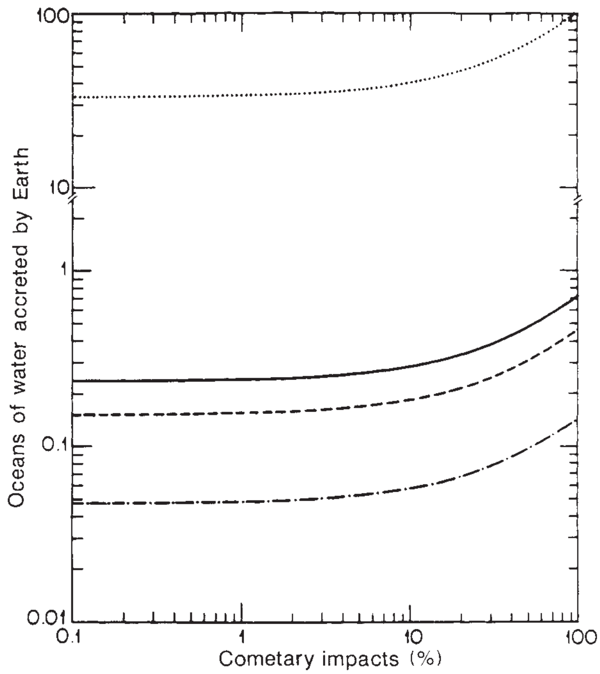

FIG. 3 Oceans of water accreted by Earth (in units of $M_{oceans} = 1.4 \times 10^{21}$ kg), as a function of the cometary percentage of impactor mass. Dot-dashed curve from equation (4*a*), dashed from equation (4*b*), dotted from equation (4*c*) and the solid curve is derived from ref. 24.

Finally, could impactors erode a significant quantity of condensed water from Earth's surface? Consider an impactor with velocity and mass sufficient to cause atmosphere erosion. For a 1-bar terrestrial atmosphere, such an impactor has a minimum mass m_* of 3.5×10^{15} kg, or a diameter $2r_p \approx 14$ km (ref. 4, equation (2)). The primordial terrestrial ocean would have accumulated its present surface-averaged depth $d_{oc} \approx 3$ km; the following calculation maximizes impact erosion of this ocean by assigning it d_{oc} since 4.5 Gyr BP. An eroding impactor of diameter $\gtrsim 2r_p$ deposits nearly all its impact energy at a depth $\sim 2r_p$; that is, it penetrates a 3-km ocean and continues many km into the sea bed[16]. The ocean water in the path of the impactor is rammed into the sea bed and severely shocked; this cylinder of water will enter the escaping vapour plume. For hypervelocity impact, the shock pressure in the ocean around this cylinder falls off as $\sim r^{-3}$ (ref. 16), so that little more water is sufficiently shocked to contribute to the escaping vapour plume (H. J. Melosh, personal communication)—of course, much more water is vaporized, but not lost from the planet.

Thus, the mass of ocean water eroded by an impactor of mass m, and diameter $2r_p$ is $M(m) = \pi r_p^2 d_{oc} \rho_{oc} = \pi(3/4\pi\rho_p)^{2/3} m^{2/3} d_{oc}\rho_{oc}$, where ρ_{oc} and ρ_p are the densities of the ocean and impactor, respectively. The total mass $M_{lost}(t)$ of condensed ocean eroded by impactors of mass $> m_*$ is then

$$M_{lost}(t) = \int_{m_2}^{m_*} M(m)[\partial n(>m, t)/\partial m]\,dm \quad (5)$$

Performing this integration, and taking into account that only ~10% of asteroids and ~50% of comets have v_{min} large enough to cause erosion, I find that $\leqslant 5\%$ and $\leqslant 15\%$ of the water delivered by asteroids and comets, respectively, is subsequently eroded by incorporation of condensed water into vapour plumes.

The fractional volume of atmosphere removed by an eroding impact is $\sim 3 \times 10^{-4}$. With $\sim 10^{-5} M_{oceans}$ residing in the present atmosphere[31], a calculation analogous to that above shows that loss by erosion of vapour-phase water is negligible. This result holds even in hypothesized 10-bar CO_2 early atmospheres, with ~0.5-bar H_2O in the vapour phase[32]. Finally, certain earlier models[33] of impact erosion indicated that erosion could not occur for worlds with $v_{esc} \gtrsim 10$ km s^{-1}; in this case Earth would have retained all incident volatiles.

Volatile inventories in the inner Solar System

Combining all of these considerations, the total amount of water accreted by Earth because of impacts over the past 4.5 Gyr is given by $M_{H_2O} = [(1-\alpha)(0.95)(0.9)(0.5)(0.17) + \alpha(0.85)(0.5) \times (0.5)]M_{tot}$, where the first term is the asteroidal contribution (5% of condensed oceans eroded, 90% of impactors accreted, 50% type C, which are ~17% water by mass), and the second is cometary (15% of oceans eroded, 50% accretion, 50% water by mass). M_{H_2O} scales erosion appropriately for the case $d_{oc} >$ 3 km (as for equation (4*c*)). M_{H_2O} is plotted as a function of α in Fig. 3, for M_{tot} calculated from both equation (4) and the empirical determination of Sleep *et al.*[24]. Because this latter determination represents a lower limit, I conclude from Fig. 3 that Earth accreted a net (after impact erosion) amount of water $\gtrsim 0.2$–$0.7 M_{oceans}$ during the period of heavy bombardment.

Even taking impact erosion into account, the cratering decay constant $\tau_c = 70$ Myr requires Earth to have accreted ~30–100M_{oceans} of water since 4.5 Gyr BP. Barring creative, and *ad hoc*, mechanisms to bury this much water in Earth's mantle, this result indicates either that this choice of τ considerably overestimates the severity of the early terrestrial impactor flux, or that the assumptions used above regarding the water content of heavy-bombardment asteroids are incorrect. For example, if S types were in fact responsible for 90% of terrestrial collisions,

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and if the cometary contribution were negligible, net water accretion would be reduced by a factor ~25, so that $\tau_c = 70$ Myr would deliver only $\sim 1.2 M_{oceans}$. It could also be that early in Solar System history, the impacting flux was dominated by water-poor objects in Earth-like orbits. The flux of such short-half-life objects would, however, have been superimposed over the flux of longer-lived objects represented by equation (1) (see for example ref. 12).

If α were as large as ~20%, the water accreted by Earth would be dominated by comets. Is such a result consistent with the terrestrial inventory of other volatile elements, especially carbon and nitrogen? The elemental abundance ratios found for comet Halley[29] indicate that Earth, were its carbon primarily cometary in origin, should have about five times as much nitrogen as it actually has[31]. But both the preferential impact erosion of nitrogen and the likelihood that comets supplied only a fraction of the terrestrial volatile inventory defuse this dilemma.

Finally, these calculations have implications for the volatile inventories of the other terrestrial planets. For reasons noted above, the Moon probably did not retain more than an insignificant fraction of the volatiles incident upon it (see ref. 2 for a further discussion). Delivery of volatiles to Venus and Mars depend on their impact rates relative to the Moon, but the analysis presented above indicates that Venus should have accreted approximately as much water as Earth. Mars, more subject to impact erosion[4], should have accreted an ocean equivalent to a layer of water ~10-100 metres deep uniformly distributed over the planet. This result is in agreement with other estimates[34], based on different methods, of the martian water inventory.

Note added in proof: W. K. Hartmann (personal communication and ref. 35) has considered spectrophotometric measurements of Solar System satellites thought to be captured, and concluded that asteroids scattered through the Solar System towards the end of the period of planet formation were probably >70% C type.

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Co-localization of molecules involved in antigen processing and presentation in an early endocytic compartment

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The pathways of intracellular traffic involved in antigen processing and presentation have been defined by immunoelectron microscopy. The export pathway for class II histocompatibility molecules and the antigen import pathway meet in a peripheral endocytic compartment having all the molecular machinery believed to be required for antigen processing and presentation, including internalized surface immunoglobulins, proteolytic enzymes and invariant chains. This compartment defines a site where peptides from endocytosed antigen can bind class II molecules *en route* to the cell surface for presentation to T cells.

T-CELL receptors recognize fragments of protein antigen bound to histocompatibility molecules[1-3]. The formation of such molecular complexes is called antigen presentation and requires the processing of antigen by denaturation and degradation[4]. Helper T cells are stimulated by class II histocompatibility molecules which generally present peptides from antigens that have entered B cells or macrophages by endocytosis[5]. Antigen uptake by B cells occurs by means of specific cell-surface immunoglobulins[6]. Cytotoxic T cells preferentially recognize class I histocompatibility molecules that bind to fragments from endogenously synthesized antigens, such as viral antigens[3]. The intracellular pathways involved in antigen processing and presentation have yet to be defined. But it is apparent that class I molecules bind antigenic fragments early during their biosynthesis[7-10], whereas class II molecules can bind peptides from exogenous antigens acquired by a different intracellular route[5,11,12].

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