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## Technical Memorandum **80251**

(NASA-TM-80251) CRUSTAL EVOLUTION OF THE  
EARLY EARTH: THE ROLE OF MAJOR IMPACTS  
(NASA) 36 p HC A03/MF A01 CSCI 08G

N79-23560

Unclas  
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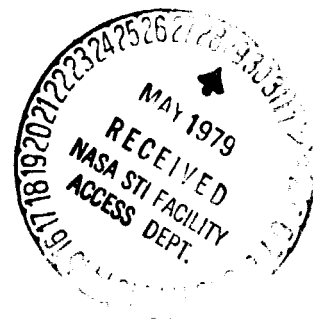
# Crustal Evolution of the Early Earth: The Role of Major Impacts

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**MARCH 1979**

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**CRUSTAL EVOLUTION OF THE EARLY EARTH:  
THE ROLE OF MAJOR IMPACTS**

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## ABSTRACT

Impact cratering was an important--even dominant--process affecting the crustal evolution of the small terrestrial planets. The fundamental highlands/maria dichotomy of the Moon's surface can be traced to a late heavy bombardment by basin-forming, asteroid-sized bodies which produced not only a topographic division in the lunar crust also localized the later eruptions of mare basalts. Major impact basins with diameters in excess of 200 km are recognized throughout the inner solar system from Mars to Mercury. Similar craters must have formed on the Earth prior to 4 billion years ago, and the minimum number of such basin-forming impacts can be calculated by scaling from the observed (minimum) number preserved on the Moon. When allowance is made for differences in impact velocity, gravitational cross-section and the effects of gravity on crater diameter, it is found that at least 50% of a presumed global sialic crust would have been converted into impact basins by 4 billion years ago. Among the effects resulting from the impact of an asteroidal object on the early crust were: (a) establishment of a topographic dichotomy of 3-4 km (after isostatic adjustment), (b) pressure-release partial melting of the upper mantle and rapid flooding of the basin floor by basalt, and (c) enhancement of thermal gradients in the sub-basin lithosphere and upper asthenosphere. Comparative planetary data such as impact scaling can be used as important constraints on models of the early terrestrial crust. For example, the topography resulting from impact bombardment produced discrete oceans and dry land by 4 billion years ago, making unreasonable models of a globe-encircling ocean on the Earth after that time.

## INTRODUCTION

The understanding of the present nature of the continental crust of the Earth requires knowledge of the evolution of the crust through time, as well as information on the composition and structure of the original crust. While the recent evolution of the lithosphere may be described reasonably accurately, the accuracy of that description decreases rapidly the further into the past we look. In particular, the scarcity of rocks on the Earth older than ~3.5 billion years makes models of the earliest Archaean based only on terrestrial data highly speculative at best. Yet this is the very time when so many of the processes which today dominate the Earth's development were becoming established. Those processes have been sufficiently active that older rocks were obliterated as newer ones were produced. It is extremely unlikely that the clues necessary to understand the earliest Earth will be found only on this planet.

But important constraints on modeling the early Archaean are available. The smaller planets are generally less active and have experienced more sluggish evolution than the Earth; they have retained a greater fraction of the oldest rocks on their surfaces. In general evolutionary activity and complexity increases with size of the planet, because of the greater thermal reservoir available for those bodies. When allowance is made for differences in size, mass, bulk composition and location in the solar systems, studying the crustal evolution of the smaller terrestrial planets offers important information about the crustal evolution of the Earth as well (Frey and Lowman, 1978).

This is particularly true for early cratering effects. The Moon, Mercury, Mars and Venus all retain, to some degree, ancient cratered terrains. The tremendous geologic and meteorologic activity of the Earth erase traces of most craters within a few million years, and there is no evidence for any large craters dating from the early Archaean (Grieve and Robertson, 1978). Yet cratering was a fundamental, even dominant process throughout the inner solar system during the early stages of crustal formation. In this paper the role of major impact basins (such as those which formed on the Moon before 4 billion years ago) is examined to determine the effects of such impacts on the early crustal evolution of the Earth.

Specifically we address the following fundamental problem: what is the origin of the Earth's fundamental crustal dichotomy of low density continental and high density oceanic crust? Is this in any way related to the superficially similar highlands/maria crustal dichotomies of the Moon, Mercury and Mars? The lunar two-fold crustal division can be traced to late heavy bombardment by basin-forming, asteroid-sized objects. These impacts not only produced a topographic dichotomy, but localized the later eruptions of mare basalts which produced a compositional dichotomy in the lunar crust. Is a similar series of events likely for the early Earth?

The formation of the Earth's original crustal dichotomy is obscure. Today the crustal dichotomy is maintained by plate tectonic processes, but these same processes were not likely to have produced the original separation of high and low density crust. Comparative planetary data (Lowman, 1976) as well as terrestrial data (Shaw, 1972, 1976; Hargraves, 1976) suggest that

originally the Earth was covered by a thin, low density, sialic crust. If true, plate tectonic processes could not have converted some 60% of that crust into high density "oceanic" crust because of the difficulty of subducting continental lithosphere once it has formed (McKenzie, 1973; Molnar & Gray, 1979). This especially true for the early Earth when the lithosphere should have been thinner. It can be argued (Frey, 1977a) that the crustal dichotomy must have been established first; that is, subductable crust must have been present before large scale horizontal displacement of the lithosphere was possible. But plate tectonic processes may have been active for much of the Earth's history (Burke et al., 1976; McKenzie and Weiss, 1975). Therefore it becomes desirable to look at the crustal evolution of the earliest Earth for the origin of the crustal dichotomy; that is, before 4 billion years ago, when cratering was still a very significant mechanism of crustal modification.

#### BASIN-FORMING IMPACT CRATERING

Large impact basins with diameters greater than 200 km are recognized throughout the inner solar system, being well displayed in the ancient highland crusts of the Moon, Mercury and Mars (Wood and Head, 1976; Frey and Lowry, 1979). It would be impossible for the Earth to escape this bombardment which produced the lunar Imbrium and Orientale basins before terminating about 4 billion years ago. This remains true regardless of whether the bombardment was part of the late accretional sweep-up (Wetherill, 1976) or part of a special flux of objects (Chapman, 1976). Dynamically plausible orbits exist for both cases (Wetherill, 1975, 1976). In this paper we do not consider whether the period

of basin-forming impacts was a discrete "event" in the inner solar system, but simply calculate the cumulative effects of such a bombardment until its termination 4 billion years ago. Because these calculations are scaled from observed lunar impact basins, they represent a minimum estimate of the total effects on the Earth. Such a scaling is reasonable because the Earth and Moon would see approximately the same incoming flux for most likely impacting bodies from outside the Earth-Moon system.

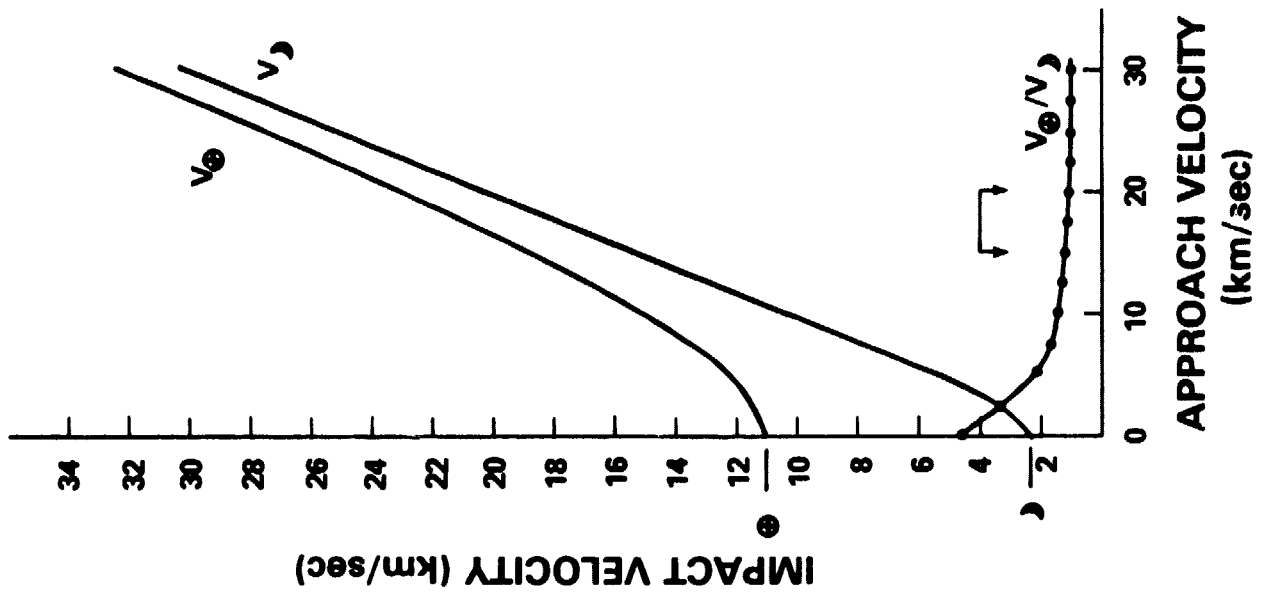
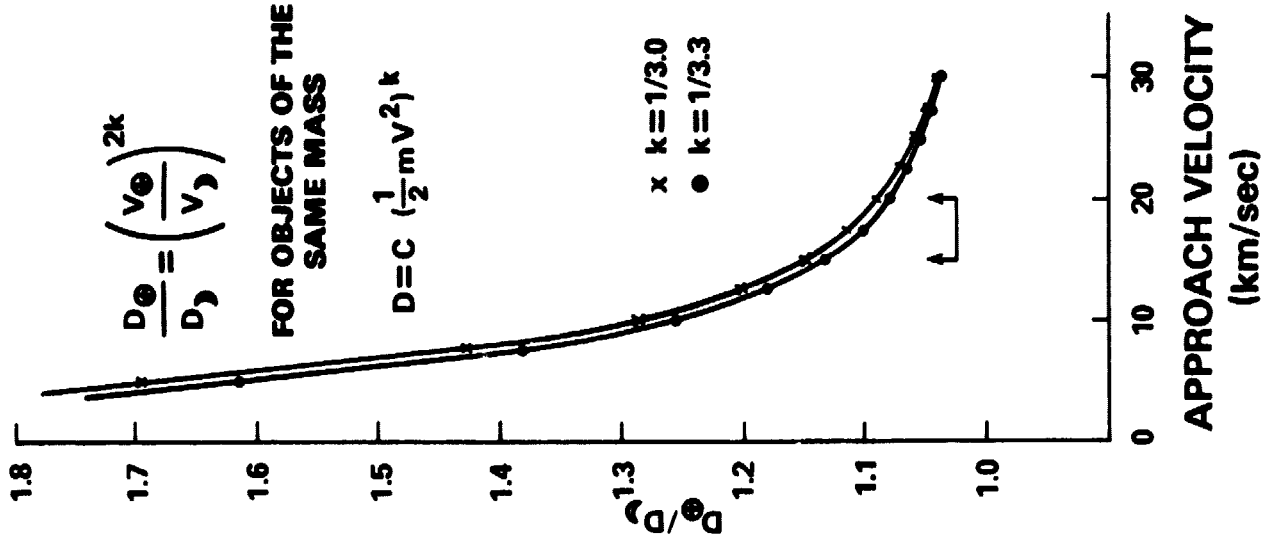
Consider objects in Apollo asteroid-type orbits, which are reasonable modern analogs of the bodies responsible for the late heavy bombardment of the Moon (Wetherill, 1975). These objects approach the Earth-Moon system with a relative velocity between 15 and 20 km/sec (Öpik, 1966). Impact velocities (Figure 1a) at the surface of the Earth will range from 18.7-22.9 km/sec, and at the lunar surface the objects will hit at 15.2 to 20.1 km/sec. The 15-20% greater impact velocity at the surface of the Earth is due to the greater gravitational acceleration of the Earth. Gravitational field differences have additional effects, as described below.

Faster impacts produce larger impacts on any planet. Impact velocity is generally related to crater diameter through an energy scaling law

$$D = CE^k = C \left(\frac{1}{2}mV^2\right)^k \quad (1)$$

where D is the crater diameter, E the energy released by the impact and taken to be equal to the kinetic energy of the impacting body, C and k are constants. C is not well determined, but, if the same for the Earth and Moon, then the ratio of diameters of impact basins formed on the Earth to those formed on the Moon by objects of equal mass can be written





$$\frac{D_E}{D_M} = \left( \frac{V_E}{V_M} \right)^{2k} \quad (2)$$

Where  $V_E$  and  $V_M$  are the impact velocities at the surface of the Earth and Moon, respectively. The constant  $k$  is also uncertain. Values between 1/3 and 1/4 have been suggested, but most authors adopt 1/3.3 or 1/3.4 (see Hartmann, 1965, 1977; Dence et al., 1977; Dence, 1978). In reality, the ratio above is not very sensitive to  $k$ , as shown in Figure 1b (Frey, 1977b). Here 1/3.3 is used. For approach velocities of 15-20 km/sec, craters on the Earth might be expected to be some 11-15% larger than those formed on the Moon by identical objects. In fact the basins formed on the Earth may be somewhat smaller than expected due to greater gravitational confinement of the cavity on larger planets. This can be represented by a gravitational acceleration ratio (e.g., Hartmann, 1977)

$$\frac{D_E}{D_M} \propto \left( \frac{g_E}{g_M} \right)^{-z} \quad (3)$$

The value of  $z$  is uncertain. Hartmann (1977) adopted 0.2, but Dence et al. (1977) used a value of about 0.1. Below we treat three cases, retaining  $z$  as a variable yet to be determined.

Therefore the diameter ratio of basins on the Earth and Moon is

$$\frac{D_E}{D_M} = \left( \frac{V_E}{V_M} \right)^{2/3.3} \left( \frac{g_E}{g_M} \right)^{-z} \quad (4)$$

Here we are concerned with the area of impact basins formed on the Earth compared with the area of lunar impact basins. The ratio  $(D_E/D_M)^2$  is plotted in the bottom portion of Figure 2a as a function of approach velocity, for

$z = 0, 0.1$  and  $0.2$ . At  $15 \text{ km/sec}$  approach velocity the area ratio of basins ranges from  $1.29$  ( $z = 0.0$ ) to  $0.63$  ( $z = 0.2$ ). At  $20 \text{ km/sec}$ , basins on the Earth would be  $1.17$  ( $z = 0.0$ ) to  $0.57$  ( $z = 0.2$ ) times as large as those formed on the Moon by identical objects. This clearly shows the importance of the gravitational acceleration term  $z$ : for  $z = 0.2$ , terrestrial basins are much smaller than corresponding lunar basins despite the greater energy involved in their formation.

One other important effect must be considered. Because of its larger gravitational cross section, the Earth will collect many more of the incoming bodies. This results in a much larger number of impacts per unit area of all sizes. This in fact is the dominant effect in these scaling relations, as shown elsewhere (Frey, 1977b). The gravitational radius

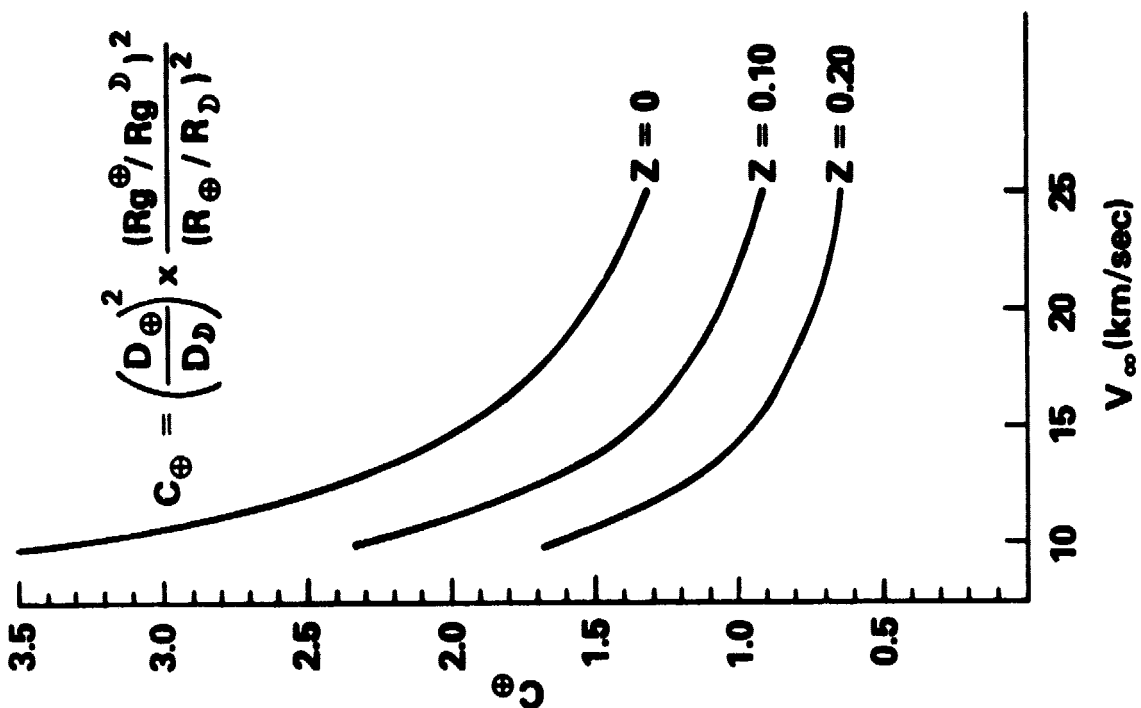
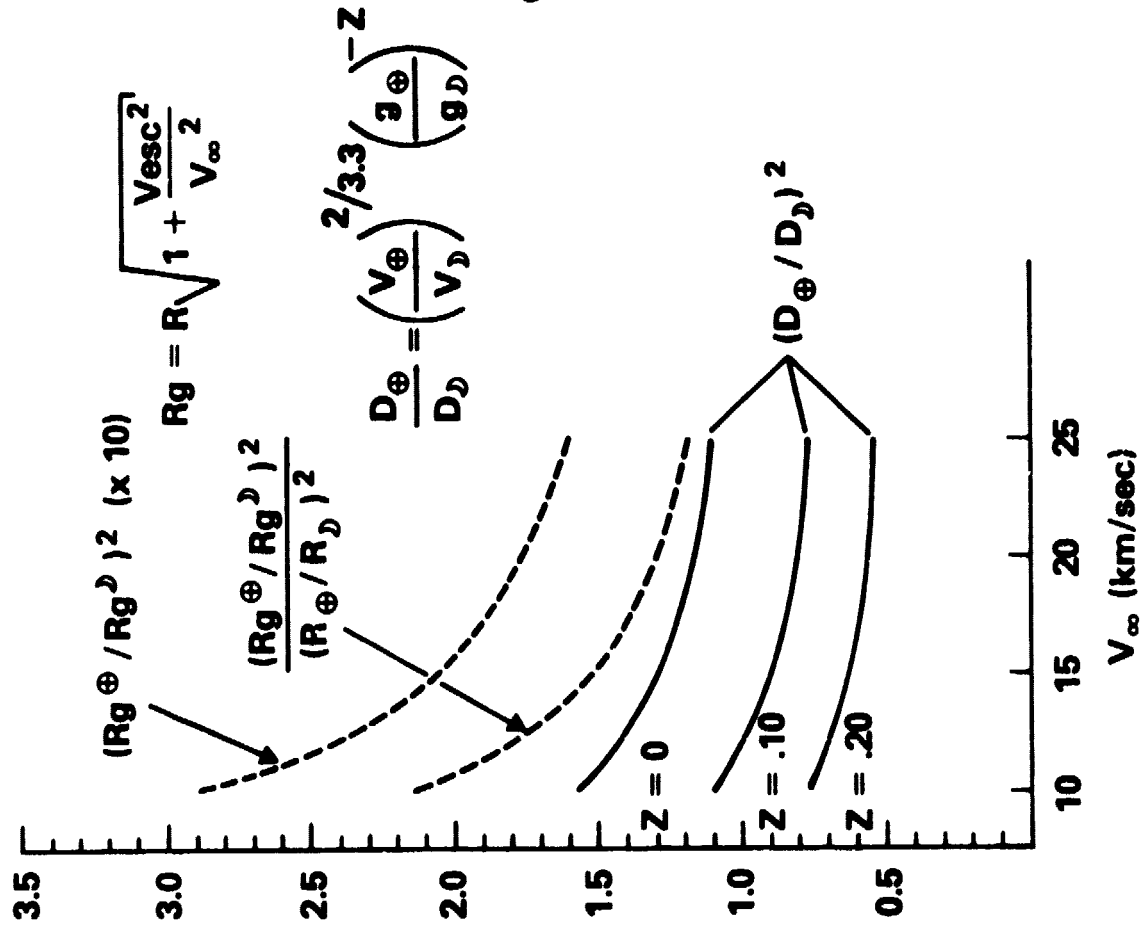
$$R_g = R \sqrt{1 + \frac{V_{esc}^2}{V_{\infty}^2}} \quad (5)$$

where  $R_g$  is the gravitational radius,  $R$  the physical radius,  $V_{esc}$  the escape velocity, and  $V_{\infty}$  the approach velocity. The ratio of the gravitational cross sections of the Earth and Moon is shown in Figure 2a (times 10), as is the ratio normalized by the physical cross-section ratio of the two planets.

This latter term is the relevant one:

$$\frac{(R_{g\oplus} / R_{g\text{Moon}})^2}{(R_{\oplus} / R_{\text{Moon}})^2} \quad (6)$$

as it shows, as a function of approach velocity, the ratio of the number of impacts per unit area that will occur on each body. For approach velocities between  $15\text{-}20 \text{ km/sec}$ , the above term ranges from  $1.52$  to  $1.30$ , decreasing with increasing velocity. That is, over the approach velocities of interest, the



Earth collects 1.3 to 1.5 times as many craters per unit area as does the Moon.

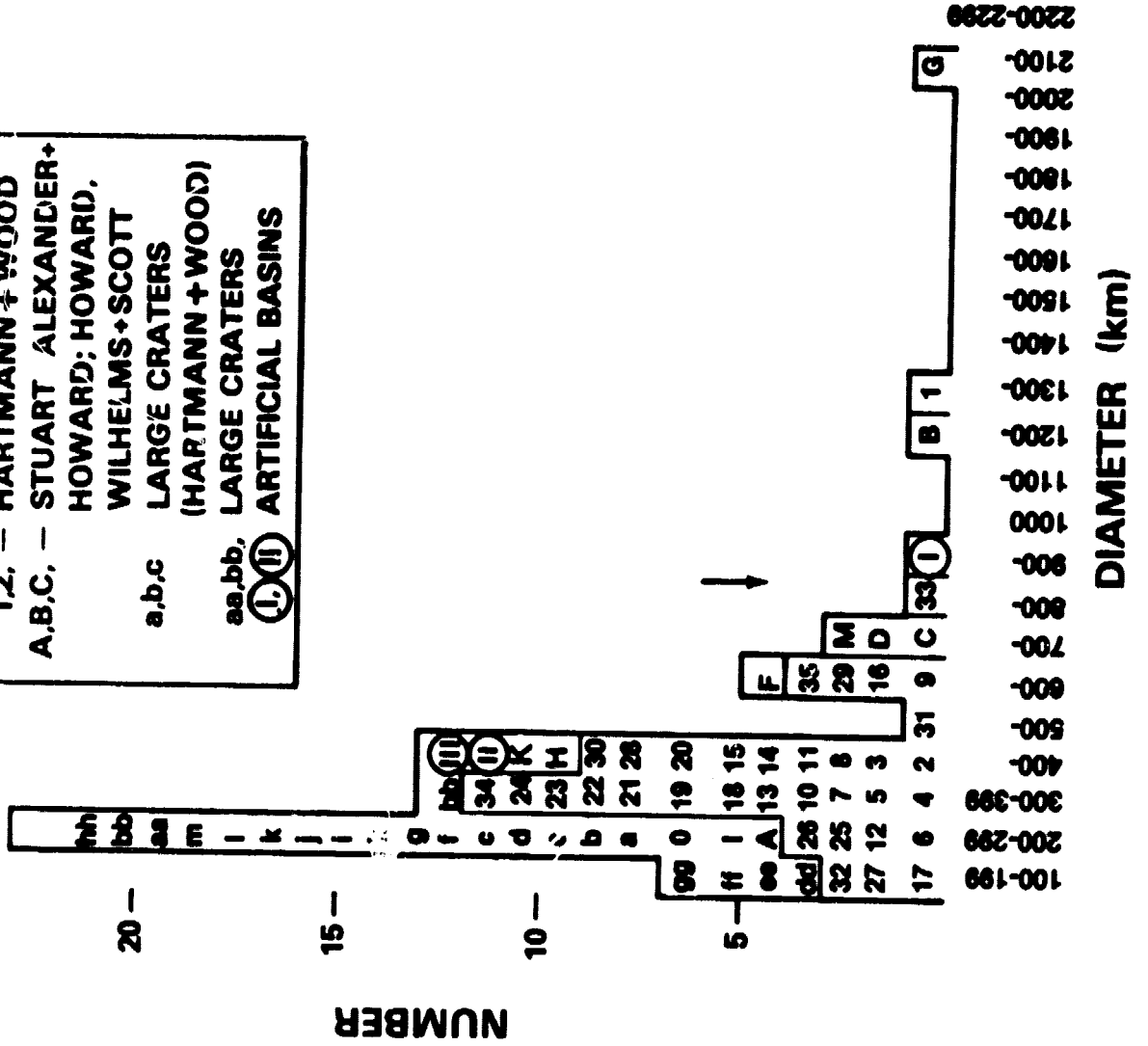
The effects of basin diameter and basin density may be combined into a single conversion factor  $C_{\oplus}$

$$C_{\oplus} = \left(\frac{D_{\oplus}}{D_{\text{J}}}\right)^2 \times \frac{(R_{\text{g}\oplus}/R_{\text{g}\text{J}})^2}{(R_{\oplus}/R_{\text{J}})^2} \quad (7)$$

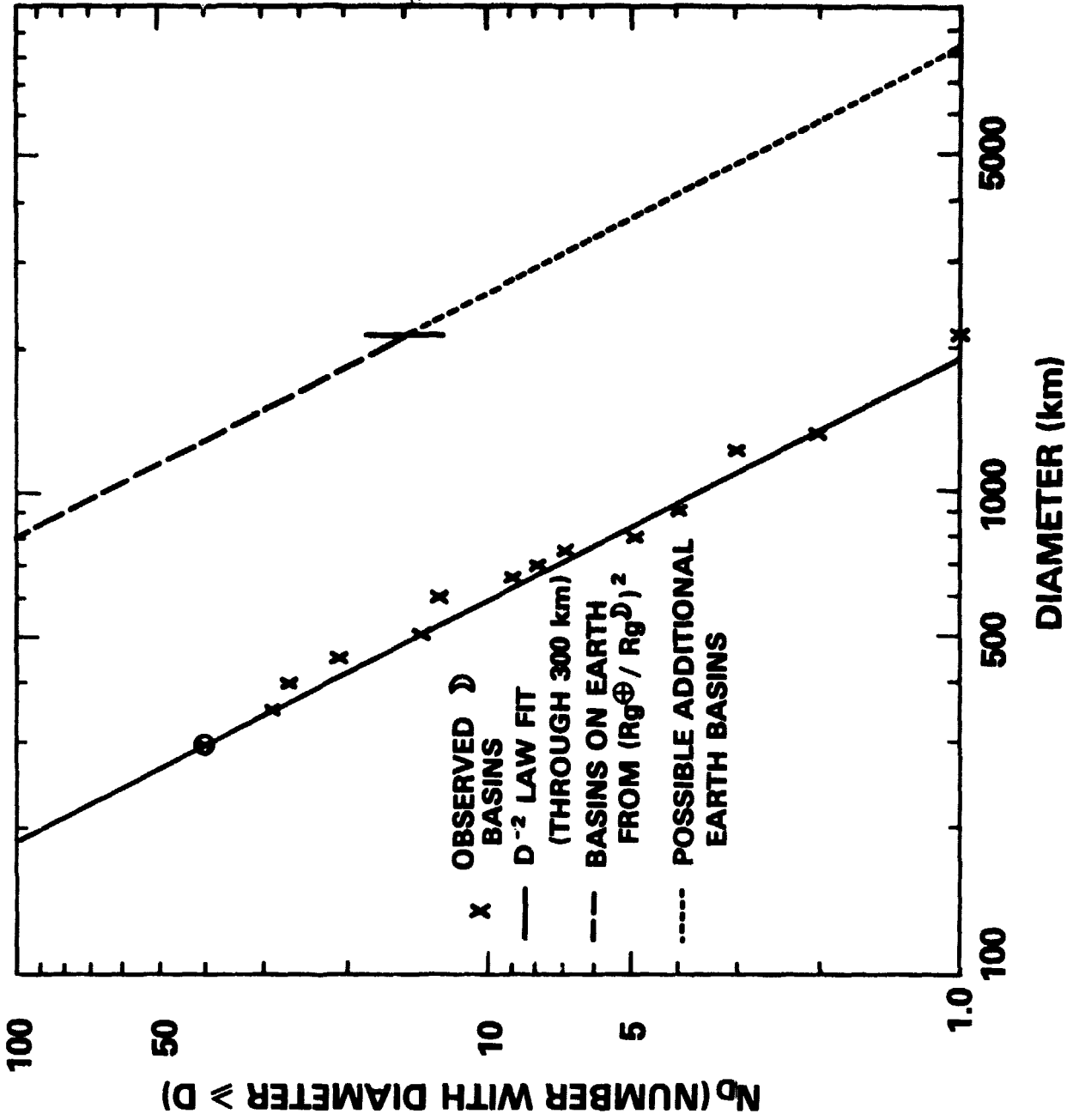
This is shown in Figure 2b, as a function of approach velocity, for three values of  $z$ . In this form it is possible to calculate the equivalent area of basins formed on the Earth (in units of the surface area of the Earth) directly from a knowledge of the total area of lunar basins, normalized by the lunar surface area. This is referred to as "equivalent basin area" and is denoted by  $a_{\oplus}$  or  $a_{\text{J}}$ . Note that  $a$  is not the total surface area of a planet covered by basins, but the total area of all basins (normalized by surface area). In determining the total surface area covered by basins, allowance for overlap must be made, as described below.

In principle all that is needed is knowledge of  $a_{\text{J}}$ , the equivalent area of all lunar basins in units of the Moon's surface area. Figure 3 is a histogram of all lunar basins larger than 200 km (Frey, 1977b). The total number of such basins is larger than that given by Wood and Head (1976), who restricted their count to multiringed and peaked ringed basins. The histogram above does not include Mare Gargantuan, and so should be considered a minimum for the number of lunar basins. The total area of these is equivalent to  $0.40 A_{\text{J}}$ , where  $A_{\text{J}}$  is the surface area of the Moon. If Mare Gargantuan (Wood and Head, 1976) is included the equivalent basin area becomes  $0.50 A_{\text{J}}$ . It is not unreasonable that buried basins may raise this figure even higher, perhaps to  $0.60 A_{\text{J}}$ .

1,2. - HARTMANN + WOOD  
 A,B,C. - STUART ALEXANDER + HOWARD;  
 HOWARD; HOWARD.  
 WILHELMS + SCOTT  
 a,b,c LARGE CRATERS  
 (HARTMANN + WOOD)  
 aa,bb, LARGE CRATERS  
 (I, II) ARTIFICIAL BASINS



If the observed number of lunar basins are scaled to estimate the number of basins forming on the Earth, the terrestrial basins will be an under-estimate of the number that probably formed. The reasons for this are two. First, as discussed above, the actual number of basins that formed on the Moon is almost certainly larger than the number preserved. Secondly, there is the distinct possibility that the Earth sustained even larger impacts than those seen on the Moon. As described elsewhere (Frey and Lowry, 1979) the lunar basins, when plotted on a  $\log(\text{cumulative number}/\text{km}^2) - \log(\text{diameter})$  graph, closely follow a line with slope = -2 (for  $D \geq 300$  km). If we assume a  $D^{-2}$  distribution of basins on the Earth as well, then it is possible for a number of very large basins to have formed before 4 billion years ago. Table I shows this. If there are 3 basins on the Moon with diameter equal to or greater than 1000 km, then there should have been some 56 basins formed on the Earth larger than or equal to 1000 km diameter. This in turn "predicts" 6 basins larger than 3000 km, 2 larger than 5000 km, and 1 larger than 7000 km, according to a  $D^{-2}$  law. If there are 4 lunar basins larger than 1000 km, then the Earth might have had 8 larger than 3000 km and 3 larger than 5000 km. These are approximate values, and of course depend on the value of  $C_{\oplus}$  adopted (that is, on the ratio of basin areas on Earth and Moon and the number per unit area ratio for the Earth and Moon). Figure 4 shows those basins which might be ignored unless this effect is taken into account. The solid line is a  $D^{-2}$  fit through  $D=300$  km on the  $\log(\text{cumulative number})$  vs  $\log(\text{diameter})$  plot. The long dashed line is the corresponding  $D^{-2}$  distribution expected for the Earth, for  $C_{\oplus} = 1.4$  (see below). This corresponds to a value of  $z = 0.1$  at  $V_{\infty} = 15$  km/sec. The



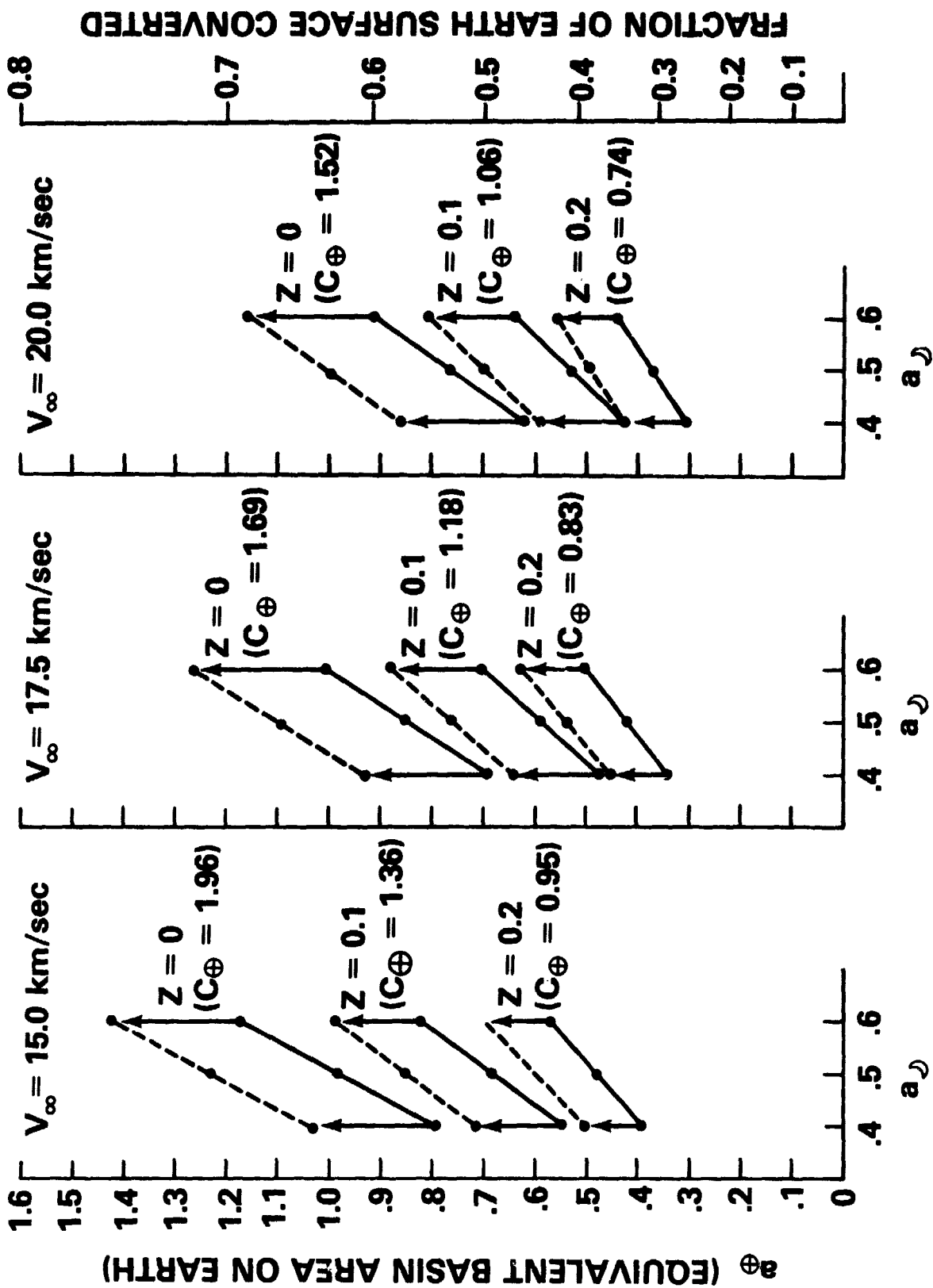


lunar distribution cuts off at 2100 km; there could be 6 basins larger than this formed on the Earth as shown by the extrapolation of the Earth  $D^{-2}$  line (short dashes). The total area of these "extra" basins is some  $1.23 \times 10^8 \text{ km}^2$ , or roughly  $0.24 A_{\oplus}$ , where  $A_{\oplus}$  is the surface area of the Earth. This compares with the total area of about  $0.55 A_{\oplus}$  for all the basins with  $D \leq 2100 \text{ km}$  that are expected to have formed on the Earth based on the number seen on the Moon. It is clear that these "extra" basins represent a significant contribution to the total equivalent area of terrestrial basins.

The above effects are combined in Figure 5. Three cases are shown, for three values of the approach velocity (15.0, 17.5 and 20.5 km/sec). Approach velocity contributes both to impact velocity (and therefore the individual basin area ratio  $[D_{\oplus}/D_{\text{L}}]^2$ ) and to the gravitational cross-section ratio  $(Rg^{\oplus}/Rg^{\text{L}})^2$ . Consider the middle panel, for  $V_{\infty} = 17.5 \text{ km/sec}$ . In this (and the two adjacent panels) the horizontal axis is  $a_{\text{L}}$ , the equivalent lunar basin area (total area of all lunar basins divided by surface area of the Moon) As described above, this value is at least 0.4 the lunar surface area (although basin overlap reduces the actual surface area covered by basins to about  $0.32 A_{\text{L}}$ ). The vertical axis is the equivalent basin area on the Earth  $a_{\oplus}$ . That is,  $a_{\oplus}$  is the total area of all basins produced on the Earth by scaling from lunar basins. The minimum value of this is given by

$$a_{\oplus} = C_{\oplus} a_{\text{L}} \quad (8)$$

where  $C_{\oplus}$  is the conversion factor shown in Figure 2b.  $C_{\oplus}$  of course depends on the gravitational factor  $z$ . For  $z = 0$ , the diameters of terrestrial basins are much larger than those formed on the Moon by identical objects (see Figure 2a). Thus  $C_{\oplus}$  is a maximum of this value (1.69). For  $Z = 0.2$ ,  $C_{\oplus}$  is reduced to 0.83, because terrestrial basins have only 60% as much area as the corresponding lunar basins (see Figure 2a).



Equation 8 above assumes no terrestrial basins larger than the largest lunar basin formed on the Earth. For this case, the calculated value of  $a_{\oplus}$  as a function of  $a_{\text{J}}$  is given by the solid line in Figure 5. However, "extra" basins may have formed on the Earth due to the large diameter fact in the  $D^{-2}$  distribution (Figure 4). The elevated dashed line in Figure 5 accounts for these extra basins, whose number and size depend on the approach velocity and the assumed value of  $z$ . The true  $a_{\oplus}$  probably lies somewhere between the solid and dashed lines, for any given value of  $a_{\text{J}}$ . That is,  $a_{\oplus}$  is probably larger than that given by the lunar size distribution alone, but is probably less than the maximum possible using a pure  $D^{-2}$  law for terrestrial basins.

In fact, the value of  $a_{\oplus}$  is not enough. This only represents the total area of basins formed on the Earth. A more significant quantity is the total fraction of the Earth's surface converted into basins,  $A$ . From the probability of each new impact basin forming in an uncratered area, it can be shown (Frey, 1977b) that

$$A = 1 - e^{-a_{\oplus}} \quad (9)$$

The scale for  $A$ , in units of the Earth's surface area, is shown on the right. The scale is compressed at the bottom as would be expected: as the total area (or number) of basins increases, it becomes harder and harder to find uncratered areas, so the total fraction of the Earth's surface not cratered decreases more slowly. That is, most impacts occur in areas already cratered.

For a middle velocity of 17.5 km/sec and assuming  $z = 0.1$ ,  $C_{\oplus} = 1.18$ . That is, the area of basins on the Earth is at least 18% greater than those on the Moon, both normalized by the planetary surface area. Again, the probable

existence of "extra" basins suggests that the effective  $C_{\oplus}$  is greater than this. If there are lunar basins corresponding to 40% the lunar surface area, the equivalent area of basins on the Earth is  $0.47-0.64 A_{\oplus}$ , which corresponds to 38%-47% of the Earth's surface converted into basins when allowance is made for overlap. A more reasonable value for  $a_{\text{L}}$  might be  $0.5A_{\text{L}}$ , to allow for uncounted lunar basins (such as Mare Gargantuan). Then  $C_{\oplus} \geq 1.18$  ( $z = 0.1$  at 17.5 km/sec) implies  $a_{\oplus} = 0.59-0.76 A_{\oplus}$  and  $A = 0.45-0.53A_{\oplus}$ .

In general a lower approach velocity increases the total basin area on the Earth as a function of  $a_{\text{L}}$  and  $z$ . This is because lower velocity of approach increases the ratio of impact velocities on the Earth to those on the Moon (see Figure 1), making the diameters (for any given  $z$ ) generally larger on the Earth (Figure 2a). Lower approach velocities also increase the ratio of gravitational cross-sections (Earth/Moon), meaning the Earth sustains more impacts/unit area because its greater gravity can pull in more slow moving objects. Therefore  $C_{\oplus}$  increases sharply with decreasing approach velocity (Figure 2b). Consider  $V_{\infty} = 15$  km/sec,  $z = 0.1$  again. Then for  $a_{\text{L}} = 0.4 A_{\text{L}}$ ,  $a_{\oplus} = 0.54-0.71 A_{\oplus}$  (allowing for extra basins) and  $A = 0.42-0.51 A_{\oplus}$ . For  $a_{\text{L}} = 0.5 A_{\text{L}}$ , these increase to  $a_{\oplus} = 0.68-0.85 A_{\oplus}$  and  $A = 0.49-0.57 A_{\oplus}$ .

Therefore it would seem that the observed number of lunar basins plus those "extra" basins suggested by the  $D^{-2}$  law yield a basin equivalent area of at least  $0.70 A_{\oplus}$ , which in turn means at least 50% of the Earth's surface was converted into basins by 4 billion years ago. This value of "at least 50%" seems warranted considering that the number of lunar basins observed is truly a minimum, and agrees with a previous estimate made using simpler arguments (Frey, 1977b). Note the highest value suggested by Figure 4 ( $V_{\infty} = 15$  km/sec,  $z = 0$  and  $a_{\text{L}} = 0.6$ ) has about 70% of the Earth surface covered by basins.

EFFECTS OF BASIN-FORMING IMPACTS

There are three major effects that result from the formation of a large impact basin: (a) creation of a large but shallow crater, (b) pressure-release partial melting at depth and subsequent flooding of the basin by basaltic lavas, and (c) alteration of the geothermal gradients below the basin (Frey, 1977a). All of these effects depend on the depth of the crater formed. It is not obvious that either lunar, mercurian or terrestrial depth (d)-diameter (D) relations apply to the case of very large impact basins (diameter  $\geq 1000$  km). All three planets show a distribution of depths that can be described by two straight lines. For the Moon (Pike, 1974)

$$d = 0.196D^{1.010} \text{ 10km} \quad (10a)$$

$$d = 1.044D^{0.301} \text{ 10km} \quad (10b)$$

Mercury craters follow this broken distribution to within 10% for fresh craters (Malin and Dzurzin, 1978). The transition to a lower slope probably involves modification of a the crater form during impact. Terrestrial craters change from simple to complex (for those formed in crystalline rock) at about 3-4 km diameter (Grieve and Robertson, 1978) while lunar and mercurian craters change slope at about 10 km (Malin and Dzurzin, 1978). Grieve and Robertson (1978) give for the Earth

$$d = 0.326D^{0.786} \text{ (simple, in crystalline rock)} \quad (11a)$$

$$d = 0.52D^{0.189} \text{ (complex, in crystalline rock)} \quad (11b)$$

but point out the large scatter in the d-D relation for complex craters makes the slope so uncertain that it could easily be  $\sim 0.3$ , as is the case for lunar craters.

The lunar curve for large craters, if extrapolated into the diameter range for large impact basins, suggests a 1000 km basin would be ~8 km deep, the terrestrial curve gives a value of ~2 km deep. These very shallow depths are undoubtedly due to crater modification by a variety of processes including isostatic adjustment, wall failure and slumping both during and following crater formation. Extrapolation of the simple crater curves, which might be thought of as representing a transient cavity or "instantaneous crater form," yield incredible depths of 210 km (from lunar curve) and 74 km (from terrestrial curve). These depths transcend the crust and, in the extreme, the lithosphere. Under these circumstances it is very unlikely that normal cratering forms and processes apply. The equation of state of the target material changes over the presumed depth to which a shock wave might penetrate; propagation of a shock wave will no longer be spherical and the resulting cavity may be significantly more shallow than for craters formed in layers which are thick and homogenous compared to crater dimensions.

Here we are dealing with only the first order effects of major impacts; details of crater forms are beyond the scope of this paper. Below it is assumed that initial crater depths were shallow: a 1000 km wide basin is assumed to have been ~13 km deep at the outset (Baldwin, 1963). All of the effects described below are significantly enhanced if the basin was initially much deeper. The descriptions included here should therefore be viewed as a conservative "worst case."

To describe even the mechanical effects of a major impact, a model for the early Earth lithosphere is required. This becomes even more crucial for discussion of thermal effects. For reasons discussed earlier it is assumed that the Earth originally had a global crust of intermediate composition (Lowman, 1976; Shaw, 1972). This crust is represented as andesite (Lowman, 1978) and assumed to have been 20 km thick. The crust overlies a pyrolite mantle below which is peridotite (Frey, 1977a,c). Consider a late-arriving asteroidal object which impacted this crust roughly 4 billion years ago and formed a basin 1000 km wide.

Initially the crater was  $\sim 13$  km deep, but, in the thin lithosphere of the Earth this basin would soon isostatically adjust upward. The basin floor would rise over several thousand years, and leave a relief between basin rim and floor of 3-4 km (Frey, 1977a). This value is relatively insensitive to the initial depth of the crater, but is quite dependent on the assumed thickness of the low density crust. Decreasing the crustal thickness decreases the total adjusted relief, as a larger percentage of the excavated material is then of higher density.

Formation of such a crater produced a pressure drop at the surface. Removal of the overburden means that pressure-temperature relations at depth were altered in the direction that favors melting at more shallow depths. Once solid material near its pressure-melting temperature would have partially melted when the pressure was reduced. For a pyrolite mantle the partial melt would have been basalt; the liquid would have been less dense than the surrounding solid but highly fractured rock and should have risen rapidly through the thinned and broken lithosphere (Frey, 1977b). Basin filling by basalts would

have begun almost immediately following impact. Note the above has not taken account of any extra heat deposited at depth due to the impact itself (Safronov, 1978). The appearance of basaltic material in the topographic lows of major impact basins produced a compositional dichotomy on the early Earth akin to that of the Moon. By the time the bombardment had come to an end ~4 billion years ago, at least 50% of an original global sialic (continental) crust had been converted into terrestrial maria. These maria were the original oceanic basins of the Earth, as described elsewhere (Frey, 1977b). Thus the fundamental crustal dichotomy of the Earth was established very early in Earth history, making possible early plate-tectonic processes requiring subductable crust.

Note that if the impact was initially deeper than described above, the flooding of the basin would have been even more rapid. Deeper excavation produces a greater pressure drop leading to more complete melting at shallower depths. The liquid then had even less distance to rise to erupt onto the basin floor. In the extreme limit of depths  $\geq 70$  km the impact would have penetrated into already molten regions and basaltic lava would have been immediately exposed.

It has been suggested that some highly altered remnant of this original mare-type volcanism may still exist on the Earth as the greenstone belts found in Archaean terrain (Green, 1972; Glickson, 1976). Many ultrabasic Archaean rocks seemed to have formed in the presence of unusually steep thermal melting gradients (Green, 1975), such as might be associated with large impacts (see below) and the corresponding pressure change such impacts produced. No known



rocks on the Earth data from 4 billion years ago, but greenstones with ages of 3.5 billion years old do exist. At best such rocks would be late eruptions. Other possible modes of origin are also likely (see Weiblen and Schulz, 1978; Burke et al., 1976; Glickson, 1976; Green, 1975).

The question of early geothermal gradients and their variation due to impact is an important problem. In order to examine the first order effects a simple thermal history was calculated as follows. The Earth was assumed initially molten, but differentiated into a 20 km thick andesitic (or diorite) crust (density  $2.7 \text{ gm cm}^{-3}$ ), an upper mantle of pyrolite (density 3.3 from 20 to 300 km depth) and a lower mantle of peridotite (density 3.5 from 300 to 500 km depth). Specification of these rock types determines the radioactive heat contributions from  $\text{U}^{238}$ ,  $\text{U}^{253}$ ,  $\text{Th}^{232}$  and  $\text{K}^{40}$ . The one dimensional equation for thermal conduction is

$$\frac{\partial T(Z)}{\partial t} = \frac{\bar{K}}{\rho C_p} \frac{\partial^2 T}{\partial Z^2} + \frac{1}{\rho C_p} H(Z,t) \quad (12)$$

where  $\bar{K}$  is the mean thermal conductivity (here taken to be  $0.008 \text{ cal cm}^{-1} \text{ sec}^{-1} \text{ }^\circ\text{C}^{-1}$ ),  $C_p$  is the specific heat at constant pressure ( $0.25 \text{ cal gm}^{-1} \text{ }^\circ\text{C}^{-1}$ ),  $\rho$  is the density,  $T$  the temperature,  $t$  is time and  $H(Z,t)$  is the rate of heat generation per unit volume due to radioactivity. For this simple case consider only the outer 500 km of the Earth, divided into layers 10 km thick. Assume the lower boundary has fixed temperature. The surface temperature is set equal to zero and the equation solved by finite differences in time steps of 1 million years (a requirement fixed by the stability criterial for the finite difference technique). After 500 million years (i.e. roughly 4.1 billion years ago) the

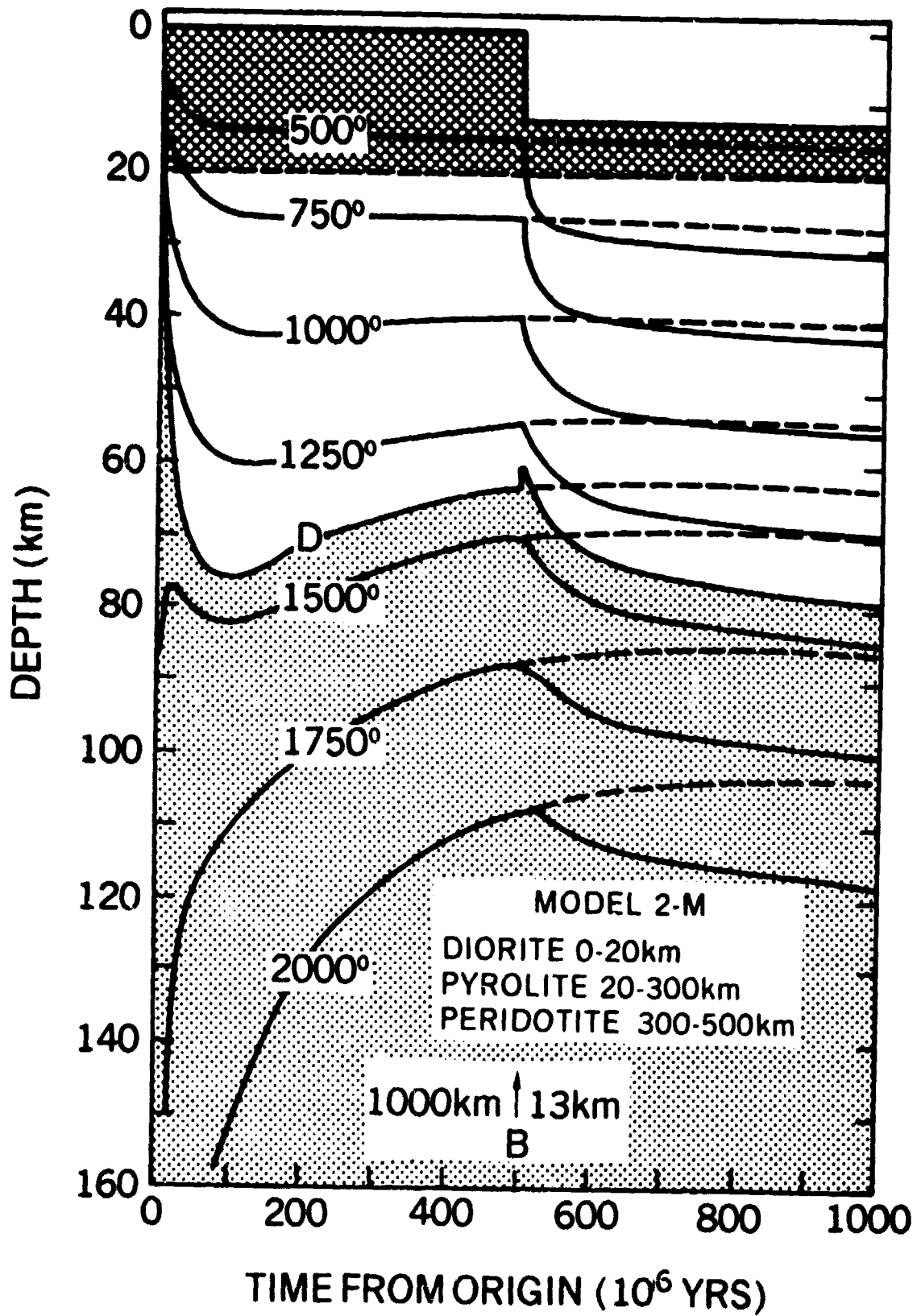
evolution is interrupted to allow for a 1000 km wide, 13 km deep impact. The newly exposed surface layer temperature is set equal to zero, the pressure and melting temperatures at depth recalculated to allow for the excavation, and the evolution allowed to continue for another 500 million years. Details are available elsewhere (Frey, 1977c).

Some of the results are shown in Figure 6. Here depth to various isotherms is plotted as a function of time since origin. The line labeled D refers to the depth at which the temperature exceeds the melting temperature at that depth; this is taken to be the lithosphere/asthenosphere boundary. Solid lines show the effect of the impact (labeled B, at 500 million years after origin) on the isotherms. Dashed lines show what would be the case if no impact had occurred; this may be taken as the thermal profile for adjacent highlands.

Initially the isotherms deepen with time as the molten model cools, but then begin to rise slightly after 150 million years as radioactive heating becomes important. The lithosphere deepens to about 75 km after 100 million years, but then thins to about 64 km just prior to the basin-forming impact. Had no impact occurred, the isotherms would remain more or less stable in this simple model, especially over the outer few hundred kilometers.

The effect of the impact is obvious. All isotherms except the melting line (D) drop rapidly. This is a consequence of penetration to deeper, hotter layers which cool rapidly by radiation, causing the loss of heat at depths to increase. The depth to melting (strictly speaking, partial melting) initially rises several km immediately after impact. This is due to the pressure-released melting associated with excavation of the basin. Melting occurs

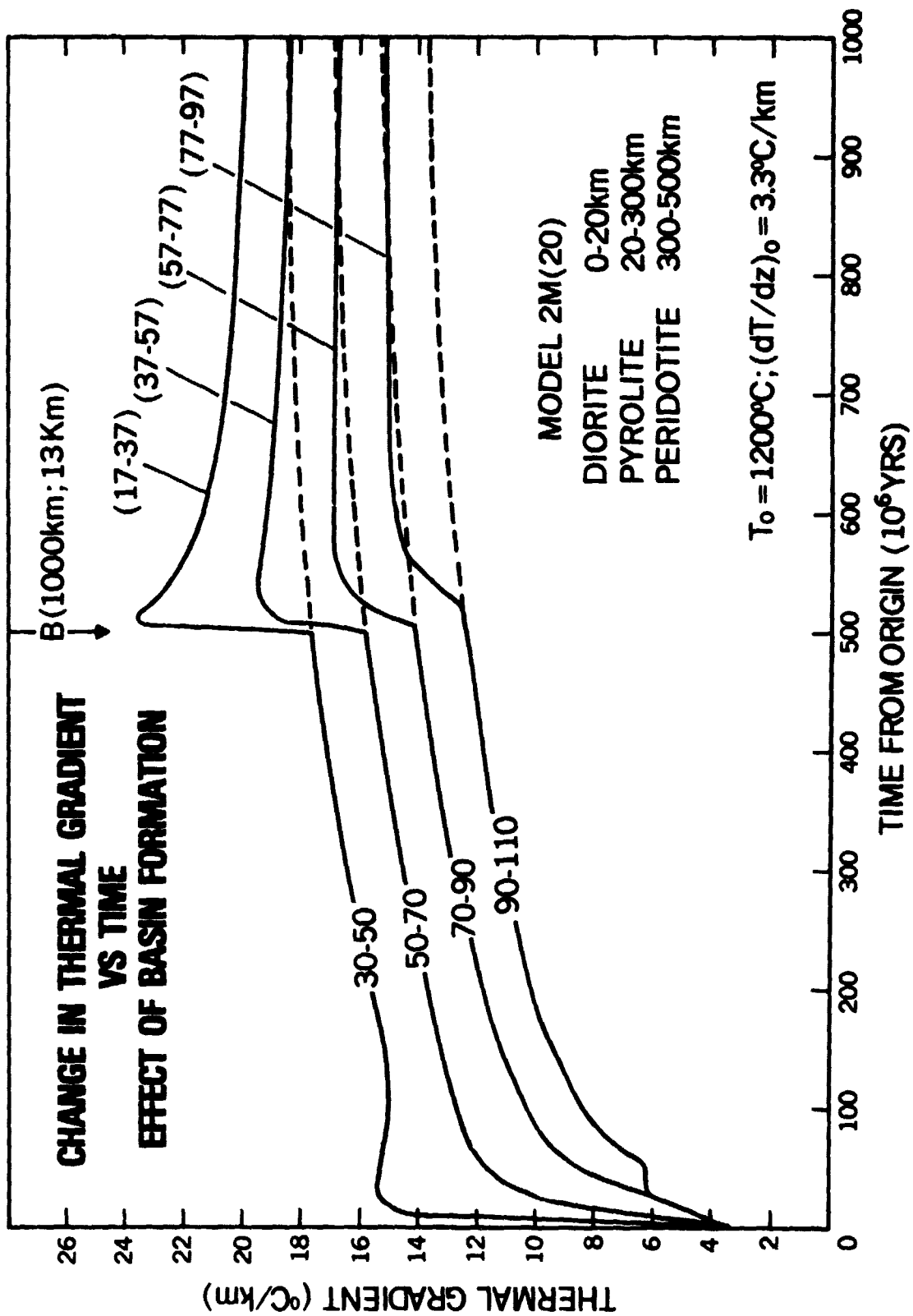
# DEPTH TO SELECTED ISOTHERMS (°C)



closer to the surface than before. The lithosphere is therefore thinner, especially due to the impact having stripped off at least 13 km of crust, as shown in Figure 6. Following the "D" isotherm also shows how the thickness of the lithosphere changes with time. All the isotherms increase with depth as the model cools; 500 million years after the impact "D" has deepened to 79 km. But the excavation of the crater removed 13 km of crust so the true thickness of the lithosphere is only 66 km, roughly equal to its pre-impact value. That is, in this simple conductivity model it takes some  $10^8$  years for the lithosphere below the basin to cool enough to thicken to the highland lithospheric thickness.

This time to cool will obviously depend on the depth of the impact; deeper penetration to still hotter layers results in more rapid cooling and thickening. For a 10 km deep impact, the time to rethicken to highland lithospheric values is 100 million years; if the impact was 50 km deep then 220 million years are required (Frey, 1977c). Qualitatively these results are in agreement with a similar case done for the Moon (Arkani-Hamed, 1974), although subbasin lithospheric thickening is some 2.5 times slower than in the lunar case. The major point here is that the subbasin lithosphere remains thin (mostly due to impact excavation) compared with the highland lithosphere for some  $10^8$  years following impact.

Formation of such a large basin would also have altered geothermal gradients. Exposure of hot deep layers to rapid cooling by radiation created very high gradients across this new surface. Gradients at depth were also affected, as shown in Figure 7 (Frey, 1977a,c). Here gradients averaged over 20 km thick layers are shown as a function of time before and after the basin-forming impact



(solid lines). The no-impact case is shown by dashed lines. At all depths thermal gradients are enhanced. In the 30-50 km layer (13-37 km below the new basin surface after impact) the change is from  $18^{\circ}\text{C}/\text{km}$  before to  $23^{\circ}\text{C}/\text{km}$  immediately after impact. At greater depths the effect is somewhat less, but it is clear that thermal gradients were enhanced by about 20% below the lithosphere by the impact. Note also that this steepening of the gradients persists over time. Had the impact been deeper than the 13 km used here the enhancement would have been even greater due to exposure of even hotter layers to rapid cooling.

The above simple thermal history has not included possible convection on the Earth, and is therefore unrealistic. McKenzie and Weiss (1975) have shown that some form of mantle convection has most likely always been part of the Earth's solution. Although not included explicitly, the qualitative effects of the basin-forming impact on early convection are obvious. Steepening the thermal gradients by 20% stirred up convective transport below the basin by 20%. This more vigorous convection operated below a thinned (by at least 25%) and badly fractured lithosphere. These were conditions favorable for rapid rifting or break up of the subbasin-lithosphere. It is not unreasonable that early plate formation and plate motion may have rapidly followed the formation of major impact basins and their flooding by basalt. That flooding provided subductable crust at the surface. The badly fractured lithosphere could easily have been broken into a large number of small microplates (Frey, 1977a,c) which might be expected to have moved more rapidly than the large plates of today. McKenzie and Weiss (1975) and Burke et al. (1976) have also argued for small plates and/or rapid motions in the Archaean.

The above discussion does not include the possibility of significant amounts of heat being deposited within the lithosphere by the impact itself (Safronov, 1978). This is a difficult problem to assess quantitatively, because for impacts of the size described here there are enormous uncertainties in energy scaling, energy partition and the depth dependence of the thermal energy deposited. However, it seems that most high temperature effects due to impacts are concentrated near the surface (Grieve et al., 1977). It may be that the near surface thermal gradients were overturned, but this should have been a short-term effect compared with others discussed here. There is not likely to have been a significant effect at the depths of the lithosphere/asthenosphere boundary unless the impact penetrated this far; in this case the incremental heat added at these levels (60-70 km depth) would not significantly add to the thermal energy of the layer (see Figure 6).

#### DISCUSSION

It is clear that the Earth, like the other terrestrial planets, must have experienced severe impact bombardment early in its history by basin-forming, asteroid-sized objects. A conservative scaling from the observed lunar basins indicates that the terrestrial bombardment was enough to convert at least half of a global sialic crust into basins some 4 km deep. For any reasonable thermal profile for the early Earth the lithosphere would have been thinner than today. Eruption of basaltic lavas onto the basin floor is expected due to pressure-release melting below the impact. A crustal dichotomy like the lunar highlands/maria division and superficially like the modern continents/ocean basins was established early in the history of the Earth, by 4 billion years ago.

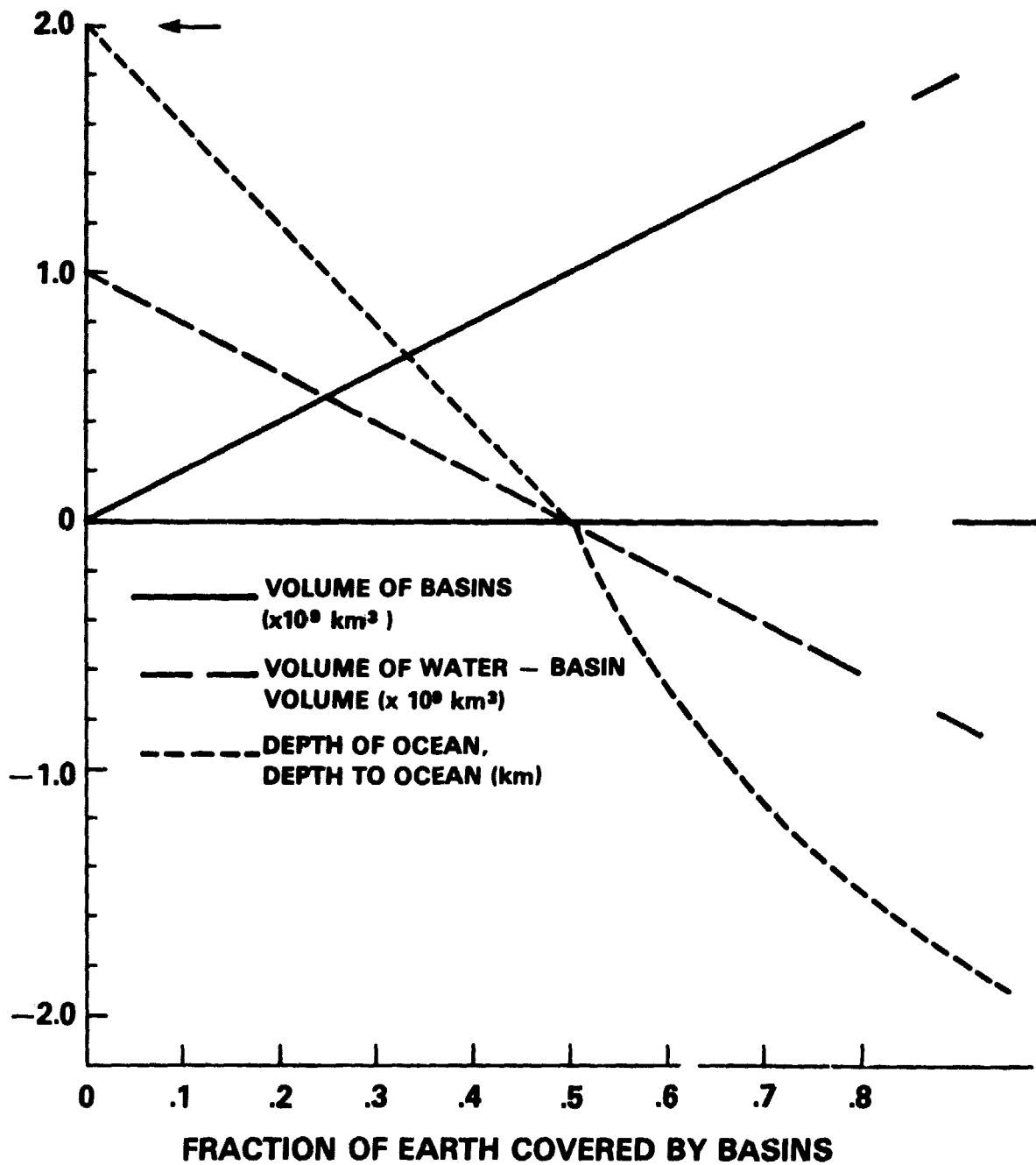
This is the general picture provided by comparative planetary evolution. The details require more study. But the above-described scenario is more realistic than many stories written about the early Earth because it is based on observational data. Too often models of the Earth are constructed based on terrestrial data alone. These are generally unsatisfactory for the earliest evolution of the Earth because of the lack of terrestrial data from the early Archaean. Many such models have completely ignored the effects of impacts on the early crust. One example is discussed below.

Hargraves (1976) has suggested a model for the primitive Earth in which a global sea of some 2 km depth overlies a primordial sialic crust. In his model continental crust thickens with time, finally emerging above sea level quite late in Earth history at 1.7 to 2.3 billion years ago, or, in his preferred version, not until the late Precambrian. Comparative planetary data would suggest that such an emergence of dry continental crust occurred much earlier, by no later than 4 billion years ago, and that it was due to the same impact bombardment described above (Frey, 1979).

If the early Earth was originally covered by a 2 km deep global sea, the intense cratering by basin-forming objects would have caused sea level to progressively lower. This is shown in Figure 8. If we assume an average adjusted basin depth of  $\sim 4$  km, then the volume of basins increases linearly (as shown by the solid line) as the percentage of the Earth's surface converted into basins increases. The volume of water on the surface of the Earth minus the cumulative volume of basins formed decreases at the same rate, as shown by the large dashes. The two curves cross at 0.25 of the Earth's surface converted to basins; at this point the cumulative volume of basins equals half the original



# CHANGE IN MEAN SEA LEVEL

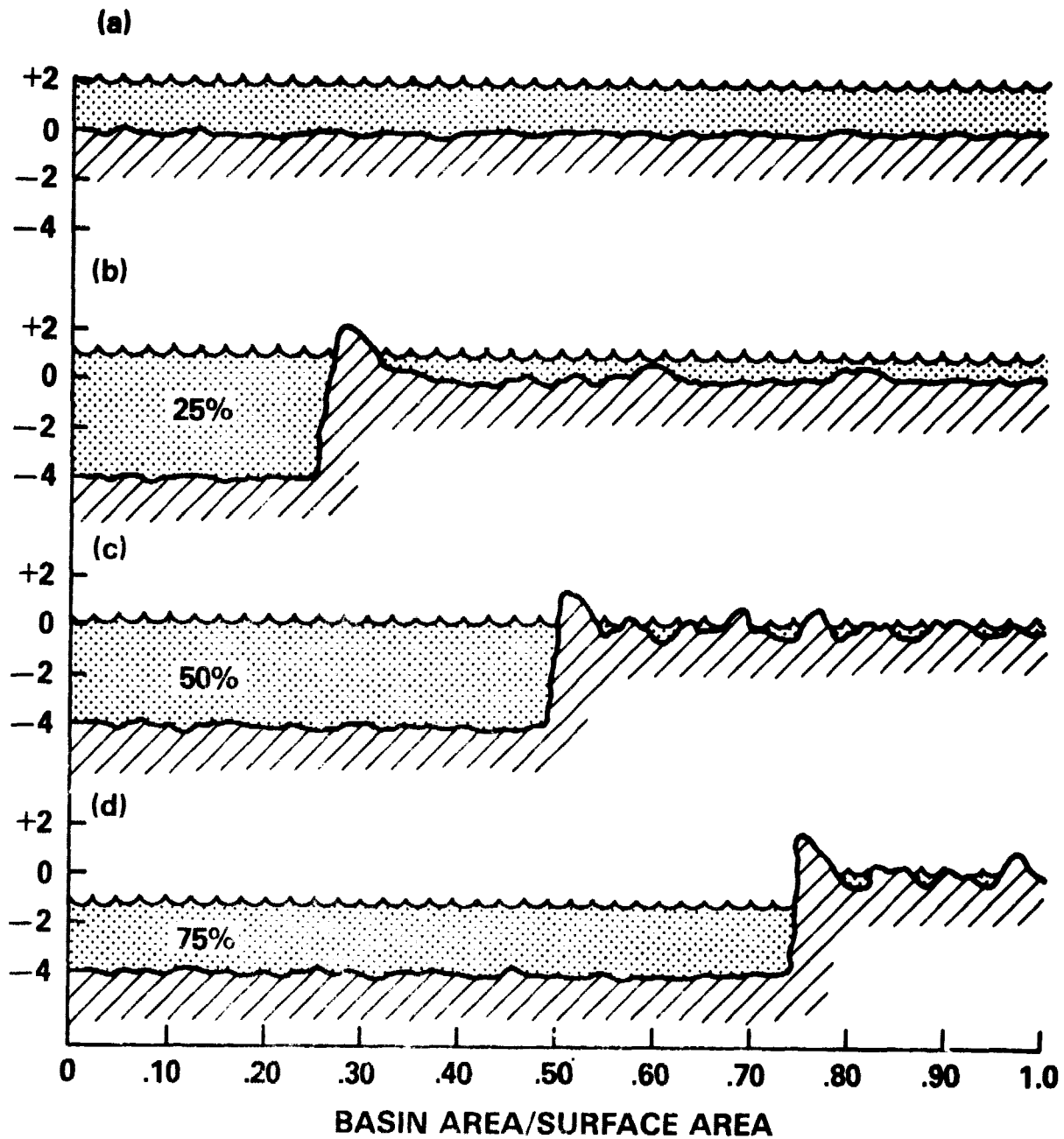


volume of water and sea level has dropped from 2 to 1 km average depth (short dashed line). Impacts which produced basins 1000 km across also produced rim relief on those basins in excess of 1 km, if lunar cratering is any guide (Pike, 1967). Therefore the highest basin rims were above sea level when basins covered only 25% of the Earth's surface. As an increasing percentage of the surface was converted into 4 km deep basins, water continued to drain into these sinks, and sea level continued to drop. Figure 9 is a schematic plot that represents this progressive emergence of dry "continental" or high-land crust. By the time the bombardment came to an end 4 billion years ago, at least 50% of the Earth's surface had been converted to basin topography; sea level had lowered to about the original crustal radius of the Earth leaving nearly half of the surface (at least slightly) above sea level.

The above says nothing about the degassing history of the Earth. It is not clear that a global ocean did exist on the primordial Earth. If it did, the topography produced by the basin-forming impact bombardment prior to 4 billion years ago reduced sea level to a point where a large percentage of the Earth's "continental" crust stood dry by the end of the bombardment. If degassing was less catastrophic, the already existing basins would have provided natural sinks into which late accumulating water would drain. It seems clear, therefore, that dry continental crust has been present on the Earth from at least 4 billion years ago, and did not remain submerged for much longer as Hargraves (1976) suggested.

The implications of an early "emergence" of continents have been discussed elsewhere (Frey, 1979). The point to be made here is the value of comparative planetary data in constraining models of the early Archaean. Scenarios which ignore the obviously important effects of early cratering are not likely to be realistic. The late heavy bombardment was a dominant process affecting the crustal evolution of the small terrestrial planets; it was also significant in the crustal development of the Earth.

# GLOBAL OCEAN INITIALLY 2km DEEP



### CONCLUSIONS

The lack of terrestrial data from 3-4 billion years ago severely restricts the accuracy of models of the early Archaean based on this information alone. The more sluggish crustal evolution of the smaller terrestrial planets has preserved a more complete record of the earliest processes affecting these bodies. When allowances for differences in size, mass, bulk composition and location in the solar system are made, the data from the smaller planets can be used to constrain models of the early Earth. This is especially true for the dominant external process which affected the crusts of all terrestrial planets, the late heavy bombardment by asteroid-sized objects. These impacts occurred throughout the inner solar system, and terminated on the Moon about 4 billion years ago. The Earth could not have escaped such impacts, and they must have significantly modified the early crust of this planet. That crust was probably thin, intermediate in composition and global in extent, again based on comparative planetary analogy.

An estimate of the effects of a late heavy bombardment on the early terrestrial crust can be made by scaling from the observed number of lunar impact basins with diameters in excess of 200 km. This number is certainly a minimum and therefore any effects should be considered lower limits for the Earth. Craters forming on the Earth will be larger than those forming on the Moon due to identical objects unless gravity severely restricts the growth of the cavity. The ratio of impact basin area on the Earth to that on the Moon increases as approach velocity decreases, because the ratio of impact velocities increases as approach velocity decreases. A much more significant effect is that of the relative gravitational cross-sections of the Earth and Moon: for reasonable

approach velocities the Earth should have collected 1.3-1.5 times as many impacts per unit area as did the Moon. By the time the lunar bombardment came to an end roughly 4 billion years ago at least 50% of the Earth's original global sialic crust would have been converted into large basins. Some 60 of these would have been more than 1000 km across. Furthermore, it is possible that the Earth sustained a few very large impacts, which would have produced basins larger than those seen on the Moon.

There were several immediate effects of these large-basin-forming impacts. After isostatic adjustment typical basin floors would have been some 4 km below the cratered highland surface. Thus the bombardment created a topographic division with half of the Earth at higher elevations. These low lying basins would have been rapidly flooded by basalts produced at depth below the basin by pressure-release partial melting. Excavation of the crater provided the reduced pressure. Penetration to deeper, hotter layers caused these levels to cool rapidly and steepened the thermal gradients below the basin. This would have stirred up whatever convection existed in the asthenosphere. Combined with the highly fractured lithosphere, it is plausible that rapid rifting and breakup of that lithosphere produced a number of small and rapidly moving "microplates."

If the degassing of the Earth was complete enough to have produced an early global ocean overlying the original crust, the topography which resulted from impact bombardment lowered sea level such that dry continental masses existed before the bombardment was complete. The large impact basins provided natural sinks into which more and more water drained as an increasing percentage of the Earth's surface as converted into these basins. Models which suggest a global ocean persisting until the late Precambrian are unlikely.

The late heavy bombardment by basin-forming asteroidal-sized objects which terminated roughly 4 billion years ago on the Moon was a dominant process affecting the crustal evolution of the smaller terrestrial planets; it was also significant in the crustal development of the Earth. That bombardment was responsible for the original ocean basins on this planet and may have triggered plate tectonic processes. These effects should not be ignored in models of the primitive Earth, and can best be understood in terms of comparative planetary studies of the evolution of the terrestrial planets.

#### ACKNOWLEDGEMENTS

This paper is based on a presentation given at a special symposium "Comparative Planetary Evolution: Implications for the Proto-Archaean" held at the 1978 meeting of the Geological Society of America in Toronto, Canada. I would like to thank Dick Grieve for his kind invitation to contribute to that symposium. My thanks also to Paul Lowman of NASA's Goddard Space Flight Center, Mike A'Hearn of the University of Maryland Astronomy Program and Bill Hartmann of the Planetary Science Institute for encouragement and helpful discussions on various aspects of this work, some of which is derived from a Ph.D. Thesis submitted to the University of Maryland as partial fulfillment of the requirements for a Doctoral degree. This work was supported by NASA Grant NGL-21-002-033.

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FIGURE CAPTIONS

Figure 1: Impact velocity and relative crater diameter as a function of approach velocity. (a) Impact velocity of the surface of the Earth ( $\oplus$ ) and Moon ( $\lrcorner$ ) and the ratio of these. Note that the ratio increases with decreasing approach velocity due to the greater effect of gravitational acceleration on slow-moving objects. (b) Basin diameter ratio versus approach velocity. Crater diameter is related to impact velocity through an energy scaling law. The ratio of diameters is not very sensitive to the exponent  $k$ , as shown by the two plotted curves.

Figure 2: (a) Ratio of areas of basins formed on Earth to those formed on the Moon by identical objects for different values of a gravitational acceleration parameter (bottom). Larger values of  $Z$  correspond to greater confinement of crater cavity, resulting in terrestrial basins being smaller than lunar basins despite greater energy released at impact on the Earth. (top) Ratio of gravitational cross-section for the Earth and Moon ( $\times 10$ ) versus approach velocity, and ratio of gravitational cross-sections normalized by physical cross-section ratio. At lower approach velocities the Earth collects many more impacts per unit area than does the Moon. (b) Conversion factor  $C_{\oplus}$  versus approach velocity for different values of the gravitational acceleration parameter. This factor combines the ratio of basin areas and the gravitational cross-section ratio normalized by physical cross-section ratio (a) into a single term which allows calculation of the equivalent surface area of basins formed on the Earth from the total area of basins formed on the Moon. This yields a minimum total area of terrestrial basins (see text).

Figure 3: Histogram of lunar basins with diameter larger than or equal to 200 km from Frey (1977b). All craters are counted without regard to specific morphologic subtypes.

Figure 4:  $\log$  (cumulative number)- $\log$  (diameter) plot for observed lunar basins (Figure 3) and calculated terrestrial basins. Solid line is  $D^{-2}$  slope fitted through  $D=300$  km. Long-dashed lines represent scaled terrestrial basins based on observed lunar basins with  $D \leq 2100$  km. Short dashed line is extrapolation to possible additional large diameter basins which might have formed on the Earth but did not on the Moon.

Figure 5: Calculated area of basins formed on the Earth as a function of total lunar basin area. Both are normalized by planetary surface area (i.e.,  $a_{\text{L}} = \text{total area of lunar basins divided by lunar surface area}$ ). Each panel corresponds to a given velocity of approach (shown at the top). Three values of  $Z$  (0.0, 0.1 and 0.2) are shown. The solid line represents  $a_{\oplus}$  calculated from  $C_{\oplus}$  (Figure 2b), that is, from extrapolation of only observed lunar basins. See text for details. Scale on right shows actual fraction of the Earth's surface covered by the total basin area given on the left scale.

Figure 6: Depth to selected isotherms versus time for a simple, purely conductive thermal history. "D" represents depth to (partial) melting, here taken to be the depth to base of the lithosphere. A 1000 km wide impact basin is shown as having occurred at 500 million years after origin.

Figure 7: Change in thermal gradient in 20 km thick layers versus time for the same thermal history model shown in Figure 6. After impact the 30-50 km deep layer is only 17-37 km below the new surface layer. All thermal gradients are enhanced by the impact which is introduced 500 million years after origin; those nearest the impact surface are steepened the most.

Figure 8: Change in sea level with increasing fraction of the Earth's surface covered by basin. After adjustment basins are assumed to be 4 km deep. Total volume of basins increases with time, causing sea level to decrease as water drains into the depressions. Global sea is assumed to be 2 km deep at the outset.

Figure 9: Schematic diagram of progressive changes in sea level with increasing basin area. When 25% of the Earth's surface has been converted into basins 4 km deep, sea level has dropped to 1 km and the highest basin rims have already emerged from the sea. When 50% of the Earth is covered by basins, roughly half of the Earth's crust stands above sea level.