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Terrestrial Mantle Siderophiles and the Lunar Impact Record

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A simple analytical fit to the lunar cratering record, scaled from final to transient crater diameters, then to impactor masses, implies a total mass $\sim 1.0 \times 10^{20}$ kg incident on the Moon subsequent to the solidification of the lunar crust ~ 4.4 Gyr ago. About half this mass would be retained, and a comparable lunar mass would be eroded. These results are in good agreement with geochemical estimates of the meteoritic component mixed into the lunar crust, which give $(0.4-1.5) \times 10^{20}$ kg. Gravitationally scaling to Earth, and taking account of the statistical probability that the largest impactors incident on Earth were more massive than the largest incident on the Moon, gives an estimate of 1.5×10^{22} kg of material accumulated by Earth subsequent to 4.4 Gyr ago. This result is in excellent accord with geochemical estimates of post-core formation meteoritic input. These estimates, based on abundances of highly siderophile elements in the terrestrial mantle, lie in the range $(1-4) \times 10^{22}$ kg. The significant result is the approximate agreement of the lunar cratering record scaling with both lunar and terrestrial geochemical constraints, numerous uncertainties render exact comparisons pointless. Nevertheless, the close agreement suggests the model developed here may credibly be used to estimate exogenous volatile and prebiotic organic delivery. © 1991 Academic Press, Inc.

I. TERRESTRIAL VOLATILES: A LATE-ACCRETING VENEER?

The role of impacts in the delivery of volatile elements and prebiotic organic molecules has a speculative history extending back at least to the beginning of this century (Chamberlin and Chamberlin 1908). Such speculation has, over the last two decades, received support from inhomogeneous terrestrial accretion models (Wasson 1971, Turekian and Clark 1975, Anders and Owen 1977, Chou 1978, Sun 1984, Wänke *et al.* 1984, Dreibus and Wänke 1987, 1989, Newson 1990) as well as dynamical models for outer planet formation (Fernández and Ip 1983, Shoemaker and Wolfe 1984), in which Earth receives the bulk of its surface volatiles as a late-accreting impactor veneer. Moreover, chemical equilibrium models predict one to several orders of magnitude less water and nitrogen on Earth than is in fact present (Prinn and Fegley 1989), indicating that some terrestrial accretion of volatile-rich material from greater heliocentric distances must have occurred (see, e.g., Wetherill 1990) or that nonequilibrium models must be considered. However, independently of solar nebula chemistry or planetary formation models, we may ask what the observed lunar cratering record tells us about terrestrial accretion of volatiles and organics during the period of heavy bombardment (Chyba 1987, 1990, Grinspoon 1988, Chyba *et al.* 1990). Such an approach intends to minimize the model-dependence of the conclusions, by basing the calculations as much as possible on the available data.

In practice, this procedure is much less successful in narrowing the uncertainties of the problem than one might hope. The oldest lunar province for which a radiometric date actually exists (the Apollo 16 and 17 uplands) is only 3.85–4.25 Gyr old; the ages of more heavily cratered provinces can at present only be estimated (BVSP 1981). For that matter, the entire interpretation of the heavy bombardment as representing exponentially decaying remnants of planetary formation is occasionally questioned by those favoring a lunar cataclysm (e.g., Ryder 1990). I have previously cautioned (Chyba 1990) that a range of impactor decay rates are permitted by the extant lunar data, and that different choices for these rates can lead to substantially different conclusions about terrestrial mass influx during the heavy bombardment. I suggested that lunar siderophile abundances may be used to constrain fits to the cratering record. This paper extends the procedure for generating simple analytical fits to the lunar data, to include recent work (Croft 1985, McKinnon and Schenk 1985, McKinnon *et al.* 1990) in the scaling of transient to final crater diameters, and considers constraints imposed by both lunar and terrestrial geochemical data. These constraints are complicated by impact erosion of both the lunar surface and the terrestrial atmosphere. I consider a first-order treatment of these effects, employing the Melosh and Vickery (1989) model for erosion by rapidly expanding post-impact vapor plumes, and scaling relationships (Housen *et al.* 1983) for ejecta speeds

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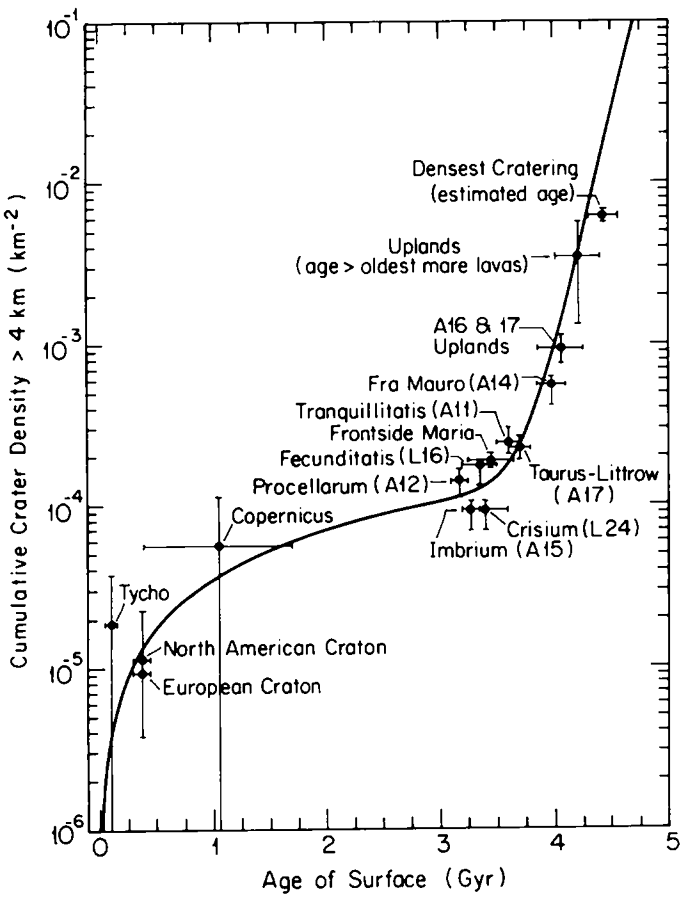

FIG. 1. Analytical fit to cumulative lunar crater density as a function of surface age [Eq. (1)], with a decay time constant τ = 144 Myr (100 Myr half-life).

greater than escape velocities. The paper concludes with an account of volatile delivery and erosion during the heavy bombardment that hopes to provide a best fit to the variety of data currently available to constrain the problem.

II. PROCRUSTEAN FITS TO THE LUNAR IMPACT RECORD

Figure 1 shows cumulative lunar crater density as a function of surface age, from the Basaltic Volcanism Study Project (BVSP 1981). Also shown is an analytical fit to these data, which is discussed below. Elsewhere I have compared other authors' fits to this, as well as an alternate lunar data set (Chyba 1990). The data, for craters bigger than a diameter D, are well modeled by the equation

$$N(t, D) = \alpha[t + \beta(e^{t/\tau} - 1)](D/4\text{ km})^{-1.8}\text{ km}^{-2}, \quad (1)$$

where t is in billions of years (Gyr), and α and β are determined by two-dimensional χ^2 minimization. [Eq. (1) is just the mathematical statement of the observation that cratering has been roughly constant for the past 3.5 Gyr, while increasing exponentially into the past prior to that time. Since Fig. 1 is a cumulative crater plot, its data are fit by the integral (over time t) of this constant plus exponential, giving a sum of a linear and exponential term. The logarithmic ordinate in Fig. 1 results in the term linear in t extrapolating to $-\infty$ as t goes to 0.] Choices of the decay constant τ in the literature have ranged from 70 to 220 Myr. Fundamentally, however, fitting any single decay "constant" to the cratering flux is a procrustean exercise, as the impactor flux cannot actually have decayed at a constant rate, but must rather have made a transition from rapidly swept-up objects in Earth-like orbits, to objects from comparatively long-lived, slowly decaying orbits (Wetherill 1977, Hartmann 1980, Grinspoon 1988). The fit derived in this section should therefore represent the impactor flux subsequent to the time after which the flux of those impactors with very short ($\tau \sim$ 20 Myr) sweep-up timescales has decayed to a negligible level, compared with that of more slowly ($\tau \sim$ 100 Myr) decaying populations (Grinspoon 1988, Chyba 1990).

I have argued (Chyba 1990) for choosing a heavy bombardment decay half-life of 100 Myr, or τ = 144 Myr [which gives $\alpha = 3.5 \times 10^{-5}$ and $\beta = 2.3 \times 10^{-11}$ in Eq. (1) by two-dimensional χ^2 minimization]. This choice is attractive because an ~100 Myr half-lie 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 (Fernández and Ip 1983, Shoemaker and Wolfe 1984). Therefore it might at least correctly represent the time-dependence of the cometary fraction of the heavy bombardment flux. In addition, this choice has the advantage of lying in the middle range of the values of τ with which the lunar cratering record is reasonably consistent. Finally, an independent estimate of τ by Oberbeck and Fogleman (1989), which correlates absolute age estimates for lunar impact basins from a crustal viscosity model (Baldwin 1987a) with crater counts for these basins (Baldwin 1987b), yields τ = 150 Myr. However, the most compelling reason for a choice of τ roughly consistent with a 100-Myr decay half-life is the discovery (see Section V) that such a choice is also in excellent agreement with lunar and terrestrial geochemical constraints.

The initially excavated (or transient) crater diameter D_{tr} (in meters, MKS units are used throughout) is related to the mass m of the incident impactor by (Schmidt and Housen 1987)

$$m = \gamma v^{-1.67} D_{tr}^{3.80}, \quad (2)$$

where v is impactor velocity, and γ is a constant that depends on surface gravity, impactor and target densities, and impactor incidence angle θ (taken to be 45°):

$$\gamma = 0.31 g^{0.84} \rho^{-0.26} \rho_t^{1.26} (\sin 45°/\sin \theta)^{1.67}. \quad (3)$$

Here $\rho_t = 2900$ kg m^{-3} is the lunar crustal density (Haines and Metzger 1980), and I take $\rho = 2200$ kg m^{-3} to be the density of a typical impacting asteroid and $g = 1.67$ m sec^{-2} to be the gravitational acceleration at the lunar surface (Melosh 1989). These values give $\gamma = 1.4 \times 10^3$ kg sec$^{-1.67}$ m$^{-2.13}$.

In previous work (Chyba 1987, 1990), I have cited investigations of wall collapse for craters in the 3- to 100-km range, showing final crater diameters enlarged over initial excavation (transient) diameters by ~30% (Shoemaker 1983), relating final to transient crater diameter via $D_f = c_f D_{tr}$, where the collapse factor $c_f = 1.3$. This treatment of the transient to final crater transition is a typical approximation [see, e.g., Melosh and Vickery (1989), who set $c_f = 1.25$], but more extensive empirical studies show that D_f cannot be related to D_{tr} by a simple proportionality factor (Croft 1985, McKinnon and Schenk 1985, McKinnon *et al.* 1990). Rather, D_f scales as D_{tr} to a power greater than one. Croft (1985) has suggested the relation

$$D_f \approx D_Q^{-0.18} D_{tr}^{1.18}, \quad (4)$$

where D_Q is the diameter of the simple-to-complex crater transition. For the Moon, Croft takes $D_Q = 15$ km. Equation (4) is not applicable to diameters smaller than 15 km; formally it gives $D_f < D_{tr}$ for such craters, an evident absurdity.

McKinnon *et al.* (1990), following McKinnon and Schenk (1985), present a somewhat different treatment, arguing that simple craters on the Moon near the simple–complex transition (which they define, on the basis of depth–diameter data, to be $D_c \approx 11$ km) are in fact ~15–20% wider than their original transient craters. McKinnon *et al.* (1990) write

$$D_f = k D_{tr}^{1.13}, \quad (5)$$

where

$$k = \kappa D_c^{-0.13}. \quad (6)$$

The requirement that D_f is 17.5% (the average of 15% and 20%) larger than D_{tr} when $D_f = D_c$, combined with Eqs. (5) and (6), allows κ to be determined; one finds $\kappa = (1.175)^{1.13} = 1.2$, whence

$$D_f = 1.2 D_c^{-0.13} D_{tr}^{1.13}, \quad (7)$$

with $D_c = 11$ km.

Equation (7) may be taken together with Eq. (2) to relate impactor mass and velocity to final (collapsed) crater diameter D_f:

$$m = 0.54 \gamma v^{-1.67} D_c^{0.44} D_f^{3.36}. \quad (8)$$

D_f is, of course, what one actually observes on the Moon. Eq. (1), with D properly interpreted as D_f, may then be combined with Eq. (8) to give the number of objects with mass > m that have impacted the Moon as a function of time t:

$$n(>m, t) = \alpha[t + \beta(e^{t/\tau} - 1)][m/m(4\text{ km})]^{-b}\text{ km}^{-2}, \quad (9)$$

where $b = (1.8/3.36) = 0.54$, and m(4 km) is given by Eq. (8) with $D_f = 4$ km. The total mass, $M(t)$, incident in impactors with masses in the range m_{min} to m_{max} on a lunar surface of age t is then:

$$M(t) = \int_{m_{max}}^{m_{min}} m[\partial n(>m, t)/\partial m]dm, \quad (10)$$

which yields

$$M(t) = \alpha[t + \beta(e^{t/\tau} - 1)]$$
$$[b/(1 - b)][m(4\text{ km})]^b m_{max}^{1-b}\text{ km}^{-2}, \quad (11)$$

where I have assumed $m_{max} \gg m_{min}$. Note that

$$M(t) \sim m_{max}^{0.46}[m(4\text{ km})]^{0.54} \sim v^{-1.67} D_{max}^{1.55}; \quad (12)$$

that is, $M(t)$ depends much more weakly upon D_{max}, the largest crater diameter used in the calculation, than one might have anticipated from the Schmidt–Housen scaling, Eq. (2). Uncertainties in D_{max} and v lead to final uncertainties in $M(t)$ of a factor of a few (see Sections III and IV).

$M(t)$ from Eq. (11) may be gravitationally scaled to determine the total mass incident on Earth during the heavy bombardment, subsequent to a time, t. Several variables must first be determined, however: the mass, m_{max}, of the largest impactor (or, equivalently, D_{max}), and a "typical" impactor velocity, v. These choices are the topic of the next two sections.

III. IMPACT VELOCITIES ON THE EARTH AND MOON

A common approach (BVSP 1981, Melosh and Vickery 1989) for determining v in Eq. (8) has been to use the root mean square (rms) impact velocity v_{rms} for the known Earth-crossing asteroids. This can at best provide a "snapshot" of asteroid–Earth collisions, as evolution of orbits (mainly due to Jovian perturbations) will alter perihelia (q), which strongly influence collision probabilities and velocities. Note, however, that the calculations cited below (Kessler 1981, Steel and Baggaley 1985) do average over simple secular precession of asteroid perihelia, which occurs on timescales of $\sim 10^4$–10^5 years (Shoemaker *et al.* 1979), short compared to typical asteroid dynamical lifetimes in the inner solar system. For the Earth-crossing asteroids these lifetimes are $\sim 10^7$–10^8 years (Steel and Baggaley 1985); an implicit assumption of this approach, then, is that the current Earth-crossing asteroid swarm is approximately representative of a "typical" distribution of Earth-crossing impactor velocities.

Using rms velocities in impact calculations skews "typical" velocities toward misleadingly high values (Chyba 1990). Previous work (Melosh and Vickery 1989, Sleep *et al.* 1989, Chyba 1990) used velocities calculated from the 20 Earth-crossing asteroids known in 1981 (BVSP 1981). But the orbits of 65 such objects were known by May 1989 (Olsson-Steel 1990), so velocity calculations may now be updated, with a substantial improvement in statistical reliability.

Figure 2a shows the percentage of asteroid–Earth collisions that occur at a given velocity, using Olsson-Steel's recent (1990) compilation of Aten and Apollo asteroids. The method used by Olsson-Steel to calculate individual asteroid collision probabilities and velocities (Kessler 1981, Steel and Baggaley 1985) results in minimum (v_{min}) and maximum (v_{max}) possible velocities at infinity for a given mean collision probability. I have averaged these extremes to find the values of v_∞ used here; the uncertainty thereby introduced in <0.5 km sec^{-1}, small compared with other unknowns in the problem.

Given values of v_∞ and probabilities of terrestrial collisions for each object, it is a simple matter to calculate the percentage of Earth-crossing asteroid collisions with Earth as a function of impact velocity v, where

$$v^2 = v_\infty^2 + v_{\text{esc}}^2, \tag{13}$$

and the escape velocity $v_{\text{esc}} = 11.2$ km sec^{-1} for Earth. The results are shown in Fig. 2a. For asteroid–Earth collisions, these data yield $v_{\text{rms}} = 18$ km sec^{-1}, and an average velocity $v_{\text{av}} = 17$ km sec^{-1}. However, about 50% of collisions occur at velocities below $v_{\text{med}} = 15$ km sec^{-1}, which I therefore take to be the "typical" value in the median sense. These values are well below those calculated from the much smaller BVSP (1981) data set, for which $v_{\text{rms}} =$ 25 km sec^{-1}, $v_{\text{av}} = 20$ km sec^{-1}, and $v_{\text{med}} = 17$ km sec^{-1}.

These calculations may be easily extrapolated to the Moon, as shown in Fig. 2b. This extrapolation is not merely a case of replacing v_{esc} for Earth in Eq. (13) with $v_{\text{esc}} = 2.4$ km sec^{-1} for the Moon, because collision probabilities, which depend on the quantity $[1 + (v_{\text{esc}}/v_\infty)^2]$, also change relative to one another. For lunar collisions, I find $v_{\text{rms}} = 16$ km sec^{-1}, $v_{\text{av}} = 14$ km sec^{-1}, and $v_{\text{med}} = 12$ km sec^{-1}; this last value is also, to within 2%, the median value of v_∞. Note that throughout this paper, the effects of the Moon's orbital velocity (~ 1 km sec^{-1}) with respect to Earth are neglected; this approximation never introduces an error >1% in any of the following calculations.

The Moon should have been much closer to Earth 4.5 Gyr ago than it is now. Lunar orbital evolution timescales, however, suggest that the resulting enhanced gravitational focusing effects are unimportant subsequent to ~4.4 Gyr ago (Chyba 1990). As (see Section V below) we are concerned only with the impactor flux subsequent to this time, we may ignore this possible very early near-Earth stage of lunar history.

What of the Earth-crossing short-period (SP) comets? Because models of the early cometary bombardment of the inner Solar System (Fernández and Ip 1983, Shoemaker and Wolfe 1984) 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, it is appropriate to use contemporary SP comet orbits as analogues to those of the cometary component of the heavy bombardment. Previously (Chyba 1990), I used Weissman's (1982) compilation of 20 SP comet impact velocities and probabilities. These statistics may now be slightly improved, with the addition of comets Tuttle, Schwassmann–Wachmann 3, and Wilk; however, comet Giacobini–Zinner's perihelion has evolved to $q = 1.028$ AU, so it is no longer an Earth crosser (Olsson-Steel 1987). Figures 2c and 2d show the results of using Olsson-Steel's (1987) compilation for 22 SP comets, following the procedure used above for asteroids. Obviously the statistics for SP comets are considerably worse. The results are nearly identical to those found using Weissman's (1982) data: For collisions with the Earth, I find $v_{\text{rms}} = 30$ km sec^{-1} and $v_{\text{av}} = 27$ km sec^{-1}, whereas for lunar collisions, $v_{\text{rms}} = 30$ km sec^{-1} and $v_{\text{av}} = 26$ km sec^{-1}. (The effects of terrestrial gravitational focusing are relatively minor for SP comet orbits, due to their high values of v_∞.) Median collision velocities are more uncertain than for asteroids, because of the poorer cometary statistics; roughly, $v_{\text{med}} \approx 23$ km sec^{-1}. For cometary collisions with the Moon, $v_{\text{med}} \approx 20$ km sec^{-1}, equal to v_∞ to within 1%.

The asteroidal distribution of percentage collisions vs impact velocity appears very different from the cometary one. Crudely speaking, the cometary distribution has a "bell-shaped" appearance, whereas the asteroidal has a "decaying-exponential" one. This difference may be rationalized as follows. The more elliptical cometary orbits have greater values of v_∞, so the resulting percentage collisions vs impact velocity distribution reflects the "intrinsic" distribution of cometary orbits. In the case of

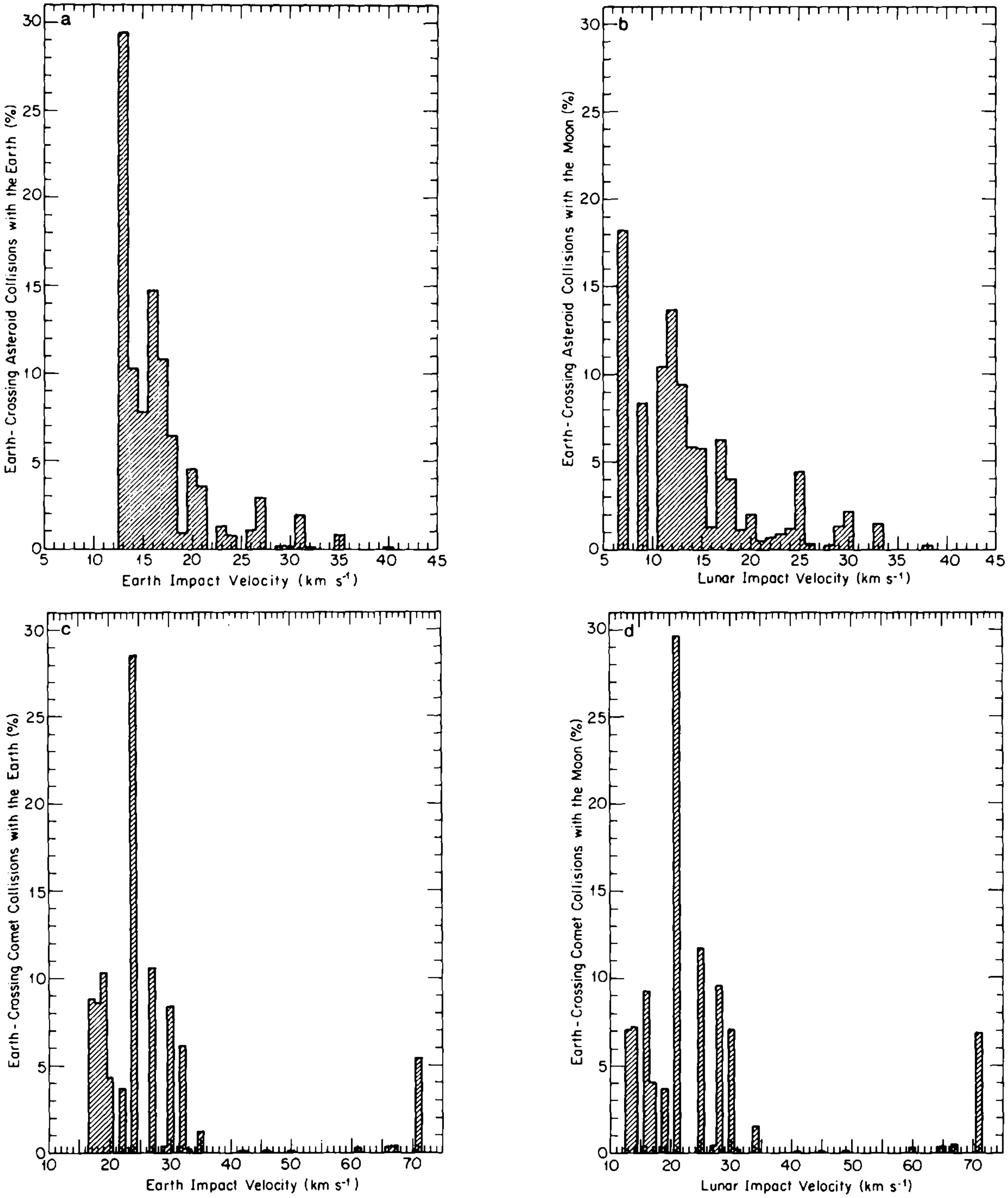

FIG. 2. Percentage of Earth-crossing asteroid and short-period (SP) comet collisions with the Earth and Moon as a function of impact velocity. Percentage collisions of asteroids with (a) Earth, and (b) the Moon, based on data from Olsson-Steel (1990). Percentage collisions of SP comets with (c) Earth, and (d) the Moon, based on data from Olsson-Steel (1987).

Earth-crossing asteroids, however, the orbits are far more circular, with resulting much smaller values of v_∞, so that there are comparatively slow (and long) overlaps with the Earth's orbit. Objects with sufficiently small values of v_∞ therefore acquire large collision probabilities.

I take the value $v_{\text{med}} = 12$ km sec^{-1} for asteroid impacts on the Moon to be the most appropriate value for v in Eqs. (8) and (11). With $v = 12$ km sec^{-1}, Eq. (8) yields $m(4 \text{ km}) = 8.9 \times 10^9$ kg. How sensitive are the results of Eq. (11) to different choices of v? Using Eq. (12), we see that, were we to use the asteroidal v_{rms} instead of v_{med} in Eq. (11), $M(t)$ would decrease by $(16/12)^{1.67} = 1.6$. If we used the cometary value of v_{med}, instead of the asteroidal one, $M(t)$ would decrease by $(20/12)^{1.67} = 2.3$. The range of values for v, then, seems to introduce an error of at most a factor of a few into $M(t)$.

IV. MAXIMUM-MASS IMPACTORS

Calculating $M(t)$ in Eq. (11) requires a value for m_{max}. For the Moon, m_{max} will be the mass of the impactor that excavated the largest lunar basin, taken here to be South Pole–Aitken. This ancient lunar farside basin was first tentatively identified photogeologically (see Wilhelms 1987); its existence has now been clearly demonstrated by Galileo observations (Head *et al.* 1991). Wilhelms (1987) suggests a pre-Nectarian age for South Pole–Aitken of 4.1 Gyr, and identifies (his Table 4.1) an average ring diameter of 2200 km (or, possibly, two rings of 1800–2000 km and 2500 km). It is true that other giant, highly degraded lunar basins have been suggested, on the basis of both photogeological evidence (Wilhelms 1987) and subsurface mass concentrations (Campbell *et al.* 1969, O'Leary *et al.* 1969), but their existence remains controversial. In particular, Spudis *et al.* (1988) have recently argued that the reality of the proposed giant "Procellarum" basin is called into question by these authors' conclusion that the large-scale, concentric structural pattern on the lunar nearside probably represents Imbrium rings (ranging up to 3200 km in diameter), rather than a "Procellarum" signature. Impacts larger than Imbrium would probably have excavated lunar mantle material, but no lunar mantle samples have ever been found (Sleep *et al.* 1989). Indeed, Head *et al.* (1991) have suggested, on the basis on Galileo multispectral imaging, that the South Pole–Aitken basin may contain lunar mantle material.

Sleep *et al.* (1989) have argued that lunar impacts similar to or larger than Imbrium must have been rare after 4.4 Gyr ago, or else the lunar regolith would be more stirred than it appears to be. If South Pole–Aitken is taken as the largest lunar basin, with $D_f = 2200$ km, how many basins as large or larger than Imbrium are implied? The observed number of lunar craters larger than some diameter D falls off as $D^{-1.8}$, a power law that is well obeyed in both the lunar frontside highlands and the heavily cratered uplands, for histogram bins in D up to $D = 512\sqrt{2} = 724$ km (BVSP 1981). Assuming this N–D power law to hold for the largest objects (an assumption examined further below), we may write:

$$N(> D_f) \propto D_f^{-1.8}. \tag{14}$$

This equation allows the number of basins as large or larger than Imbrium, which has a diameter ~1160 km (see below), to be calculated from

$$N(> 1160 \text{ km}) = N(> 2200 \text{ km}) [(1160 \text{ km})/(2200 \text{ km})]^{-1.8}. \tag{15}$$

Taking $N(> 2200$ km) to mean the number of basins as large or larger than South Pole–Aitken, $N(> 2200 \text{ km}) = 1$ so that $N(> 1160 \text{ km}) \approx 3$. Of these three basins, South Pole–Aitken is one; if we consider Imbrium to represent a second, there remains only one other giant. The statistics here are of such small numbers that this object could easily not exist. It seems that regolith stirring arguments do not exclude $D_f = 2200$ km for South Pole–Aitken.

What is the mass of the South Pole–Aitken impactor? Choosing $D_f = 2200$ km in Eq. (7) corresponds to an initial excavation (transient) diameter for the basin of $D_{tr} = 1020$ km. Eq. (2) [or Eq. (8) directly] then yields a South Pole–Aitken object mass of 1.4×10^{19} kg. Should Eqs. (2), (7), and (8) be trusted for giant basins? The answer appears to be yes, based on analogous calculations for the masses of the Orientale and Imbrium objects, which may be checked against independent estimates. For example, Sleep *et al.* (1989) obtain a mass for the Orientale object from a thermal contraction estimate of the heat buried by this impact, and then scale the result to the Imbrium impactor. They conclude that the Orientale object had an impact kientic energy of 1.2×10^{26} J, with an uncertainty of about a factor of 3. Choosing an impact velocity $v = 13$ km sec^{-1}, they find an Orientale impactor mass of 1.4×10^{18} kg. An estimate of Orientale ejecta mass of 2.4×10^{19} kg then yields a ratio of ejected mass to projectile mass of about 17. Imbrium ejected about 3.6×10^{19} kg of material (Spudis *et al.* 1988), so by this method (Sleep *et al.* 1989), the Imbrium object had a mass of about 2.1×10^{19} kg, with an evident uncertainty of a factor of ~3. These estimates may be compared to the results of calculating Orientale and Imbrium object masses from Eq. (8). Choosing D_f for these basins to be 930 km (Wilhelms 1987, Baldwin 1987a) and 1160 km (Wilhelms 1987, Baldwin 1987a, Spudis *et al.* 1988), respectively, Eq. (8) yields impactor massess of 8.1×10^{17} kg and 1.7×10^{18} kg. Both estimates agree with the independent calculations of Sleep *et al.* to better than a factor of two. Since Eq. (8) appears reliable for calculating impactor masses for those giant basins where its results may be compared with independent estimates, I will employ it as well for the South Pole–Aitken object, and take $m_{max} = 1.4 \times 10^{19}$ kg in Eq. (11).

Equation (11) may now be used to calculate the total mass incident upon the Moon subsequent to some time t, and this result extrapolated to Earth. Section V compares these results with constraints imposed by lunar and terrestrial geochemical evidence. For the Moon, this evidence depends on the meteoritic component (determined from Ir and, possibly, Ni abundances) retained in lunar samples; Ir and Ni would have been retained in the lunar regolith subsequent to that date when a mostly solid crust first existed (Sleep *et al.* 1989). These authors cite samarium–neodymium (Sm–Nd) isotopic evidence that the upper lunar crust, ferroan anorthosite, solidified as early as 4.44 ± 0.02 Gyr ago (Carlson and Lugmair 1988); whereas the age of KREEP basalt (Carlson and Lugmair 1979) implies 4.36 ± 0.06 Gyr ago as the solidification of the base of the crust. I therefore take 4.4 Gyr as the approximate time t to use in Eq. (11) for the Moon. Since $\tau = 144$ Myr, an error in this choice of ~40 Myr will change $M(t)$ by ≲30%.

Inserting these values into Eq. (11) gives $M(4.4 \text{ Gyr}) = 2.7 \times 10^{12}$ kg km^{-2}, or, with a lunar radius $r_m = 1738$ km, $M(4.4 \text{ Gyr}) = 1.0 \times 10^{20}$ kg delivered to the Moon since 4.4 Gyr ago. (The mass of the South Pole–Aitken object was ~14% of this total.) This mass flux is the appropriate value to scale to Earth. It is not, however, the net mass actually accreted by the Moon since 4.4 Gyr ago: Probably only ~50% of the objects striking the Moon during heavy bombardment were moving slowly enough for their mass to be accreted (see Section VI). Moreover, significant impact erosion of the Moon should also have taken place.

Scaling from the Moon to Earth requires accounting for Earth's larger gravitational cross section. A planet's gravitational cross section is $\sigma = \pi R_g^2$, where the gravitational radius R_g for a planet of physical radius R and escape velocity (v_{esc}) is given by

$$R_g = R[1 + (v_{esc}/v_\infty)^2]^{1/2}. \qquad (16)$$

With $v_\infty = 12$ km sec^{-1}, the ratio of the Earth's gravitational cross section to that of the Moon is then

$$\sigma_\oplus/\sigma_m \equiv \xi = 1.8(R_\oplus/r_m)^2. \qquad (17)$$

The Earth's volume is equivalent to a sphere of radius $R_\oplus = 6371$ km (Stacey 1977), so that $\xi \approx 24$. Multiplying $M(4.4 \text{ Gyr})$ for the Moon by ξ gives 2.4×10^{21} kg of meteoritic material accreted by Earth subsequent to 4.4 Gyr ago.

However, this simple scaling does not fully account for the role of Earth's larger gravitational cross section. Because $\xi \approx 24$, it is likely that the largest impactors in the Earth–Moon system were collected by the Earth. To try to quantify this effect, I follow the approach of Sleep *et al.* (1989), and write the probability P that the Moon is not hit by any of the largest n impactors:

$$P = (24/25)^n, \qquad (18)$$

so that $P \approx 50\%$ provided $n \approx 17$. Thus if South Pole–Aitken is the largest lunar impactor subsequent to 4.4 Gyr ago, 17 larger objects are expected to have hit Earth since that time. I take the sizes of these objects to be distributed according to Eq. (14).

Equation (8) implies $D_f \propto m^{0.30}$, so Eq. (14) may be transformed into an equation for the number of impactors with masses $> m$:

$$N(> m) \propto m^{-0.54}. \qquad (19)$$

If Earth collected 17 objects with masses $> 1.4 \times 10^{19}$ kg, Eq. (19) implies Earth should also have been struck by one object with mass $m > 2.6 \times 10^{21}$ kg, the terrestrial maximum-mass impactor subsequent to 4.4 Gyr ago. How much larger than 2.6×10^{21} kg was this object? Any such "statistics-of-one" question can only be answered probabilistically, and with concomitant caution. One approach to this question is to begin with Eq. (19), and ask: What mass must an impactor have had such that Earth would have collected 0.5 such objects? Equation (19) then yields

$$(17/0.5) = [(1.4 \times 10^{19} \text{ kg})/m]^{-0.54}, \qquad (20)$$

giving $m = 9.5 \times 10^{21}$ kg. The total mass, M, of all 17 impactors is, analogously, given by the sum

$$M = \sum_{i=1}^{17} (1.4 \times 10^{19} \text{ kg})[(i - 0.5)/17]^{-1.85}, \qquad (21)$$

or $M = 1.2 \times 10^{22}$ kg. (Clearly there is great uncertainty in this result.) Summing the mass incident on Earth in the largest impactors with that previously found yields a net accretion subsequent to ~4.4 Gyr ago of 1.5×10^{22} kg.

Sleep *et al.* (1989) find that an impact energy of 2×10^{28} J is sufficient to evaporate the entire terrestrial ocean; such impacts may possibly have globally sterilized the early Earth. With a typical Earth impact velocity 15 km sec^{-1}, a kinetic energy 2×10^{28} J corresponds to an impactor of mass 1.8×10^{20} kg. If 17 larger-than-South Pole–Aitken impactors are taken to have hit Earth during the heavy bombardment, Eq. (21) shows that, statistically, Earth should have sustained ~4 such impacts between ~4.4 and 3.8 Gyr ago.

The derivation of the maximum-mass terrestrial impactor found here relies on the exponent s in the relationship $N(>m) \propto m^{-s}$. Equation (19), with $s = 0.54$, has the clear advantage of consistency with the ancient lunar cratering record and crater diameter–mass equations adopted throughout this paper. Nevertheless, one may ask to what extent our results would differ had some other plausible value for s been chosen. For example, theoretical treatments of collisionally evolved planetesimal systems predict $s = 0.83$ when fragmentation dominates, 0.67 with both fragmentation and coagulation present, and 0.5 for coagulation in the absence of comminution (Greenberg 1989, and refs. therein). Empirically, contemporary Solar System asteroids with diameters $D < 260$ km approximately follow a distribution with $s = 1.0 \pm 0.1$, and those with $D > 260$ km have $s = 0.83 \pm 0.25$ (Donnison and Sugden 1984). Short-period comets with magnitudes $H_{10} < 10.8^m$ [i.e., those with radii $\gtrsim 1$ km (Hughes 1990)] have $s = 0.50$. [Unfortunately, the situation is less clear cut than these citations imply. For example, in an earlier treatment of asteroidal $N - m$ distributions than that cited above, Hughes (1982) found that asteroids with $D > 260$ km had $s = 0.58 \pm 0.25$. And it is certainly the case (Gradie *et al.* 1989) that values of s differ between different asteroid classes, as well as between different size ranges within given classes.] In any case, evidently s in Eq. (19) does not lie outside the range of size distributions for known solar system objects. Suppose that instead of Eq. (19) for the largest impactors ($s = 0.54$), we were to adopt $s = 1.0$, the empirical distribution differing the greatest from Eq. (19). Then the total mass collected by the Earth would be reduced by a factor ~4, compared with the values found above.

V. LUNAR AND TERRESTRIAL GEOCHEMICAL CONSTRAINTS

What constraints on the heavy bombardment are provided by lunar and terrestrial geochemistry? Using Ir abundances, Sleep *et al.* (1989) estimate a meteoritic component mixed into the upper half of the lunar crust (35 km) of between 1 and 4%, with 2% their preferred value. This corresponds to a total meteoritic thickness of 0.7 km of material, or, using our previous estimate for the density of the lunar crust (Haines and Metzger 1980), a total extralunar mass of 7.7×10^{19} kg (~0.2% of the current lunar mass), with an uncertainty of a factor of two. This mass would have been accreted subsequent to the solidification of the lunar crust ~4.4 Gyr ago.

In the case of the Earth, one may use siderophile abundances in ultramafic xenoliths (more or less unaltered solid mantle material brought to the surface by volcanic eruptions of basaltic magmas), as well as basalts, peridotites, and other mantle-derived rocks, to estimate the meteoritic component mixed into the Earth's mantle (Chou 1978, BVSP 1981, Sun 1984, Wänke *et al.* 1984, Dreibus and Wänke 1987, 1989, Newsom 1990). Among the xenoliths, apparently unaltered (or nearly unaltered) mantle samples have been found only among the spinel-lherzolites, which represent upper mantle material from depths up to 70 km (Wänke *et al.* 1984). The observed near-chondritic proportions of siderophile elements in these samples are evidently not the result of metal–silicate fractionation processes, as it is unlikely that the partitioning of all of these elements upon fractionation is similar (BVSP 1981). This conclusion is supported by detailed modeling of various core formation theories (Newsom 1990). It has therefore been suggested that the highly siderophile "noble metals", Ru, Rh, Pd, Re, Os, Ir, Pt, and Au, were added to the upper mantle during the heavy bombardment, subsequent to terrestrial core formation (Chou 1978, BVSP 1981, Sun 1984, Newsom 1990). [During core formation, virtually all noble metals present in the mantle should have been incorporated into the core (Sun 1984, Newsom 1990).] The nearly chondritic ratios of the noble metals excludes fractionated impactors (such as eucrites) as sources (Chou 1978).

Noble metal abundances in upper mantle samples can be explained by ~1% of CI carbonaceous chondrite input subsequent to core formation (Chou 1978, BVSP 1981, Sun 1984, Dreibus and Wänke 1989). How much total extraterrestrial mass does this represent? This depends on whether one takes the post-core formation input to have been mixed throughout the entire, or only the upper, mantle. The upper mantle appears to be well-mixed, as indicated by studies of isotope ratios and distributions of rare earth elements in midocean ridge basalts; convective mixing in the upper mantle greatly reduces heterogeneities on a timescale of several hundred million years (Turcotte and Kellogg 1986). Therefore it certainly appears reasonable to take the CI chondrite abundances implied by available samples to correctly represent typical upper mantle abundances. If one assumes that the CI input is mixed only until this depth (Chou 1978), the CI abundance may be multiplied by the mass of the upper mantle to obtain the total mass accreted by Earth during the heavy bombardment.

The depth of Earth's upper mantle is defined by the seismic discontinuity at 670 km (Stacey 1977, Turcotte and Kellogg 1986). Virtually no earthquake activity occurs below this depth; it remains controversial whether subducted lithospheric slabs ever penetrate this layer (Frohlich and Grand 1990). However, even if a barrier to mantle convection exists at 670 km depth on the contemporary Earth (Turcotte and Kellogg 1986), it nevertheless remains possible that a vigorously convecting early Earth would have experienced whole mantle convection, and therefore mixed the extraterrestrial component throughout the entire mantle (Wänke *et al.* 1984).

I estimate a mass for the terrestrial upper mantle from an Earth interior model in which density is a simple analytic function of depth (Dziewonski *et al.* 1975, Stacey 1977). Integrating from a radius of 5701 km (670 km depth) up to 6352 km (the base of the crust, for an "average structure" model), I find the mass of the upper mantle to be 1.1×10^{24} kg. Assuming this to be ~1% CI chondritic then gives $\sim 1.1 \times 10^{22}$ kg as an estimate of the total mass accreted by Earth during the heavy bombardment, subsequent to core formation. This value is a lower limit, as it assumes mixing occurred only throughout the upper mantle. An upper limit should be given by taking mixing to occur throughout the entire mantle, which yields 4.0×10^{22} kg total accreted mass [taking the mass of the whole mantle to be 4.0×10^{24} kg (Stacey 1977)]. Therefore the actual value is likely to be bracketed by the range $\sim(1-4) \times 10^{22}$ kg.

Wänke *et al.* (1984) and Dreibus and Wänke (1987, 1989) have argued that mixing should have occurred throughout the entire mantle, and find a total CI component equivalent to 0.44% the mass of the Earth, or 2.6×10^{22} kg. [Given the mass of the Earth, $M_{\oplus} = 5.98 \times 10^{24}$ kg, these authors have effectively taken the CI component of the mantle to be 0.65%.] A detailed model recently presented by Newsom (1990) requires a late veneer chondritic component equivalent to 0.2% $M_{\oplus}$, or 1.2×10^{22} kg. Again, this value falls within the range, $(1-4) \times 10^{22}$ kg, considered here.

Terrestrial core formation is believed to have taken place within $\sim 10^8$ yr of the formation of the Earth (Stevenson 1983, 1990, Sun 1984). The Earth's age may be estimated by iodine–plutonium–xenon (I–Pu–Xe) dating of the atmosphere and mantle, which suggests formation 75–100 Myr after t_0, the time of formation of primitive meteorites (Swindle *et al.* 1986). I–Pu–Xe dating of lunar samples suggests a lunar age of 63 ± 42 Myr after t_0, between 80 Myr before and 30 Myr after the terrestrial formation date. (Therefore, if the impact-trigger hypothesis for the formation of the Moon is correct, the Moon-forming event occurred no more than 30 Myr subsequent to terrestrial accretion.) A period of 75–100 Myr for the formation of the Earth subsequent to meteorite formation is in excellent agreement with dynamical models for the formation of the terrestrial planets, which yield $\sim 10^8$-yr timescales for the accretion of the Earth from an initial population of planetesimals (Wetherill 1977, Greenberg 1989; see also Grinspoon 1988). A recent study by Wetherill (1990) yields 99% accretion of the Earth after 77 Myr.

What is the value of t_0? The oldest high-precision meteorite date comes from calcium–aluminum-rich inclusions of the CV chondrite Allende, which give a $^{207}Pb/^{206}Pb$ model age of 4.559 ± 0.004 Gyr (Tilton 1988). This value in turn implies that terrestrial formation was virtually complete by 4.49–4.46 Gyr ago. The formation of Earth's core would then have taken place by ~4.4 Gyr ago.

Therefore, it is appropriate to compare the extraterrestrial mass input implied by terrestrial geochemical data, with that found by extrapolating from the lunar cratering record to the Earth using $t \approx 4.4$ Gyr in Eq. (11). The agreement between the two techniques is excellent. The value found via the lunar cratering extrapolation, 1.5×10^{22} kg, lies within the range $(1-4) \times 10^{22}$ kg implied by the geochemical data. Moreover, the extrapolation is reasonably robust against a different choice for the time of terrestrial core formation; for example, changing the time of core formation by 100 Myr will change the result by only a factor of two.

VI. IMPACT EROSION OF THE EARTH AND MOON AND ORIGINS OF THE BOMBARDING POPULATION

In Section IV, I used the lunar cratering record to derive the total impactor mass incident on the Earth and Moon subsequent to ~4.4 Gyr ago. In Section V, these results were compared with those implied by terrestrial and lunar geochemical data. In order for these comparisons to be appropriate, however, an account must be made of the erosion of the Earth and Moon by high-speed impacts. This section presents a first-order treatment of this problem for both worlds, considering erosion due both to loss of high-velocity crater ejecta, and to atmospheric erosion by expanding post-impact vapor plumes.

To treat the latter effect, I follow Melosh and Vickery (1989), who find that large impacts may cause atmospheric erosion provided that the impactor strikes the planet at a velocity high enough for a vapor plume to form and expand at a speed $> v_{esc}$, and that the mass of the plume exceeds the air mass above the plane tangent to the impact. Here I consider an impactor that satisfies these two criteria to be entirely lost as an escaping vapor plume from the target planet. The Melosh and Vickery (1989) treatment takes a mass of the target equal to that of the impactor to be incorporated into the plume, and similarly lost. (This should be considered as an extremely rough-and-ready approximation, whose uncertainty should be remembered throughout the following discussion.)

Melosh and Vickery (1989) show that the threshold impact velocity v_{min} for most of a vapor plume to exceed v_{esc} is given by

$$v_{min}^2 = 4(v_{esc}^2 + 2H_{vap}), \quad (22)$$

where H_{vap} is the vaporization energy, taken to be 13 MJ kg^{-1} for silicates and 3 MJ kg^{-1} for ice (1 MJ kg^{-1} = 1 km^2 sec^{-2}). For the Moon, $v_{\min}$ is a bit over 11 km sec^{-1} for silicate projectiles ("asteroids") and about 7 km sec^{-1} for ice impactors ("comets"). As may be seen from Fig. 2d, virtually all vapor plumes resulting from cometary impacts on the Moon (for comets of any mass, as there is no atmosphere to be overcome) would be lost. In the case of asteroids, as Fig. 2b illustrates, nearly 40% of lunar collisions occur at velocities below 11 km sec^{-1}, and about 50% below 12 km sec^{-1}. Recognizing that published treatments of impact erosion remain approximate at best, I will hereafter take 50% as a rough estimate of the fraction of asteroidal collisions with the Moon that result in escaped vapor plumes.

For Earth, $v_{\min} \approx 25$ km sec^{-1} for asteroids, and about 23 km sec^{-1} for comets. As seen in Figs. 2a and 2c, ~90% of Earth-colliding asteroids, and ~50% of Earth-colliding comets, impact Earth with velocities below the appropriate $v_{\min}$. Therefore Earth retains nearly all asteroid-delivered volatiles, and about half of those brought in by comets; comets erode about as much terrestrial mass as they deliver. Atmospheric volatiles might be significantly eroded, as discussed in Section VII below.

In addition to erosion by vapor plumes, we must also consider loss of target material in that fraction of crater ejecta propelled at velocities $> v_{\rm esc}$. Quantitative treatments of this effect, like those of atmospheric erosion, remain at an early and uncertain stage (Melosh 1989). Housen *et al.* (1983) have used dimensional analysis and scaling arguments to derive the functional form of an expression for the volume of ejecta, $V_{\rm e}(>v_{\rm e})$, with velocity equal to or greater than some velocity $v_{\rm e}$. For cratering in the gravity regime, they find

$$V_{\rm e}(>v_{\rm e})R^{-3} = K(v_{\rm e}/\sqrt{gR})^{-\nu}, \tag{23}$$

where $R = D_{\rm tr}/2$ is the radius of the transient (initially excavated) crater, and K and ν are (in general, target-dependent) constants. Experiments with targets of Ottawa sand yield $K = 0.32$ and $\nu = 1.2$ (Housen *et al.* 1983). Cratering equations previously employed in this paper [e.g., Eq. (2)] and elsewhere (Melosh and Vickery 1989, Chyba 1990) make use of results for cratering in competent rock, rather than sand—the former providing a better model for cratering in the lunar crust than the latter. However, the only experiments available for crater ejecta velocities are for sand targets, so that K is known only for sand. I will therefore model ejecta loss using exclusively parameters for sand targets. A treatment for competent rock must await relevant experimental data.

Equation (23) may be combined with a crater diameter–impactor mass scaling law to determine how much ejecta is propelled at velocities $>v_{\rm e}$ for a projectile of a given mass and velocity incident upon a certain target world. A general form for the crater diameter–impactor mass relation is (Schmidt 1980, Schmidt and Housen 1987, Melosh 1989):

$$\pi_{\rm D} = C_{\rm D}\pi_2^{-\beta}, \tag{24}$$

where

$$\pi_{\rm D} = D_{\rm tr}(\rho_{\rm t}/m)^{1/3}, \tag{25}$$

and

$$\pi_2 = 3.22gr/v^2. \tag{26}$$

Here $C_{\rm D}$ and β are constants, g is surface gravity, and m, r, and v are the incident impactor's mass, radius, and velocity, respectively. (Eq. (2) is just Eq. (24) with parameter values appropriate to competent rock, viz. $C_{\rm D} = 1.6$ and $\beta = 0.22$.) To be consistent with Eq. (23) above, in the following treatment only, I must employ parameters for Ottawa sand: $C_{\rm D} = 1.68$ and $\beta = 0.17$ (Schmidt 1980, Schmidt and Housen 1987). Combining Eqs. (23) through (26), assuming a spherical impactor, then yields

$$M_{\rm e}(> v_{\rm e}) = 0.11(\rho/\rho_{\rm t})^{0.2}(v/v_{\rm e})^{1.2}m, \tag{27}$$

where $M_{\rm e}(> v_{\rm e})$ is the mass of ejecta with velocity greater than some velocity $v_{\rm e}$, and ρ and $\rho_{\rm t}$ are the impactor and target densities, respectively ($M_{\rm e} = \rho_{\rm t}V_{\rm e}$). Using parameters appropriate to asteroid collisions with the Moon, and taking $v_{\rm e}$ to equal $v_{\rm esc} = 2.4$ km sec^{-1}, the lunar escape velocity, Eq. (27) becomes $M_{\rm e}(> v_{\rm esc}) \approx 0.7\ m$. Thus, a "typical" asteroid incident on the Moon erodes in ejecta alone ~70% as much mass (m) as it delivers. From Fig. 2b and Eq. (27), it may be seen that nearly 20% of incident asteroids will erode about half as much as this, whereas the few objects in the high-velocity tail may erode considerably more. As a very rough approximation, it seems that a net mass M of asteroids incident upon the Moon will erode about a mass M worth of lunar material in the form of ejecta. In addition, the Moon will lose mass due to escaping impact vapor plumes. Therefore the Moon appears to have experienced a net impact erosion due to "asteroid" collisions subsequent to ~4.4 Gyr ago.

Comets, with higher typical velocities, will tend to erode in high-speed ejecta slightly more lunar material than they deliver. Moreover, as previously argued, it appears that virtually all cometary collisions will result in vapor plumes that expand away from the Moon at velocities greater than $v_{\rm esc}$. These plumes will carry away not only the impactor mass, but a comparable target mass as well (Melosh and Vickery 1989). Thus a typical collision of a comet of mass m will add no mass to, and erode a mass $\sim 2m$ from, the Moon.

What of the Earth? Inserting the appropriate terrestrial values into Eq. (27) gives $M_c(> v_{sec}) \approx 0.1\ m$, suggesting terrestrial erosion by high-speed ejecta is a small effect. This conclusion is reinforced by the likelihood that much of the ejecta escaping from Earth will be reaccreted from their resulting Earth-crossing heliocentric orbits. (Note that in the lunar case, while some ejecta would be trapped in Earth orbit and reaccreted by the Moon, Eq. (27) may be used to show that the majority would escape from geocentric orbit into a heliocentric one. This is because even when the Moon was as close as 10 $R_\oplus$, Earth's escape velocity was only 3.5 km sec^{-1} at the lunar orbit, scarcely bigger than the Moon's 2.4 km sec^{-1}. Hence, perhaps most of the material eroded from the Moon would eventually be collected by Earth, rather than reaccreted by the Moon.)

In Section IV, I used a fit to the lunar cratering record to argue that $\sim 1.0 \times 10^{20}$ kg of material was incident upon the Moon in the slowly decaying component of the heavy bombardment. Only about half this mass was accreted, since $\sim$50% of asteroids incident on the Moon are immediately lost in post-impact vapor plumes. Must the net amount of extralunar material added to the Moon be significantly further reduced due to erosion of the lunar surface in vapor plumes and ejecta?

A rough treatment of this problem, presented below, argues the answer is no. While a more sophisticated model could be considered, this seems unjustified in light of the preliminary nature of the impact erosion models themselves. Nevertheless, because of the uncertainties in all these models, the following conclusions should be treated with caution. Consider the result of a total mass M of asteroids incident on the Moon. Half this mass, $0.5M$, is accreted, while the remainder, $0.5M$, is lost. The latter fraction also results in $\sim 0.5M$ of lunar regolith lost in vapor plumes. Also, for an incident mass M, a regolith mass $\sim M$ is lost in high-speed ejecta. Therefore, the Moon experiences a net mass loss of $\sim M$. Subsequent to $\sim$4.4 Gyr ago, taking $M \approx 1 \times 10^{20}$ kg, this represents $\sim$1 km of material eroded from the lunar surface.

This result does not, however, contradict the presence of $\sim$2% extralunar siderophiles in the lunar crust. There are two reasons for this. The first is that, while a typical impactor, taken to be a CI chondrite, will deliver 100% CI siderophile abundances, the resulting eroded lunar regolith material will be only $\sim$2% CI material. Therefore the Moon will accrete a net mass $(0.5M\text{–}0.02M) \sim 0.5M$ of CI siderophile-abundance material—even as it experiences a net mass loss. That is, efficient regolith mixing [Sleep *et al.* (1989) take the extralunar material to be mixed through a megaregolith extending to a depth of 35 km] would guarantee that impact erosion will have little effect on extralunar abundances found in the crust.

A second reason that impact erosion by asteroid-like projectiles does not contradict regolith siderophile abundances is that I have so far neglected to take into account those lunar impactors that derive from the residual long-lived "tail" of bodies remaining in the region of the terrestrial planets after the growth of the Earth and Venus have been completed [$\sim$99%, say; see, e.g., Wetherill 1990]. These bodies will have much lower velocities at infinity than bodies in asteroid-type orbits; typical values of v_∞ for the former should be $\sim$8 km sec^{-1} (Wetherill 1977). The terrestrial impact rate of the residual Earth-crossers will initially decay with a 15- to 20-Myr half-life, but this half-life will lengthen after $\sim$100 Myr (Wetherill 1977). If 1% of the terrestrial mass remains to be accreted from this population subsequent to 4.5 Gyr ago, numerical experiments show some 0.02%, or $\sim 2 \times 10^{21}$ kg, would remain in Earth-crossing orbits $\sim$4.3 Gyr ago, of which $\sim$2% would impact the Moon (Wetherill 1980), giving $\sim 4 \times 10^{19}$ kg of material delivered to the Moon after 4.3 Gyr ago. This is approximately equal to the amount of accreted extralunar material implied by lunar regolith Ir abundances. Moreover, most of this material would have been accreted: $v_\infty = 8$ km sec^{-1} results in a lunar impact velocity of only 8.4 km sec^{-1}, which by Eq. (22) is too low for the formation of an escaping vapor plume. Equation (25) suggests that a typical impactor of mass m striking the Moon at 8.4 km sec^{-1} would have eroded a mass of less than $0.5m$ in ejecta: The residual Earth-crossers collected by the Moon would therefore have yielded a net accretion of lunar mass, for as long as this population persisted.

It appears extremely unlikely that residual bodies from the Moon-forming circumterrestrial disk (whether this disk was formed by a giant impact or otherwise) could have accounted for any significant influx of mass to the Moon subsequent to 4.4 Gyr ago. As Stevenson (1987) has emphasized, the characteristic sweep-up time for such a disk is extremely short, due to the rapidity of tidal orbital evolution of Moon-sized bodies near the Earth's Roche limit.

The story I have presented in this section is one in which two populations of objects, short-lived ($\sim$15–20 Myr half-life) residual Earth-crossers, and longer-lived ($\sim$100 Myr half-life) bodies which are eventually fed into orbits similar to those of Earth-crossing asteroids or short-period comets, both contribute to the heavy bombardment on the Earth and Moon. About 4.4 Gyr ago, both populations would have contributed roughly equally, but after only a few short (15–20 Myr) half-lives, the longer-lived population would dominate. This account is similar to that suggested by Hartmann (1980) and Grinspoon (1988). It is complicated by the fact that some fraction of the residual Earth-crossers will evolve into Mars-crossing orbits with ~200 Myr half-lives (Wetherill 1975). Therefore the various populations are not cleanly distinguishable by orbital decay times alone.

From the point of view of total mass accreted by the Earth and Moon, the origin of the impactors is of little importance. However, this question is of critical importance in estimating the quantity of volatiles or organics delivered to the early Earth. Throughout the remainder of my discussion, this uncertainty remains, and is fundamental. It could be (Wetherill 1980) that the bulk of heavy-bombardment impactors were residual bodies from the initial Earth- and Venus-forming swarm, passed through Mars-crossing storage orbits. If this were true, these bodies might deliver roughly chondritic abundances of refractory elements, while making little contribution to the terrestrial volatile inventory. On the other hand (see Section VII below), if a substantial fraction of these impactors were CI chondrites or comets, Earth could have thereby acquired the bulk of its surface volatile inventory.

There appears at present to be no decisive evidence for the composition of the population responsible for the heavy bombardment (Chyba 1987), beyond the inference from terrestrial noble metal abundances that the bulk of the population had roughly chondritic abundances of highly siderophile elements (Chou 1978). What inferences may be drawn from lunar geochemical evidence remains highly controversial. Some authors (Gros *et al.* 1976, Hertogen *et al.* 1977, Anders 1978) have argued that enrichment of siderophile trace elements in some lunar highland breccias indicates sampling of basin-forming projectile remnants, and that independent statistical tests agree in the identification of nine distinct meteoritic component clusters, corresponding to remnants of nine distinct parent bodies. These claims have been disputed on a number of grounds, including the possibilities that clusters identified by certain elemental ratios merely reflect indigenous lunar variations (Wänke *et al.* 1978, Delano and Ringwood 1978), procedure-dependent artifacts (Delano and Ringwood 1978), or regional heterogeneities due to large-scale impact gardening (Wetherill 1980). Rebuttals to most of these objections have been offered (Anders 1978). I am in no position to resolve this controversy. Here I wish only to emphasize that, if the nine putatively identified impactors are real, only one has a siderophile composition similar to CI chondrites, and even this cluster (the group 5L clasts) is depleted in volatiles relative to CI abundances by as much as $\sim 10^{-2}$. Therefore, if these data do represent the composition of nine large heavy-bombardment impactors, they imply that no more than a small percentage of this population could have been CI chondrites. [There seems to be no contradiction here with terrestrial upper mantle noble metal element ratios, which lie within the ranges found in the lunar highland breccias, and are closest to lunar group 1H (Chou, 1978).] It is important to note, however, that essentially no restriction is placed by the lunar highland samples on the fraction of the heavy bombardment that may have been cometary. Virtually all comets incident on the Moon should have been lost as vapor plumes expanding with velocities greater than v_{esc}, and therefore should have made little contribution to the extralunar component of highland rocks.

VII. IMPACT DELIVERY OF VOLATILES AND TERRESTRIAL ABUNDANCES

If Earth collected (1–4) $\times$ 10^{22} kg of extraterrestrial material subsequent to core formation, what quantity of accreted volatile elements is implied? This depends on the fraction of the heavy bombardment population composed of volatile-rich objects such as CI chondrites or comets. Table I shows two candidate heavy bombardment compositions which, given a net accreted mass of 2.5 $\times$ 10^{22} kg (chosen as the midpoint of the range allowed by the geochemical data), yield an approximately terrestrial abundance of water. Implied abundances of certain other volatiles are also listed. If the bulk of the Earth's oceans were derived from such a heavy bombardment source, would this imply a terrestrial abundance of any other element so far in excess of known terrestrial inventories as to rule the possibility out?

Column 4 of Table I considers the consequences if CI chondrites comprised 100% of the accreted mass. Wänke *et al.* (1984) have compiled a list of the abundances of some 56 elements in the terrestrial crust and mantle, based on geochemical data, and normalized these to CI abundances. Given the resulting numerical values for noble metals, their list may be readily scanned for any elements which are present in the terrestrial mantle + crust in CI-relative abundances below these values. Only these elements would be provided by a putative CI chondrite accretion in quantities in excess of the known terrestrial inventories.

In fact these limits are exceeded for almost no elements. All noble metal normalized abundances are within ~50% of the Ir value. In the Wänke *et al.* (1984) compilation, only selenium (Se), sulfur (S), and carbon (C) would appear to be delivered in quantities greater than the known terrestrial inventories for these elements if the heavy bombardment were 100% CI chondrite. A more recent estimate of the sulfur inventory on the Earth by these same authors (Dreibus and Wänke 1989) substantially increases the estimate of S in the terrestrial mantle, and the apparent discrepancy is virtually removed. Se is explicitly identified by Wänke *et al.* (1984) as an element whose abundance, inferred from mantle samples, is suspect. Excluding this element, only C is delivered by a 100% CI

TABLE I
Estimated Terrestrial Volatile Inventories vs Accretion from Candidate Heavy Bombardment Fractions

Element or compound	Terrestrial inventory (kg) in mantle and crust	Terrestrial surface inventory (kg)[e]	100% CI chondrite (H_2O content 6%)[f]	25% Cometary (H_2O content 50%)[g]
Carbon[a]	2.7×10^{20}	8.7×10^{19}	9.7×10^{20}	5.3×10^{20}
Nitrogen[b]	6.1×10^{19}	5.2×10^{18}	4.0×10^{19}	1.4×10^{20}
Sulfur[c]	1.4×10^{21}	2.1×10^{19}	1.5×10^{21}	8.8×10^{19}
Chlorine[c]	4.7×10^{19}	4.5×10^{19}	1.6×10^{19}	9.4×10^{17}
H_2O[d]	1.9×10^{21}	1.6×10^{21}	1.5×10^{21}	1.6×10^{21}

[a] C terrestrial abundance estimates in columns 2 and 3 from Turekian and Clark (1975) and Hayes *et al.* (1983), respectively.
[b] N terrestrial abundance estimates from Schidlowski *et al.* (1983).
[c] S and Cl terrestrial abundance estimates from Dreibus and Wänke (1989).
[d] H_2O terrestrial abundance estimates in columns 2 and 3 from Dreibus and Wänke (1989) and Turekian and Clark (1975), respectively. Terrestrial oceanic mass comprises 1.4×10^{21} kg H_2O (Walker 1977).
[e] Terrestrial surface abundances taken to include atmosphere, hydrosphere, sedimentary rocks, and crustal estimates.
[f] Chondritic abundances, exclusive of H_2O, from Dreibus and Wänke (1989). See text for discussion of CI H_2O abundances.
[g] Cometary abundances from Delsemme (1991); Cl value assumes CI chondritic S/Cl ratio. Only ~50% of cometary impacts occur at sufficiently low velocities for terrestrial accretion.

chondrite heavy bombardment in excess of its estimated terrestrial abundance, and only by a factor of ~3. However, estimates of terrestrial mantle abundances vary greatly from author to author. Along these lines, Dreibus and Wänke (1989) have argued that discrepancies of factors of 4 between predicted and known C and N inventories are "not really disturbing, considering the poor knowledge on the mantle and even the crustal concentrations of these elements," so it is unclear whether the discrepancies shown in Table I should be considered real. Column 3 of Table I lists estimates of terrestrial surface (atmosphere, hydrosphere, sedimentary rocks, and crust) volatile inventories. These estimates are, of course, on a farmer basis than the mantle estimates included in Column 2; the former therefore provide reasonably firm lower limits for the terrestrial inventories.

In addition to C and N, Table I also lists abundances for water, sulfur, and chlorine (Cl). The latter two elements are included as some previous authors (Clark 1987, Dreibus and Wänke 1989) have suggested using these elements to constrain the chondritic or cometary fraction of the heavy bombardment. Table I shows that neither S nor Cl terrestrial inventories are smaller than those expected to be delivered by either a comet- or CI asteroid-delivered ocean of water. I have not included noble gas abundances in Table I. While Sill and Wilkening (1978) have argued that cometary ice clathrate could explain the observed terrestrial abundances of Ar, Kr, and Xe, it appears that noble gas isotope ratios are incompatible with any simple "veneer" scenario (Pepin 1989). Evidently some form of mass fractionation must be invoked to explain these data (Pepin 1989, Zahnle *et al.* 1990), mechanisms that lie beyond the scope of this discussion.

Column 5 of Table I considers the possible role of a cometary fraction in the heavy bombardment. A comparison of cratering curves between the inner and the outer solar system seems to imply that comets did not comprise the bulk of the heavy bombardment (Strom 1987). Table I shows that comets could have supplied about a terrestrial ocean of water if they comprised as little as ~25% of the heavy bombardment. Note that Table I incorporates the result that only ~50% of comets incident on Earth actually contribute their volatiles, as the remainder are lost to space in post-impact vapor plumes. Given the way the heavy bombardment impactor population's mass peaks in the largest objects, comets could have comprised 25% of this population only if very large comets existed. However, some giant comets are known to exist: the trans-Saturnian comet Chiron appears to have a diameter of at least 180 km, and possibly as large as 372 km (Sykes and Walker 1991) [as well as an orbit that may evolve into an Earth-crossing one (Hahn and Bailey 1990)], and the Great Comet of 1729 was a naked-eye object at a perihelion of about 4 AU (Kronk 1984).

The masses of accreted volatiles in Table I are calculated subsequent to 4.4 Gyr ago. What of volatiles delivered prior to this time? [For example, Hartmann (1987, 1990) has argued that the first $\leq 10^8$ years of solar system history witnessed an intense scattering of C asteroid materials, as evidenced by the fact that most solar system satellites thought to be capture appear to be spectral class C (Hartmann 1987), and by the preponderance of CM clasts among the foreign fragments in polymict meteoritic breccias (Hartmann 1990).] Terrestrial atmospheric volatiles present prior to the hypothesized Moon-forming impact might be lost in this event (Cameron 1986). Moreover, terrestrial water prior to core formation (~4.4 Gyr) should have been efficiently destroyed by reacting with metallic iron: $Fe + H_2O = FeO + H_2$;

large quantities of H_2 produced in this way may have removed other degassed volatiles by hydrodynamic escape (Dreibus and Wänke 1987, 1989). Some volatiles may also have been incorporated into Earth's core. In any case, the 100-Myr half-life of the slowly decaying population means that even if we were to integrate back to the time of Earth's accretion, our results would change by only about a factor of two.

VIII. UNCERTAINTIES IN CI CHONDRITE WATER CONTENT

It is a surprising fact of the carbonaceous chrondrite literature that water abundances (wt%) commonly cited for CI chondrites vary by a factor of nearly three (~7% vs ~20%), and that one or the other value is typically cited without reference to the other. Both results derived from experiments done in Harold Urey's laboratory at the University of Chicago in the mid 1950s, first by Boato (1954), then by Wiik (1956). Kaplan (1971) has summarized and briefly compared the two sets of results.

Wiik (1956) reported H_2O abundances for three CI chondrites. Two of these determinations were his own; one was taken from Christie (1914). Each determination yielded ~20% H_2O. Mason's (1971) commonly referenced *Handbook of Elemental Abundances in Meteorites* lists the mean value of these three measurements, 20.08%. Many recent textbooks on meteorites (e.g., Dodd 1981, Wasson 1985) follow the petrologic classification of Van Schmus and Wood (1967), which cites Wiik (1956) and gives the bulk water content of CI chondrites as "~20%." In previous work (Chyba 1990), I used a carbonaceous chondrite bulk water content given by the average of Mason's (1971) values for CI and CII chondrites.

However, 2 years prior to Wiik (1956), Boato (1954) had published substantially lower H_2O abundances for some eight CI chondrites. Yet the two CI chondrites analyzed by Wiik, Orgueil and Ivuna, were two of the eight used by Boato. For Orgueil and Ivuna, the latter author had found net water abundances of 11.3 and 11.7%, but argued convincingly on the basis of deuterium/hydrogen (D/H) ratios found at different heating steps that 4.0% and 4.7%, respectively, of this water was terrestrial contamination. Wiik (1956) noted the discrepancy between his and Boato's abundances, saying that, "the reason for the disagreement of the results is not clear." However, Wiik's (1956) experimental technique apparently neither employed stepwise heating, nor tested D/H ratios. Nevertheless, his compilation included Christie's (1914) determination that water in the Tonk CI chondrite was half lost after drying at 106°C, suggesting that ~50% of Wiik's measured water content might be due to adsorption of terrestrial water. On the other hand, Boato's highest experimental temperatures (800–900°C) may have been insufficient to release all hydrogen bound in hydrated silicates (Friedman and O'Neil 1978). Dreibus and Wänke (1987, 1989) have advocated an H_2O abundance for CI chondrites of either 7.15% (1987) or 7.29% (1989), following Boato's (1956) results.

Kerridge (1985) has summarized more recent determinations of hydrogen in some 30 carbonaceous chondrites. Kerridge's experimental protocol first crushes and outgasses samples at 200°C overnight to eliminate hygroscopic water. This includes water adsorbed by the sample subsequent to its fall; however, it might also include some water indigenous to the meteorite (J. Kerridge, personal communication 1991). Therefore Kerridge's (1985) results may best be viewed as lower limits to CI water content. Averaging the results from the CI chondrites Orgueil and Ivuna gives 0.66% H, or 5.9% H_2O. Attempting to subtract the effect of H originally present in organics reduces this figure to about 4.7% (J. Kerridge, personal communication 1991). On the other hand, in an impact, much of the organic H might be pyrolized into H_2O.

For definiteness, Table I uses a CI water abundance of 6%. If meteorites reach Earth with substantial hygroscopic water content, this could be a considerable underestimate. Since CI and CM chondrites consist mainly of a variety of clay-like hydrous sheet silicates (Sears and Dodd 1988), a reasonable upper limit to CI water abundance should be given by considering the water content of terrestrial clays (J. Cronin, N. Sleep, personal communications 1991). Water abundance in clay minerals ranges as high as ~25%; hygroscopic (H_2O^-) and combined (H_2O^+) water content are known to range individually as high as ~16% and ~18%, respectively, depending on the sample (Grim 1968). Several CM chondrites (Kerridge 1985) have H abundances that imply they are ~9% H_2O^+.

IX. IMPACT EROSION OF VOLATILES

Melosh and Vickery (1989) have proposed a simple analytical approximation to their more detailed numerical work treating impact erosion of planetary atmospheres. Their treatment indicates that an atmospheric cap of mass

$$m_{\text{cap}} = 2\pi P_0 H R_\oplus / g \tag{28}$$

above the plane tangent to the point of impact will be removed by impactors with masses above some threshold, m_*. In Eq. (28), P_0 is the terrestrial surface atmospheric pressure, H is the atmospheric scale height, 8.4 km for the contemporary Earth (Walker 1977), and g is terrestrial surface gravity. The threshold mass $m_* \approx m_{\text{cap}}$, which for the contemporary 1-bar terrestrial atmosphere is 3.5×10^{15} kg.

It is easy to show from Eqs. (9) and (28) that the heavy bombardment should have resulted in substantial atmospheric erosion for the Earth. With m in Eq. (9) set equal to m_* from Eq. (28), scaling to the Earth gives $n(> m_*, 4.4 \text{ Gyr}) = 1.3 \times 10^4$. For the purposes of illustration, take comets to have represented a negligible fraction of the impactors, so that only ~10% of the latter had sufficient velocity to cause erosion. Conservatively, then, $\sim 10^3$ atmosphere-eroding impacts occurred on Earth subsequent to 4.4 Gyr ago. Again for illustration, take the mass of the early atmosphere to equal that of the contemporary one, 5.3×10^{18} kg (Walker 1977). Then each eroding impact removed about 6.6×10^{-4} of the total atmospheric mass. Multiplying this result by 1.3×10^3 eroding impacts yields about 0.9 atm of mass eroded by the heavy bombardment. A denser atmosphere would lead to more eroded atmospheric mass.

To treat the problem more quantitatively, the net effect of erosive impacts during the heavy bombardment may be calculated from integrals similar to Eq. (10), of the form

$$M(t) = \int_{m_{\max}}^{m*} Q(m)[\partial n(> m, t)/\partial m]dm. \quad (29)$$

Erosion of condensed oceans appears to be an unimportant effect, amounting to $\lesssim 10\%$ of the mass of the ocean (Chyba 1990). Equation (29) gives the mass of atmosphere eroded, by setting $Q(m) = fm_{\text{cap}}$, where m_{cap} is from Eq. (28), and f is that fraction of the impactors that causes erosion. Performing the integral gives

$$M_{\text{lost}} = f\alpha[t + \beta(e^{t/\tau} - 1)]m_{\text{cap}}[m(4 \text{ km})/m_*]^b \text{ km}^{-2}. \quad (30)$$

A long-standing concern about cometary volatile delivery scenarios is that a cometary source for Earth's water may imply an abundance of nitrogen in excess of known terrestrial inventories (C.P. McKay, personal communication 1989, Chyba 1990). The gravity of this concern remains unclear in light of vast disagreement over the size of the relevant terrestrial inventories. For example, Turekian and Clark (1975) suggest 1.1×10^{19} kg N resides in the Earth's atmosphere, crust, and upper mantle. Dreibus and Wänke (1989) tabulate the terrestrial mantle N inventory as "?." Schidlowski *et al.* (1983) give as their preferred mantle estimate 5.6×10^{19} kg N (used in Table I), but cite other estimates as high as 2.0×10^{20} kg. Here I intend simply to show, using Eq. (30), that impact erosion could remove a substantial quantity of whatever "excess" N may have been delivered.

Consider, then, the case of cometary delivery of the bulk of terrestrial volatiles (Column 5, Table I). About half of the 25% of the collisions that were cometary, and about 10% of the remaining collisions, should have caused erosion of the atmosphere for masses $> m_*$. Hence $f \approx 0.2$ in Eq. (30). I consider a very simple model in which half of all the N delivered by comets remains in the early terrestrial atmosphere. Obviously, however, the exact quantity eroded depends on details of the early terrestrial atmosphere, in particular what fraction of N delivered to the early Earth remained as atmosphere N_2 (or ammonia, NH_3), and what fraction was incorporated into the sedimentary column. Detailed models could be pursued, although so much is uncertain about the early terrestrial atmosphere (see, e.g., Walker 1986) that it seems unwise to consider any but the simplest models. In any case, my purpose here is only to demonstrate the existence and possible magnitude of an effect.

To approximate cometary delivery and simultaneous atmosphere erosion, I have used Eq. (30), with m_* and m_{cap} scaled by Eq. (28) to an atmosphere with a mass equal to one in which half of the total nitrogen delivered, 7×10^{19} kg, is present in the atmosphere. Equation (30) then shows that some 3×10^{19} kg of atmospheric N_2 may thereby be removed from Earth by impact erosion. This represents about one-half of the "excess" cometary nitrogen predicted by comparing Column 5 with Column 2 of Table I. Considering the uncertainties of the various factors in Eq. (30), this result is perhaps fairly summarized by saying that atmospheric impact erosion can remove a quantity of N_2 about equal to that which is delivered in excess. Nevertheless, this result is far from a demonstration that erosion of exactly that needed to reconcile nitrogen delivery by comets with the (uncertain) terrestrial inventory did, in fact, occur.

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