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EARLY IMPACT BASINS AND THE ONSET OF PLATE TECTONICS

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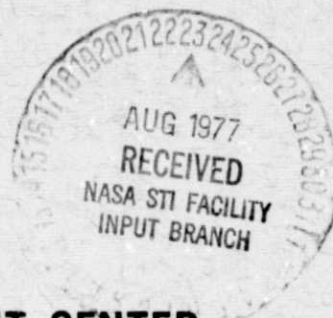
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EARLY IMPACT BASINS AND THE ONSET
OF PLATE TECTONICS

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ABSTRACT

The 4 billion year old late heavy bombardment responsible for the formation of lunar basins converted more than 50% of the Earth's original global low density crust into basin topography. These basins were rapidly flooded by basaltic lavas generated by partial melting resulting from the pressure drop due to impact excavation. In this way the fundamental crustal dichotomy of the Earth (high and low density crust) was established nearly 4 billion years ago. Therefore, subductable crust was concentrated at the surface of the Earth very early in its history, making possible an early onset for plate tectonics. Simple thermal history calculations spanning 1 billion years show that the basin forming impact not only thins the lithosphere by some 25% (or more), but also increases the sub-lithosphere thermal gradients by roughly 20%. The corresponding increase in convective heat transport, combined with the highly fractured nature of the thinned basin lithosphere, suggest that lithospheric breakup or rifting occurred shortly after the formation of the basins. Conditions appropriate for early rifting persisted from some 10^8 years following impact. We suggest a very early stage of high temperature, fast spreading "microplate" tectonics, originating before 3.5 billion years ago, and gradually stabilizing over the Archaean into more modern "large plate" or Wilson Cycle tectonics.

INTRODUCTION

Two of the outstanding problems in the crustal evolution of the Earth are the origin of continents and ocean basins (or the formation of the Earth's basic crustal dichotomy of low and high density crust) and the onset of plate tectonics. There is a relation between these two at least insofar as high density (oceanic) crust is a requirement for subduction, as shown by McKenzie (1973). Attempts in the past to construct the continents over geologic time from an initially basaltic crust have been generally unsuccessful, and recent investigations have suggested a low density crust for the early Earth, most likely of global extent (Shaw, 1972; Lowman, 1976; Hargraves, 1976). The establishment of the original crustal dichotomy then requires destruction or conversion of low density crust into higher density material. This conversion must predate the onset of plate tectonics.

The onset of, and original form of plate tectonics, on the Earth is unknown. Burke et al (1976) and Burke and Dewey (1973) argue that modern Wilson Cycle plate tectonics have existed for perhaps 2.5 billion years, and point to the existence of Proterozoic sutures as evidence of ocean closings in the early history of the Earth. Prior to this period of "large plate" tectonics they picture a "permobile regime" of fast spreading and large total length of spreading margins to dissipate the greater heat production of the early Earth. McKenzie and Weiss (1975) argue that the early convection of the Earth would prohibit the formation of large plates early in the history of the Earth. It has been impossible to date to specify when and how early plate tectonics began,

and what relation these had to later more modern Wilson Cycle tectonics. It is this problem which is addressed in this paper. We show that the early establishment of the fundamental crustal dichotomy led to an early onset of a "microplate" tectonic phase, which over time gradually stabilized into modern "large plate" type plate tectonics.

CRUSTAL DICHOTOMIES OF TERRESTRIAL PLANETS

The basic crustal characteristic of the Earth is its 60% oceanic (high density) and 40% low density (continental) crust. This characteristic is not singular with the Earth; all of the terrestrial planets for which sufficient observational data exists share this attribute. For the smaller terrestrial planets the division is between highlands and maria, but the characteristic (low/high density crust) is the same. The interesting fact is that the percentage of high density crust - and in general, newer crust - increases with increasing size and mass of the planet. This is shown schematically in Figure 1 for the Earth, Mars and the Moon (see Frey, 1977a). The crustal division into high/low density crust is for the Moon: 30/70 (maria/highlands), for Mars: 45/55 (maria/highlands) and for the Earth: 60/40 (oceanic/continental crust). Mercury also shows a lunar-type highlands/maria dichotomy, and Venera gamma-ray observations from three sites on Venus indicate two basaltic and one approximately granitic compositions (paper presented at 8th Lunar Science Conference).

This common characteristic is superficial in that modern oceanic crust on the Earth is created by seafloor spreading, whereas the high density maria of the Moon were produced by a very different process.

% "NEW CRUST" VS MASS, DIAMETER

	MASS (Gm)	LOG (MASS)	DIAMETER (Km)	LOG (DIAMETER)
EARTH	5.98×10^{27}	27.777	12,756	4.106
VENUS	4.86×10^{27}	27.687	12,118	4.083
MARS	6.60×10^{26}	26.820	6,738	3.829
MERCURY	3.30×10^{26}	26.519	4,864	3.687
MOON	7.35×10^{25}	25.866	3,476	3.541

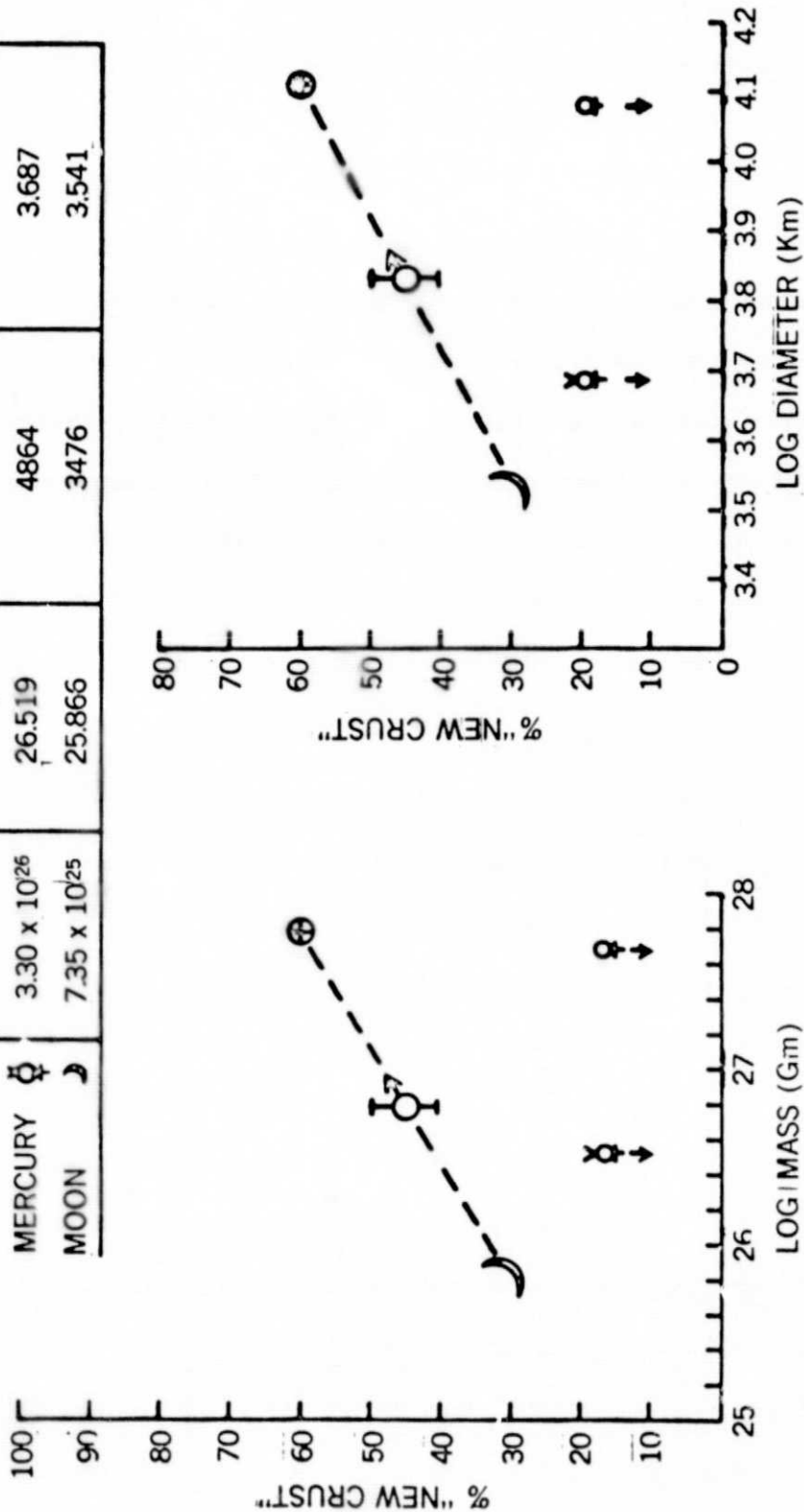


FIGURE 1

Furthermore, the high density crust of the Earth is everywhere less than about 200 million years old, whereas the lunar basaltic rocks are more than 2.5 billion years old (El-Baz, 1975). But modern oceanic crust on the Earth is only the most recently generated high density crust; the net effect of plate tectonic processes is to rework the oceanic parts of the crust by generation of new basalt at mid-ocean ridges and destruction of older seafloor at subduction zones. Only high density crust can be subducted (McKenzie, 1973); therefore high density crust has been in existence much longer than the modern ocean basins. If plate tectonics have operated on the Earth for some 2.5 billion years, then high density crust existed on the Earth at least this long ago. The fundamental question then is: When did high density crust first form on the Earth?

No trace of the original high density crust of the Earth remains today, but as shown previously by Frey (1977b), the possible origin of such crust can be examined by interplanetary comparisons. The lunar crustal dichotomy can be traced back about 4 billion years to a period of late heavy bombardment by large planetesimals. These impacts produced a large number of giant basins (Wood and Head, 1976; Hartmann and Wood, 1971; Howard et al., 1974), which were later flooded by Fe-rich basalts derived from great depths. Similar basin-forming impacts have occurred on Mars (Wilhelms, 1973) and Mercury (Murray et al., 1975). It is unclear whether these objects represent the last stages of accretional bombardment or a particular flux of objects which affected the entire inner solar system some 4 billion years ago

(Wetherill, 1975, 1976; Chapman, 1976). In either case the Earth must also have experienced such a bombardment.

A lower limit to the number of basin-forming impacts that occurred on the Earth 4 billion years ago may be obtained by scaling from the observed number of lunar basins (Frey, 1977b). Two factors enter into this scaling, the impact velocity and the capture cross-section of the Earth compared with that of the Moon. The impact velocity of objects entering the Earth-Moon system will be higher at the surface of the Earth. For Apollo-type asteroids (a modern example of the kinds of objects that could have produced the lunar basins) this velocity of approach will be 15-20 km/sec (Opik, 1966); the ratio of impact velocities on the Earth to those on the Moon will then be 1.23 to 1.14. Impact velocity may be converted into basin diameter through an energy scaling relation. For example, from Hartmann (1965),

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

Using the ratio of basin diameters on the Earth to those formed on the Moon by objects of equal mass eliminates the poorly known constant C. This relationship is not very sensitive to k, whose values probably range from 1/3 to 1/3.4 (see Frey, 1977a; also Hartmann, 1965). Basin diameters are some 11-15% larger on the Earth than on the Moon.

The much greater effect is that of the capture cross-section of the Earth compared with that of the Moon. The Earth's gravitational radius is 4.2-4.5 times that of the Moon, depending on the velocity of approach. The gravitational cross-section of the Earth is some

17-20 times larger than that of the Moon, while the physical cross-section for the Earth is only 13.5 times that of the Moon. This means the Earth collects some 1.3-1.5 times as many impacts per unit area as does the Moon.

Using the observed number of lunar basins (e.g., Wood and Head, 1976), it is then possible to calculate the total area of basins formed on the Earth. Depending on the exact number of basins counted on the Moon and the velocity considered, this figure is equivalent to some 48-78% of the Earth's surface area. A simple calculation (see Frey, 1977a) shows that these basins would be distributed over something like 50% of the Earth's surface.

Therefore 4 billion years ago at least 50% of the Earth's surface area was converted into basin topography. The actual area affected is likely to be much larger, since the number of observed lunar basins represents a minimum number for those that actually formed. Another effect is the possibility that basins much larger than those observed on the Moon formed on the Earth. D^{-2} scaling suggests that if there were 3 basins larger than 1000 km on the Moon (which implies that there were then some 56 such basins on the Earth), there could have been 6 basins larger than 3000 km and at least one of these would have been larger than 6000 km. As shown elsewhere, a large fraction of the 50% basin topography on the Earth was in the form of very large basins greater than 1000 km diameter (Frey, 1977b). If formed in a global low density crust (see, e.g., Lowman, 1976), then by 4 billion years ago one-half of this crust was in the form of low-lying basins.

The pressure drop caused by the impact excavation triggers partial melting at depths more shallow than before the impact. If the early mantle of the Earth was pyrolite (see below), then partial melting will produce a basaltic liquid whose density is less than the density of the overlying solid but highly fractured rock. The liquid basalt will rise rapidly in this situation, flooding the basin in a time-scale short compared to both the flooding of the lunar basins and the isostatic adjustment time of the terrestrial basins (Frey, 1977a). Therefore, 4 billion years ago some 50% of the Earth's original global low density crust was converted into high density maria. The original ocean basins were mare-type basins. The original crustal dichotomy was established early in the history of the Earth, thereby concentrating subductable crust at the surface, and making possible the subduction portion of plate tectonics.

THERMAL HISTORY CALCULATIONS

If the crustal dichotomy was established early in the history of the Earth, how rapidly did the onset of plate tectonics follow? One way to examine this question is to consider the thermal history of the upper Earth in terms of the effects of a basin-forming impact. The impact will significantly thin the lithosphere, which will favor the formation of plates. The concentration of high density crust at the surface of the Earth makes possible subduction, which is required to compensate the formation of new crust at spreading ridges. Plate motion probably requires some form of convection, although the exact details are not clear (see McKenzie and Weiss, 1976). It is therefore

of interest to examine the thermal conditions of the Earth at the time of basin formation to see whether or not plate formation and motion are favored.

Thermal gradients were higher in the past. This should be obvious because the rate of heat generation per gram was higher in the past due to radioactive Uranium, Thorium and Potassium. Figure 2 shows the heat generation as a function of time in units of 10^{-6} cal gm⁻¹yr⁻¹ for five representative rock types. Two crustal rock types are shown: granite and diorite. Three possible mantle rock types are displayed: basalt, pyrolite and peridotite, where pyrolite is a synthetic mixture of basalt: peridotite = 1:3. That is, pyrolite represents a primitive parent mixture before the basalt fraction has been removed. Also shown in this Figure are the individual contributions from U²³⁸, U²³⁵, Th²³² and K⁴⁰ for granite. The upturn in the curves prior to 3 billion years ago is due to the greater contribution of U²³⁵ before that time. The significance of the greater heat production and inferred greater heat flow in the early Earth is that the thermal gradients would have been steeper. This in turn means the lithosphere would have been thinner than the modern (continental) value. As shown below, prior to about 4 billion years ago the lithosphere below the continental crust was probably less than about 65 km thick.

In its general form, the equation which governs the thermal evolution of a planet is:

$$\frac{\partial T}{\partial t} = \frac{1}{\rho C_p} \nabla \cdot (K \nabla T) + \dot{V} \cdot \nabla T + \frac{1}{\rho C_p} H(t) \quad (2)$$

RADIOACTIVE HEAT GENERATION THROUGH TIME SELECTED ROCK TYPES

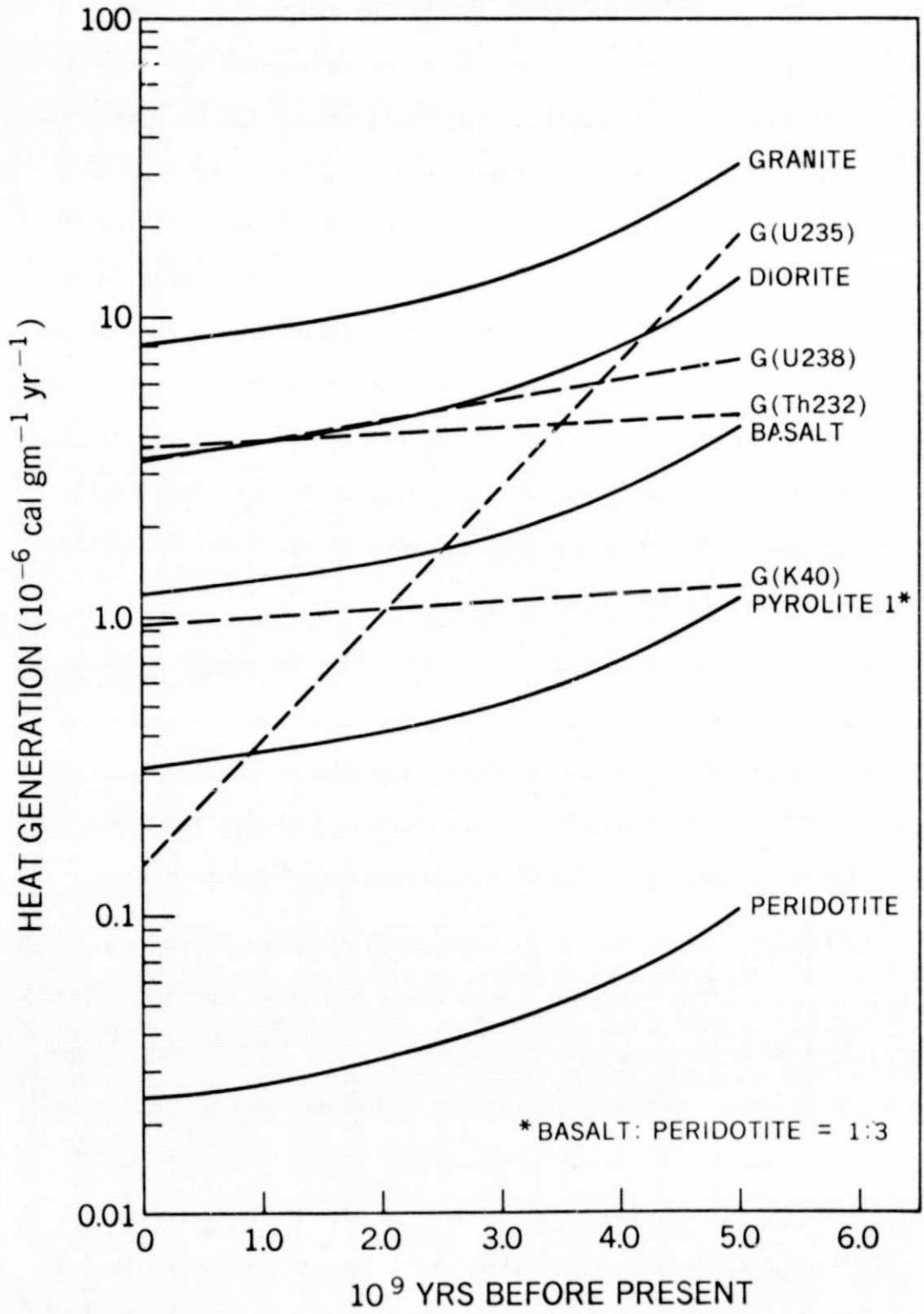


Figure 2.

where K is the thermal conductivity, ρ is the density, C_p is the specific heat at constant pressure, \vec{v} is the convective velocity and $H(t)$ is the rate of heat generation per unit volume due to radioactive decay. No analytic solution to this equation exists. We treat a simpler version of the above by assuming no convective transport of heat. As pointed out by McKenzie and Weiss (1975), such models are likely to be unrealistic, as convection has always been a part of the Earth's thermal evolution. But we are considering here only the outer few hundred kilometers of the Earth; the results below do not depend on the details of convection. We will discuss the expected effects convection would have on these models, but will not treat convective transport in the models themselves.

We also assume a constant average thermal conductivity. Shatz and Simmons (1972) have shown that for many earth materials the conductivity is dependent on temperature. Because we treat only relatively shallow depths (compared to the radius of the Earth), no great errors are introduced by assuming a constant value for this term.

The equation is solved in one dimension, and the Earth is assumed to have already differentiated into a crust and upper mantle. Concentrations of the radioactive species U^{238} , U^{235} , Th^{232} and K^{40} are specified by rock types assigned to the crust and mantle. Evolution of the outer 500 km is treated only; deeper effects are considered negligible on the 1 billion year timescale examined here. The base of this region is assumed to have constant temperature determined by the initial temperature profile. That is, the outer five hundred km

is assumed to be in contact with a thermal reservoir, which is a reasonable assumption considering the high thermal inertia of the Earth. The simplified equation is then

$$\frac{\partial T(z)}{\partial t} = \frac{\bar{K}}{\partial C_p} \frac{\partial^2 T}{\partial z^2} + \frac{1}{\partial C_p} H(z, t) \quad (3)$$

which is solved every 1 million years for depth steps of 10 km. This 1 million year time step is required by stability conditions of the finite difference scheme used here. The calculations were run over a 1 billion year time period which was interrupted 500 million years after origin by a basin-forming impact.

We must specify the composition and density of the crust and mantle, and the thickness of each rock type, as well as the original temperature profile. The other parameter which we vary is the depth of impact excavation, with the corresponding change in subsurface pressures and therefore melting temperatures. In terms of the thermal history, the distribution of the radioactive species is of paramount importance. This is determined by the choice of the rock types used to represent the crust and mantle, and the assumed thickness of the crust.

No rocks survive from the origin of the Earth, but recently thinking has turned away from continental growth over geologic time toward concepts of a global sialic crust forming early in the history of the Earth (Hargraves, 1976; Shaw, 1972). If we assume the early mantle was pyrolitic in composition, then the composition of the original crust is a very important parameter in that most of the radioactive

elements are concentrated here. If global in extent, the original crust could not have been granitic, as such a composition (over a pyrolite mantle) leads to unacceptably high values of the modern heat flow for reasonable thicknesses of the original crust. For example, a 20 km thick (see Frey and Lowman, 1976) granite crust over a pyrolite mantle, if redistributed by basin formation over 50% of the Earth, would produce a "decayed" heat flow of some 2.4×10^{-6} cal $\text{cm}^{-2}\text{sec}^{-1}$, compared with the observed modern value of about 1.5×10^{-6} . A more reasonable model is that proposed by Lowman (1976) of an intermediate (andesite or diorite) crust of global extent. More recent granitic rocks can then be derived by partial melting, or "redifferentiation" (Lowman's term) of this andesitic material. More to the point, the decayed heat flow from a dioritic crust 20 km thick over a pyrolite mantle is about 1.6×10^{-6} cal $\text{cm}^{-2}\text{sec}^{-1}$, which is quite close to the modern value. To meet this boundary condition of the modern heat flow decayed from the assumed original crust, we adopt the following model:

0 - 20 km:	diorite	density = 2.7 gm/cm^3
20 - 300 km:	pyrolite	density = 3.3 gm/cm^3
300 - 500 km:	peridotite	density = 3.5 gm/cm^3

The crustal thickness is of course an important constraint in these models, as it determines the total volume of the radioactive-rich material. It is unlikely that the original thickness was much greater than the value quoted above. If global in extent, the redistributed (by basin-forming impacts) crustal thickness would be exceptionally thick compared to modern day values if the original

thickness was greater than about 25 km. As shown below, cooling of the Earth early in its history to depths greater than 40 km was rapid, which suggests that the original differentiation would have produced a crust greater than 10 km. We consider the 20 km thickness a reasonable value for these models.

The original temperature distribution is also important in that the base temperature is fixed by the original profile. We consider two cases below: a melted model (M) where the Earth begins totally melted, and a colder model (C) where the crust and upper mantle start solid. In the first case the initial surface temperature is 1200°C , and the thermal gradient is $3.3^{\circ}\text{C}/\text{km}$; in the second case, the initial surface temperature is 0°C and the thermal gradient changes from $25^{\circ}\text{C}/\text{km}$ to $15^{\circ}\text{C}/\text{km}$ at 10 km, then to $5^{\circ}\text{C}/\text{km}$ at 100 km and finally to $0.7^{\circ}\text{C}/\text{km}$ at 200 km. In this cold model, melting occurs initially at depths of some 100 km. This is then the depth to the lithosphere/asthenosphere boundary at the outset. These two models are designated 2M(20) and 2C(20) respectively, and, as will be shown below, both converge to essentially the same condition by the time the basin-forming impact is introduced.

In addition to those parameters already specified, the following values were used:

$$\begin{aligned} C_p &= 0.25 \text{ cal gm}^{-1} \text{ }^{\circ}\text{C}^{-1} \\ \bar{K} &= 0.008 \text{ cal cm}^{-1} \text{ sec}^{-1} \text{ }^{\circ}\text{C}^{-1} \end{aligned}$$

In computing the melting temperature with depth, a pressure-melting gradient of $10^{\circ}\text{C}/\text{kbar}$ was assumed.

The evolution was begun by setting the temperature at the surface equal to 0°C (simulating cooling by radiation), then iterating the equation of thermal conductivity (Equation 2). After 500 million years, the basin-forming impact was introduced. The excavated layers were all set equal to 0°C , as was the new surface layer. Pressures and melting temperatures with depth were recomputed and the iteration continued. The results are shown below.

RESULTS OF THERMAL CALCULATIONS

Figure 3a shows the temperature versus time curves for various depths for the 2M(20) model. This model begins molten, but the outer layers cool quickly, forming a lithosphere some 70 km deep within the first 50 million years. Shortly before the impact, the outer 150 km or so have reached a near steady state condition; we therefore expect these models to be relatively insensitive to the choice of initial temperature profile by the time of basin formation. Figure 3b shows the corresponding temperature versus time plot for the 2C(20) model, and comparison of these two figures shows that both converge to essentially the same values after 150-200 million years. Even at depths of 150 km the differences are less than 100°C . Further discussion will therefore treat only the initially molten model, designated 2M(20), as shown in Figure 3a.

The effect of the impact is obvious. The 1000 km wide basin is assumed to have a depth of excavation of 12.7 km (using Baldwin's 1963 values; see below). Thus the impact penetrates to regions where the

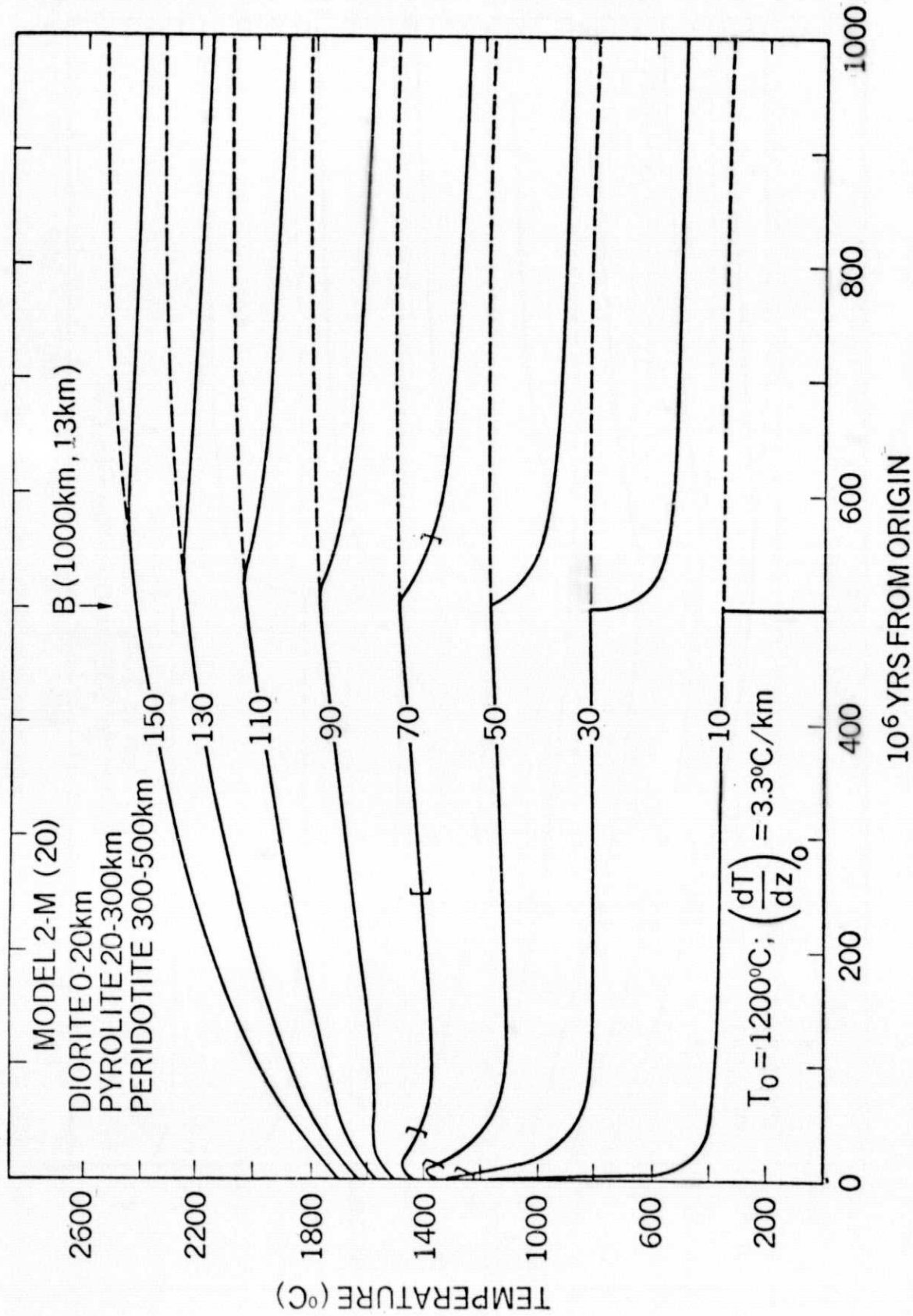


FIGURE 3a.

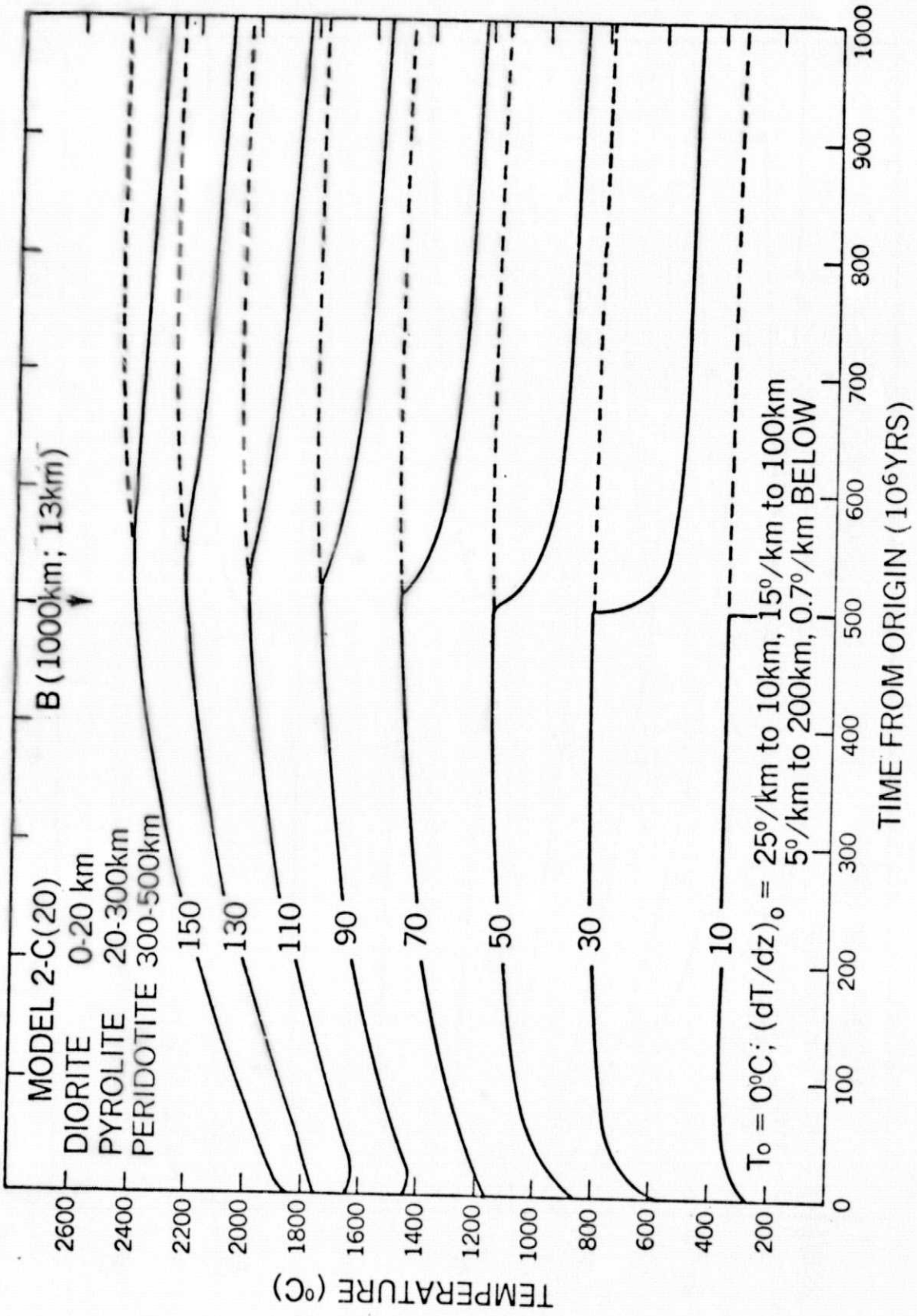


FIGURE 3b.

temperature is some 600°C . These exposed layers cool rapidly by radiation, leading to a rapid cooling by conduction of the outer layers, as shown by the steep drop of the curves. This rapid conductive cooling is a consequence of the very high instantaneous thermal gradient across the newly exposed layers. Cooling of deeper layers is progressively less and is delayed, as would be expected. At a depth of 150 km, the post-basin cooling is only some 150°C over the next 500 million years.

Another representation of these effects is shown in Figure 4, where the depth to various temperature values is shown as a function of time for the initially molten model. D refers to the depth to melting, here taken to be that point where the temperature exceeds the melting temperature for that depth. Strictly speaking this region is only partially melted, but we shall consider this the effective depth of the lithosphere for these models. As with the previous figures, the solid lines show the effect of basin-formation while the dashed line shows the no-impact case. The no-impact case may be considered to be the situation under the adjacent highlands.

As shown previously, the initially molten model cools rapidly in the outer layers, achieving a near steady state by 500 million years after origin. The low temperature isotherms stabilize after some 150 million years, but higher temperature isotherms, after an initial deeping, rise slowly as the long-term effects of radioactive decay begin to heat up the model. The lithosphere thickness deepens to about 75 km after 100 million years, but then thins due to internal heating to about 64 km just before the impact occurs.

DEPTH TO SELECTED ISOTHERMS (°C)

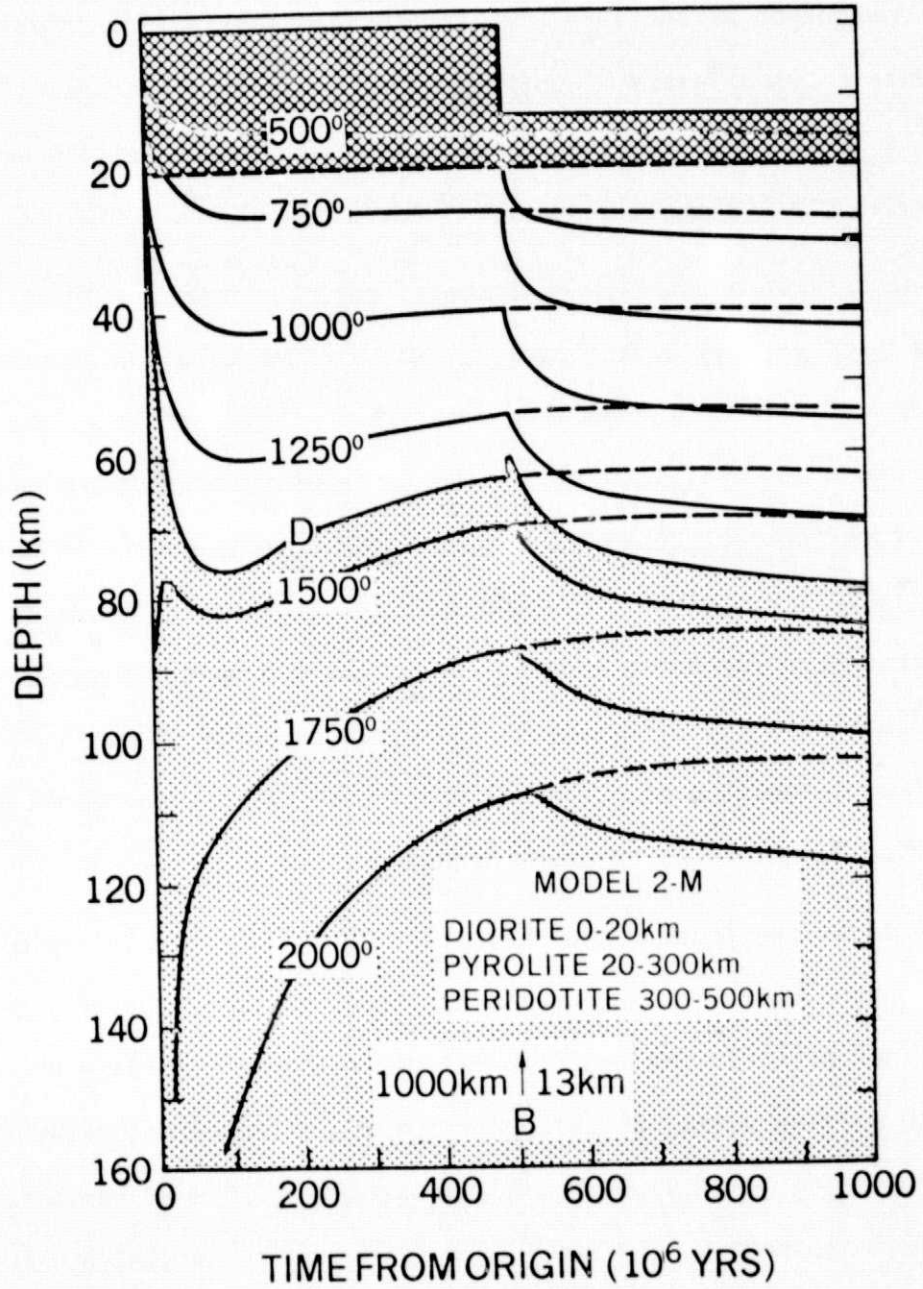


FIGURE 4.

Immediately after impact the depth to melting decreases by some 3 km. This is a consequence of the pressure drop due to the excavation of nearly 13 km of material by the impact. The reduced pressure means that material at depths of around 60 km is hot enough to begin melting. Partial melting produces basalt, which should flood the basin rapidly, as discussed before (Frey, 1977b). The net effect is to thin the lithosphere by some 3 + 13 km or about 16 km. The greater part of this is in the impact excavation, and therefore the assumed depth of the impact is important. For reasons discussed below, the value adopted here is probably conservative. The lithosphere is thinned by some 25%, with a thickness immediately after impact of 48 km of highly fractured rock.

Figure 4 also shows how the lithosphere thickens with time. Exposure of the deeper layers produces rapid cooling, as shown by the 500 and 750°C isotherms. At deeper layers the cooling is slower, but still significant in that the lithosphere base has dropped to about 79 km after 500 million years. However, the lithosphere thickness is then 79-13=66 km, or roughly its value before the impact. In more detail, the lithosphere remains thinner than its no-impact value for some 150 million years after the impact. Even though it cools faster than the highland lithosphere, the great thinning due to the impact is dominant for 10^8 years after formation of the basin.

The time to rethicken will obviously depend on the depth of the basin formed, as this is the major source of lithosphere thinning. The situation described above is conservative; Baldwin's (1963) depth-

diameter relations yield a minimum depth for large basins. Pike's (1967) curves would suggest significantly deeper initial craters for the same diameter. It is not obvious that either of these relations apply to the case of the very large craters discussed here, but the depth effect was examined by running a series of 2M(20) models with arbitrary depths ranging from 10 to 70 km. Arkani-Hamed (1974) has performed calculations similar to those above for a large lunar basin. He finds a 40 km deep impact (including fallback thickness) in a lithosphere originally 70 km deep resulted in thickening of the sub-basin lithosphere to 200 km over the first 500 million years after impact. During this same time period, the highland lithosphere increased in thickness by an additional 30 km to 100 km; the average rates of thickening were then 2.6×10^{-7} and 6×10^{-8} km/yr below the lunar basin and lunar highlands respectively. Calculations done here for the Earth show that the rate of thickening depends on crater depth, and that thickening is rapid at first with the rate decreasing over time (see Figure 4). Table I summarizes the results for basins with depths originally 10, 30, 50 and 70 km. Terrestrial sub-basin thickening is some 2.5 times slower than the comparable lunar case. This is a consequence of the greater heat generation within the Earth. The terrestrial highland lithosphere remains approximately 63 km thick over the 500 million years after the impact, as shown in Figure 4.

Of greater relevance here is the time over which the sub-basin lithosphere remains thin compared to the adjacent highland lithosphere. The 40 km deep lunar basin has a lithosphere which surpasses the highland

TABLE I - - LITHOSPHERE THICKENING vs. BASIN DEPTH FOR THE EARTH

BASIN DEPTH (km)	10	30	50	70
lithosphere thickness after impact (km)	50.1	26.7	15.2	10.9
thickening - first 50 million years (km)	8.9	24.5	29.5	29.2
rate of thickening (km/yr)	1.8×10^{-7}	4.9×10^{-7}	5.9×10^{-7}	5.8×10^{-7}
thickening - first 500 million years (km)	19.0	45.7	56.8	61.4
rate of thickening (km/yr)	3.5×10^{-8}	9.1×10^{-8}	1.1×10^{-7}	1.2×10^{-7}
time to rethicken to highland depth (10^6 yrs)	100	157	220	257
total thickening (km)	12.4	35.5	47.0	51.3
rate of thickening (km/yr)	1.2×10^{-7}	2.3×10^{-7}	2.1×10^{-7}	2.0×10^{-7}

lithosphere thickness with 100 million years (Arkani-Hamed, 1974). On the Earth the comparable timescale for sub-basin thickening is 180 million years. As shown in Table I, the deeper the impact, the longer the time to achieve a thickness comparable to that of the highland lithosphere. Even though thickening is faster with deeper craters, the thinning of the lithosphere is so great that the total time to recover the "original" thickness increases with increasing depth of excavation. In all cases discussed above, the sub-basin lithosphere remains thinner than the highland lithosphere for at least 100 million years.

The case discussed here represents a "worst case." Deeper impacts only would lead to more rapid establishment of the crustal dichotomy of the Earth (Frey, 1977b), but would result in a longer enduring of the thinner lithosphere below the basin. As discussed below, these effects are all in the right direction to make lithosphere breakup more likely with deeper basins.

The basin-forming impact will alter the thermal gradients in the lithosphere and upper asthenosphere below the basin: the abrupt cooling at depths $\gtrsim 100$ km is a response to the change in thermal gradients near the newly exposed surface. Figure 5 shows the change in thermal gradients over time in 20 km thick layers in the lower lithosphere and upper asthenosphere. Dotted lines indicate the no-impact case. The effect of the impact is obvious. Thermal gradients at all depths are enhanced; those closest to the (new) surface are affected the most. In the 30-50 km deep layer the gradient is enhanced from about $18^{\circ}\text{C}/\text{km}$

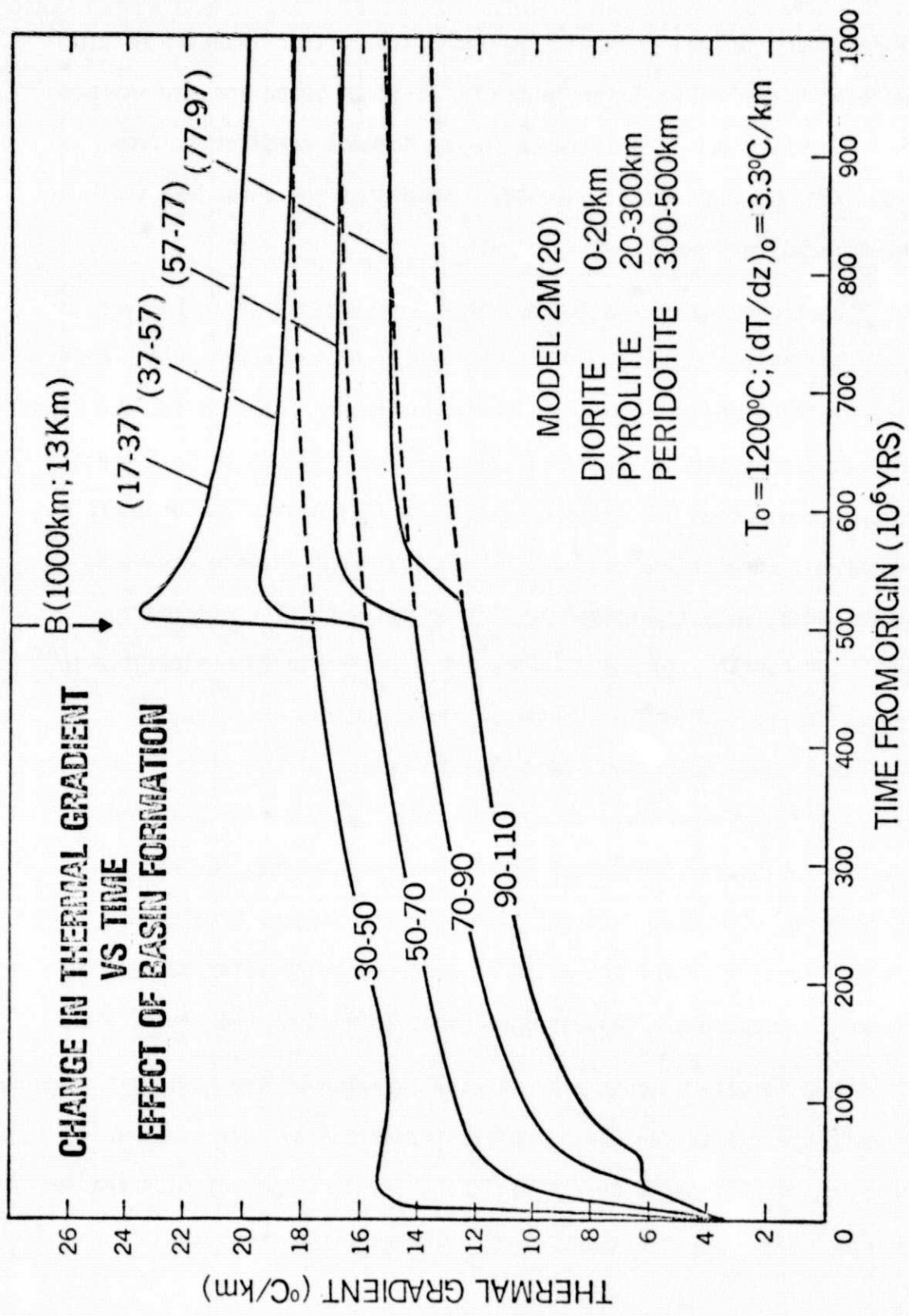


FIGURE 5

before impact to more than $23^{\circ}\text{C}/\text{km}$ immediately after impact. Melting occurs in the 50-70 km layer (which is 37-57 km below the new surface after impact). Here the increase in the thermal gradient is from $16\text{-}19^{\circ}\text{C}/\text{km}$, an increase of some 20%. Note that the enhancement of the thermal gradients occurs at all levels.

The steepening of the thermal gradient suggests that whatever convection exists below the lithosphere will be enhanced by this same factor. Immediately below the lithosphere the enhancement is some 20%; convective transport of heat will also increase by 20% in this region. Although convection has not been explicitly included in these models, the thermal effects are obvious: the impact triggers more vigorous convection below a significantly thinned lithosphere of fractured rock. The basin lithosphere is therefore particularly susceptible to breakup due to convective overturn. From Figure 5 and Figure 4 it is clear that these conditions favorable to rapid rifting of the sub-basin lithosphere persist for several times 10^8 years despite the fact that the sub-basin lithosphere cools more rapidly than the highland lithosphere. The most favorable time for lithosphere breakup due to internal convective motions would have been shortly after basin-formation; the most susceptible lithosphere was that thinned by the impact event.

These results suggest a rapid early breakup of the basin-lithosphere or oceanic parts of the Earth. Plate formation may have been rapid following establishment of the basic crustal dichotomy of high and low density crust. The consequences of this are discussed below.

DISCUSSION

Burke et al. (1976) have suggested that one way to dissipate the higher heat flow of the Archaean was through the existence of a greater total length of spreading centers, or more rapid spreading of new oceanic crust, or, more likely, both. McKenzie and Weiss (1975) argue that early convection in the Earth prohibits the formation of large plates, and that the stresses involved allow only small plates to exist. This they point out may account for the rather small size of the greenstone belts compared with modern orogenic belts (which were formed in a period of "large plate" tectonics). The results of the above simplified models support these conclusions.

The great extent of early mare type oceanic crust on the Earth (at least 50%), the extensive thinning of the lithosphere over this same area by basin formation, the enhanced thermal gradients below this thinned lithosphere, and the highly fractured nature of the basin lithosphere all imply that breakup or rifting of the basin lithosphere was likely to have been rapid and extensive. The result was most likely the formation of a large number of small plates, as opposed to the rather small number of large plates of the colder modern era. If crustal updoming over asthenospheric upwellings (Menard, 1973; Dewey and Burke, 1974) had dimensions at this time comparable to those now developing in Africa (Burke and Whiteman, 1973), then each of the major basins of the early Earth could have fractured into a dozen or so small plates. The numbers of large basins on the early Earth suggests that the total number of small plates formed in the early basin lithosphere probably

was several hundred, perhaps more. Rapid motion of these small plates by fast spreading would satisfy the heat flow condition described by Burke et al. (1976), and likewise be compatible with the stress requirements of McKenzie and Weiss (1975).

We therefore suggest that the earliest form of plate tectonics was a high temperature, rapid spreading, "microplate" tectonic phase. This stage occurred shortly after the formation of the basic crustal dichotomy; "microplate" tectonics probably began before 3.5 billion years ago.

Spreading rates in the early Archaean could have been an order of magnitude greater than they are today (McKenzie and Weiss, 1975). Subduction rates would have been comparably higher 3.5 billion years ago, and collisions between continental (highland) blocks would therefore have been much more violent than today. The zone of rocks seriously affected by the recent collision between India and Asia is measured in hundreds of kilometers; this is comparable to the size of the continental microplates expected following basin rifting. Therefore rapid collisions between very small plates could be expected to completely rework the rocks over the entire extent of the blocks, leaving little trace of previous events. This may well account for the lack of rocks dating from the basin-forming and filling period. It is unlikely that any direct record exists of the earliest microplate stage of evolution.

The Earth cools with time; as heat is lost the rate of spreading

should decrease and collisions between the (already) reworked continental plates would become less violent. The extent of reworking of these blocks should decrease. Small blocks become sutured together into larger units, which are affected by collisions over a progressively smaller percentage of their volume. The centers of these enlarging plates are preserved against the alteration which occurs at their margins as new collisions take place. Over time, large stable units would form, perhaps preserved today as the major shields of the continents. This period of stabilization is reflected in the Archaean geologic record, leading up to a general world-wide stabilization around 2 billion years ago (Reed and Watson, 1975). From this period of time until the present, large plate or Wilson Cycle tectonics have dominated the crustal development of the Earth (Burke and Dewey, 1973; Burke et al., 1976).

In this interpretation the greenstone belts represent early oceanic crust caught in the rapid collisions of the slowly stabilizing microplate period. As suggested by Green (1972) some of these rocks may be terrestrial maria; that is, they represent remnants of the basin-filling volcanics. Glickson (1976) argues that these ultramafic rocks are the products of unusual conditions early in the history of the Earth, and yet post date the basin-forming period because they show no shock metamorphic effects. The small lateral extent of greenstone belts is likely due to the small plate size of the early microplate phase (McKenzie and Weiss, 1975). More recent orogenic belts are longer because plates, and therefore plate margins, are larger.

Some of the implications of this theory should be examined. We suggest a very early establishment of the crustal dichotomy of the Earth. Not only does this imply the early existence of subductable crust (and therefore a possible early beginning of plate tectonics), but it also implies early existence of oceans themselves. Water would naturally drain into these large basins. The highly fractured rocks of the basin rims on the adjacent highlands could be easily eroded, leading to extensive sedimentation at the margins of the basins. Thick sedimentary sequences were then likely to mix with the basin-filling volcanics during continental collisions. Remelting of the dioritic or andesitic rocks of the basin rims could produce early granites which are found as part of some greenstone belts (Green, 1975).

The suggestion that ocean basins existed early in the history of the Earth has implications for the evolution of life on this planet. This has been treated elsewhere (Frey, 1977c). We add only the thought that rapid spreading and high thermal conditions may have aided the early pre-biological evolution of organic compounds.

Finally, the stabilization of a microplate stage over the time period 3.5 to 2 billion years ago, which is consistent with the Archaean geologic record, suggests a slight change in character of the rifting events which are the first stage of Wilson Cycle tectonics. Thickening of the lithosphere during this time required larger plates to become immobile over hotspots prior to rifting. Therefore, there is a greater chance of crustal melting and the termination of the rifting process, as suggested by Burke and Whiteman (1973) for African

rifts today. The number of failed rift systems might thus be expected to increase with time, but considering the poor preservation of the earliest rifting events, it may not be possible to trace such a change through the rock record.

CONCLUSIONS

The late heavy bombardment of the Earth by the same flux of objects which created the lunar mare basins was responsible for the early establishment of the Earth's crustal dichotomy. The Earth would have had more than 50% of an original global low density crust converted into lowlying basins by the bombardment. These basins would flood rapidly due to partial melting at depth triggered by the impact excavation and associated pressure drop. The original ocean basins were mare type basins, and high density "oceanic" crust, which is subductable, was concentrated at the surface nearly 4 billion years ago.

The thinning of the lithosphere by these basin-forming impacts and the enhancement of the sub-basin thermal gradients made the basin lithosphere highly susceptible to breakup or rifting by convective heat transport. Early rifting was likely to have produced a large number of small plates. High heat flow in the past and the inability of large plates to form in a convecting early Earth suggest that the original form of plate tectonics was a high temperature, fast spreading, "microplate" stage which began more than 3.5 billion years ago. As the Earth cooled, this fast, microplate phase gradually stabilized into a slower, thicker, large-plate style of tectonics similar to modern Wilson Cycle evolution of the crust.

It is unlikely that many continental rocks could have survived the earliest, high velocity collisions during the microplate stage. Suturing of small blocks into larger ones led to gradually increased preservation of old rocks. Greenstone belt volcanics may be remnants of the early basin-filling rocks caught between small colliding blocks during the stabilization period around 3.5 to 2 billion years ago.

Plate tectonics in a modified form has existed for most of the Earth's history. 3.5 billion years ago a microplate tectonics dominated the evolution of the terrestrial crust, giving way over time to larger and larger plates that moved progressively slower and slower.

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FIGURE CAPTIONS

Figure 1: Percentage of "new crust" versus mass and diameter of the Earth, Mars and Moon. "New crust" refers to maria on the Moon and Mars and oceanic basalt on the Earth. Data for Mercury and Venus are too scarce to be included; but for the other three planets the relationship is roughly linear.

Figure 2: Heat generated by selected rock types as a function of time before the present. Contributions from U^{235} , U^{238} , Th^{232} and K^{40} only are considered. For granite the individual contributions are shown; for all rock types the total effect from all the radioactive species is given.

Figure 3: Temperature versus time for two models having different initial temperature profiles. Temperatures at various depths (10, 30, 50km) are shown. The dashed line indicates the model values for the case of no impact; the solid line shows the effect of a 1000 km wide, 13 km deep impact basin forming after 500 million years. (a) Initially molten model. (b) Initially colder model, with a solid lithosphere down to about 100 km. See text for details.

Figure 4: Depth to selected temperatures versus time for the initially molten model. D refers to the depth to (partial) melting, here taken to be the base of the lithosphere. Dashed lines indicate no-impact case. Note effect of basin formation on depth to melting immediately after impact.

Figure 5: Change in thermal gradient versus time. Gradients are averaged over 20 km thick layers as shown. Dashed line indicates the no-impact (or sub-highland) case. In this model melting occurs in the 50-70 km deep layer at the time of impact.