GIANT IMPACTS AND THE THERMAL STATE OF THE EARLY EARTH

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The putative giant collision between the proto-Earth and a protoplanet half its diameter that created the Moon 4500 m.y. ago would have had a profound effect on the thermal state of the early Earth. The history of thermal models of the accreting Earth began with consideration of impacts by bodies small enough that most of their energy was radiated to space before it became incorporated in the growing Earth. Such models had difficulty explaining the onset of melting and differentiation in any planet. More recent models examined the physics of energy deposition by impacts of bodies larger than a few tens of kilometers and concluded that a large fraction of their energy is retained as heat. These models indicate that melting should have begun in the Earth when it had grown to about 10% of its present mass. In both types of model the Earth is assumed to grow gradually from bodies much smaller than the growing planet itself. However, recent work on the size-frequency distribution of the planetesimal swarm indicates that most of the mass and energy of the swarm resides in objects at the large end of the spectrum, so that most of the Earth's growth may have occurred in catastrophic impacts with bodies up to half its own diameter. In such collisions heat is deposited deep within the growing Earth, and at the latest stages of accretion the impacts may have melted the entire impacted hemisphere. The detailed effects of such a giant impact are explored using a 3-D numerical hydrocode to simulate the collision between the proto-Earth and a body half its diameter at two different impact velocities.

INTRODUCTION

The idea that the Moon was born in a gigantic collision between the proto-Earth and a Mars-sized protoplanet has continued to gain adherents since it was first proposed by *Hartmann and Davis* (1975) and by *Cameron and Ward* (1976). This theory received a great deal of attention at the 1984 Conference on the Origin of the Moon in Kona, Hawaii (*Hartmann et al.*, 1986), and has subsequently been the subject of a number of recent reviews (*Boss,* 1986; *Newsom and Taylor,* 1989; *Stevenson,* 1987). It is probably not an exaggeration to claim that it has become the current consensus theory of the Moon's origin. Although a great deal of work has been, and is being, done on the details of this scenario, most work up until now has concentrated on the Moon; little consideration has been given to the effects of such a large impact on the Earth. Nevertheless, it is clear that such an event would have profound effects on the Earth, most especially on its thermal state. Although the last word has by no means been written on this problem, in the following pages I present a summary of what is currently known about how impacts, particularly giant ones, may have affected the thermal regime of the early Earth. Although this paper is largely tutorial, near the end I will present some of the results of recent computations I have been doing in conjunction with Marlan Kipp on the early stages of a giant impact and its implications for the early thermal evolution of the Earth. Another view of this event can be found in the paper by *Benz and Cameron* (1990).

ENERGY DEPOSITION BY IMPACTS ON GROWING PLANETS

The history of accretion models of the Earth and planets mirrors a gradually increasing appreciation for the importance of large impacts. In one of the earliest models of the Earth's accretion (*Hanks and Anderson*, 1969; *Urey*, 1952, p. 110 ff) the gravitational energy of the infalling debris was exactly balanced by thermal radiation from the surface of the growing planet. Although these models did not include the effect of conduction into the planet's interior, this was included in later work (*Mizutani et al.*, 1972). I will distinguish models of this type as the "physicist's model" of accretion. Shown in Fig. 1, such models assume material falling onto the planet in time dt is added to the growing planet in thin uniform shells of thickness dR. The mass added in this time is thus $dM = 4\pi\rho R^2 dR$, where ρ is density. Each unit mass deposits a net energy consisting of gravitational binding energy and initial kinetic energy of the mass, $GM/R + \nu_\infty^2/2$, where G is Newton's gravitational constant, M is the planet's mass when it has grown to radius R, and ν_∞ is the velocity of encounter between the growing planet and the infalling material at great distance from the planet (I neglect here any initial thermal energy). The basic equation describing the equilibrium between infall energy, radiation and conduction is

$$\rho\left(\frac{GM(t)}{R(t)} + \frac{\nu_\infty^2}{2}\right)\frac{dR(t)}{dt} = \epsilon\sigma[T^4(R,t) - T_a^4] + \rho c_p T(R,t)\frac{dR(t)}{dt} + k\left(\frac{\partial T}{\partial r}\right)_{r=R} \qquad (1)$$

where T(R,t) is the temperature at the surface of the growing planet, T_a is the effective temperature of the atmosphere, σ is the Stefan-Boltzman constant, ϵ is the emissivity (generally set equal to 1), c_p is heat capacity and k is the thermal conductivity. Radius r is the position *within* the growing planet of total radius R(t). The second term on the right side accounts for the heat content of the hot added material and should include latent heat if melting occurs; it is a standard term in the heat conduction equation when moving boundaries are present (*Turcotte and Schubert*, 1982, p. 174 ff). The thermal state of a growing planet is determined by equation (1) as soon as the growth rate, dR(t)/dt, is specified. Equation (1) implicitly assumes that the added layer of mass is both laterally uniform and infinitesimally thin, assumptions that, as we shall see shortly, lead to major errors.

The studies based on equation (1) had great difficulty explaining why any of the planets should be differentiated. Thermal radiation is so efficient at removing energy that the planets would have had to have grown extremely fast

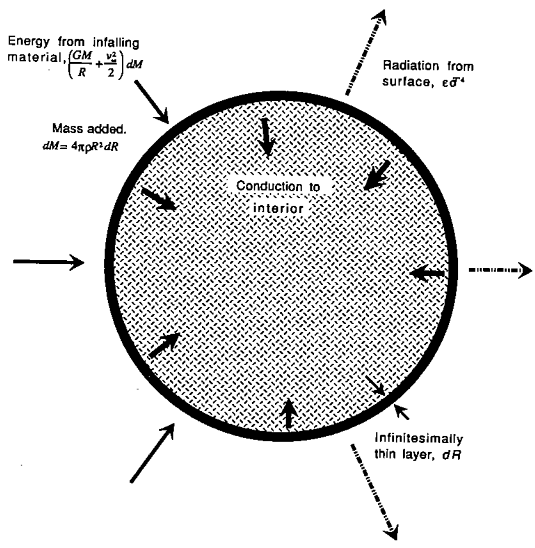


Fig. 1. *The "physicist's model" of planetary accretion, current between about 1950 and 1975, assumes that the gravitational energy of infalling material is balanced primarily by thermal radiation from the surface with a small additional contribution from conduction. This model is valid only if the planet grows from planetesimals smaller than about 10 km in diameter.*

for temperatures to reach the melting point anywhere in their interiors. Thus, *Hanks and Anderson* (1969) require the Earth to grow in 10^5 to 10^6 years for melting to occur, while *Mizutani et al.* (1972) find the Moon must accrete within 1000 years if melting is to occur in its outer portions. These times are far longer than the 10^7- to 10^8-yr timescale derived from standard planetesimal accretion models (*Wetherill,* 1980) and, taken at face value, suggest that the Earth and Moon might have accreted cold, after which the Earth differentiated when enough radiogenic heat had accumulated to cause internal melting. On the basis of this model the Moon never differentiated, a prediction that was quickly proved wrong when the Apollo missions returned Moon rocks to Earth for analysis.

The solution to this quandary was discovered by *Safronov* (1969, 1978) and more recently was elaborated by *Kaula* (1979). They realized that the physicist's idealization of the gradual addition of infinitesimally thin global layers is not a valid representation of the actual process in which individual planetesimals impact the planet's surface. Instead, each impact deposits an ejecta blanket of finite thickness and localized extent in addition to heating the target rocks directly beneath the impact site, as shown in Fig. 2. Under these conditions radiation is able to cool only a thin layer on the top of the ejecta blanket before yet another ejecta sheet is deposited on top of it. A large fraction of the initial energy of the infalling material is thus retained in the overlapping ejecta blankets.

It is easy to estimate the thickness δ of an ejecta sheet for which heat retention outweighs radiative loss. If the planet grows at a rate dR/dt the average time interval Δt between deposition of ejecta sheets of thickness δ at any given site is simply $\delta/(dR/dt)$. But the conductive cooling time of an ejecta sheet is of order δ^2/κ, where κ is the ejecta's thermal diffusivity. This cooling time equals the average time between deposition events when $\delta_0 = \kappa/(dR/dt)$. Layers thinner than δ_0 radiate all of their heat before the next ejecta deposition event, whereas thicker layers do not have time to cool between events. Making the conservative assumption that the Earth grew over a period of about 100 m.y. and that $\kappa \approx 10^{-6}$ m^2/sec yields a crossover thickness δ_0 of about 3 km. The maximum ejecta thickness

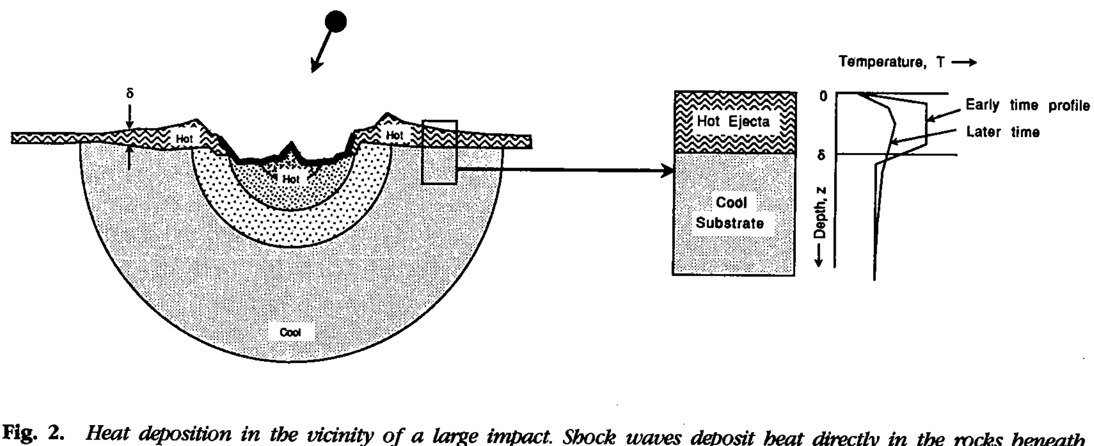


Fig. 2. *Heat deposition in the vicinity of a large impact. Shock waves deposit heat directly in the rocks beneath the crater, the amount of heating decreasing with increasing distance from the impact site, illustrated by the schematic temperature contours. The heavy black line inside the crater indicates a thick layer of melted target rocks. Such layers are observed inside large fresh lunar craters such as Copernicus and Tycho. The ejecta also contain a considerable amount of heat and thickly blanket the terrain to a distance of about 1 crater diameter from the rim. The temperature in the ejecta blanket (inset) declines as a result of thermal radiation from the surface and thermal conduction into the cooler substrate.*

of a fresh crater is about 0.02 of the crater diameter, which itself is about 10 times larger than the projectile diameter (*Melosh*, 1989), so the crossover condition is met by planetesimals about 15 km in diameter. Moreover, much of a projectile's energy is deposited more deeply beneath the crater floor than in the ejecta blanket, so that it is plausible that most of the heat added by 10 km or larger diameter planetesimals is retained by the growing planet rather than lost by radiation.

The details of the overall heat deposition process are complex and depend sensitively on impact mechanics. Thus, of the projectile's initial kinetic energy, about 30% is initially partitioned into the kinetic energy of the ejecta and most of the remainder is directly deposited as heat in the target rocks (*Melosh*, 1989, p. 66 ff). The kinetic energy of the ejecta is converted into heat when it comes to rest on the surface after mixing with a variable amount of preexisting surface material, resulting in a mixed sheet of hot rock debris thickly covering the surface within one or two crater diameters of the crater's original rim. The rocks beneath the crater (and the expelled ejecta) are directly heated by the shock from the impact which, depending upon the impact velocity, may melt or vaporize large quantities of material. The amount of material thus heated depends mainly upon the projectile's size and the square of its velocity (*Melosh*, 1989, p. 122 ff). On the other hand, a large impact cools the target planet by raising deeply-buried materials closer to the surface, where their heat may be more readily conducted to the surface and ultimately lost by radiation. Some of the initial heat of the ejecta may be lost by radiation during its ballistic flight from the impact crater to its site of deposition, while at high impact velocities a portion of the ejecta (especially that in the vapor plume) may travel at velocities greater than escape velocity and thus leave the planet entirely. The net thermal effect of an impact is thus a sum of gains and losses that tend to offset one another and hence make accurate estimation of the net heat deposition difficult. Crude consideration of these processes indicates that large impacts are more efficient at depositing energy than small impacts, and that for kilometer-sized planetesimals the radiation term in equation (1) is almost entirely negligible.

Kaula's (1980) solution to the uncertainty in estimating the net heat deposition of impacts is to lump all the poorly known processes into a single numerical factor h and write equation (1) in a form that neglects both radiation and conduction. Rearranging (1)

$$T(R,t) = \frac{h}{c_p}\left(\frac{GM(t)}{R(t)} + \frac{\nu_\infty^2}{2}\right) \qquad (2)$$

Note that with the neglect of radiation and conductive heat loss the accretion rate $dR(t)/dt$ drops out, so that the interiors of planets growing over the 10^8-yr accretion timescale may easily reach temperatures high enough to initiate melting and differentiation. The unknown dimensionless factor h must lie somewhere between 0 (no net burial of heat) and 1 (all kinetic energy of infalling matter is retained as heat). In the absence of better information, h is generally assumed to be about 0.5 for kilometer-sized or larger planetesimals (*Stevenson*, 1981).

In *Safronov's* (1969) theory the random velocity component ν_∞ of approaching planetesimals at any time t is proportional to the escape velocity $\nu_{esc} = (2GM/R)^{1/2}$ of the planet growing in their neighborhood, so the entire right-hand side of equation (2) is proportional to GM/R times factors of the order of unity. Since M is proportional to $R^3(t)$, the surface temperature T of a growing planet at time t is proportional to $R^2(t)$. This temperature is "locked in" at radius $r = R(t)$ as more material accumulates on the former surface of the planet, establishing an internal temperature distribution of the form $T(r) \sim r^2$. This relation holds until T approaches the melting temperature in the outer portion of the planet, at which time convection begins and the mantle temperature remains near the melting point (*Kaula,* 1980). Melting in the Earth begins when it has reached about 10% of its final mass, or about half its final diameter. A core is presumed to form at about this time, further heating the mantle and stirring it so that the original $T \sim r^2$ thermal structure is wiped out in a manner similar to that described by *Stevenson* (1981).

Subsequent to core formation the mantle temperature must have been closer to the solidus than the liquidus, since convection in a completely liquid mantle is so vigorous that its cooling time is only about 10^4 years (*Tonks and Melosh,* 1989), far shorter than the timescale for Earth's growth from planetesimals. Once more than about 50% of the mantle material crystallizes, however, convection is regulated by the high viscosity of solid-state creep in the crystalline fraction, thus greatly lowering the cooling rates. Craters formed by impacts subsequent to core formation would therefore have excavated either the conductive boundary layer or, if sufficiently large, the semimolten convecting mantle. The extent to which the deposition of impact energy is altered in a planet with a hot mantle has not yet, to my knowledge, been investigated [although *Minear* (1980) studied the disrupting effect of impacts on the boundary layer of a cooling lunar magma ocean, concluding that impacts decrease the cooling time], nor has the related problem of convection in the presence of rapid accumulation of thick ejecta sheets been studied.

Although equation (2) is more realistic than equation (1), it is nevertheless limited by the implicit assumption that the impacting planetesimals are small in comparison to the growing planetary embryo. Impact energy is still assumed to be added in thin (although not infinitesimally thin) shells that are, on average, uniformly distributed over the planet's surface (Fig. 3). Thermal radiation losses can be neglected because most of the infalling planetesimals' energy is buried below the surface, but the overall pattern of energy deposition is much the same.

In recent years, however, it has seemed increasingly likely that the distribution of planetesimal sizes follows a rough power law extending from the smallest sizes up to objects half the size of the largest planetary embryo (*Hartmann and Davis,* 1975; *Wetherill,* 1985), at least during the later stages of accretion. This power distribution is expected to be of the form $N_{cum}(D) = CD^{-b}$, where N_{cum} is the number of planetesimals with diameters greater than or equal to D, C is a constant, and the power b is frequently observed to be close to 2 in numerical simulations of accretion processes (*Greenberg et al.,* 1978), impact fragmentation experiments (*Fujiwara* 1986), and in the size distribution of comet nuclei (*Delsemme,*

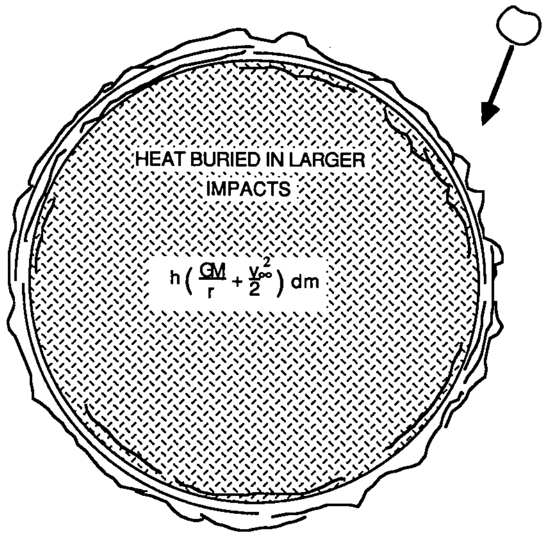


Fig. 3. *A more realistic model of planetary accretion in which a large fraction of the planetesimal's energy is trapped in the growing planet as heat. Melting temperatures are reached by the time the Earth has grown to about 10% of its present mass. In this model the planetesimals are assumed to be much smaller than the diameter of the planet so that growth is gradual; no provision is made for the effects of very large impacts that deposit their heat catastrophically through a substantial depth of the planetary embryo's mantle.*

1987). The $b = 2$ distribution has the special property that the total surface area of planetesimals in successive logarithmically increasing size intervals is constant. That is, the surface area of planetesimals with diameters between, say, D_1 and $2D_1$ is the same as the surface area of planetesimals with diameters between $2D_1$ and $4D_1$ (see Fig. 4). It seems natural that a distribution of this kind should arise from processes such as impact or coagulation that depend upon cross-sectional area (*Chapman and Morrison,* 1989, p. 57).

While the largest numbers of planetesimals in a $b = 2$ distribution are concentrated at the small sizes, most of the mass and energy resides in the largest objects (see Fig. 5). Thus, although a growing planetary embryo would be constantly battered by small planetesimals, most of the overall mass and heat transfer would occur in large, rare events that bury their heat deep within the embryo's interior. If this catastrophic mode of growth were indeed important, then the assumption that the infalling planetesimals are much smaller

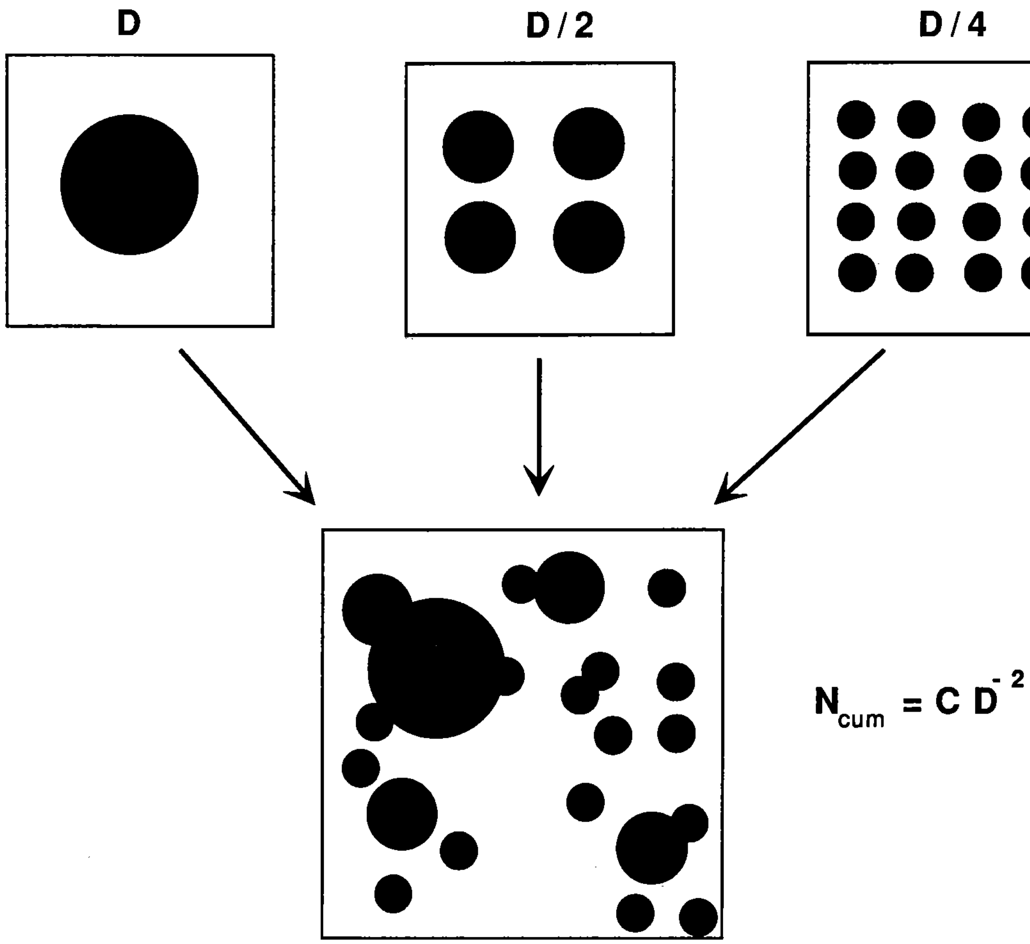
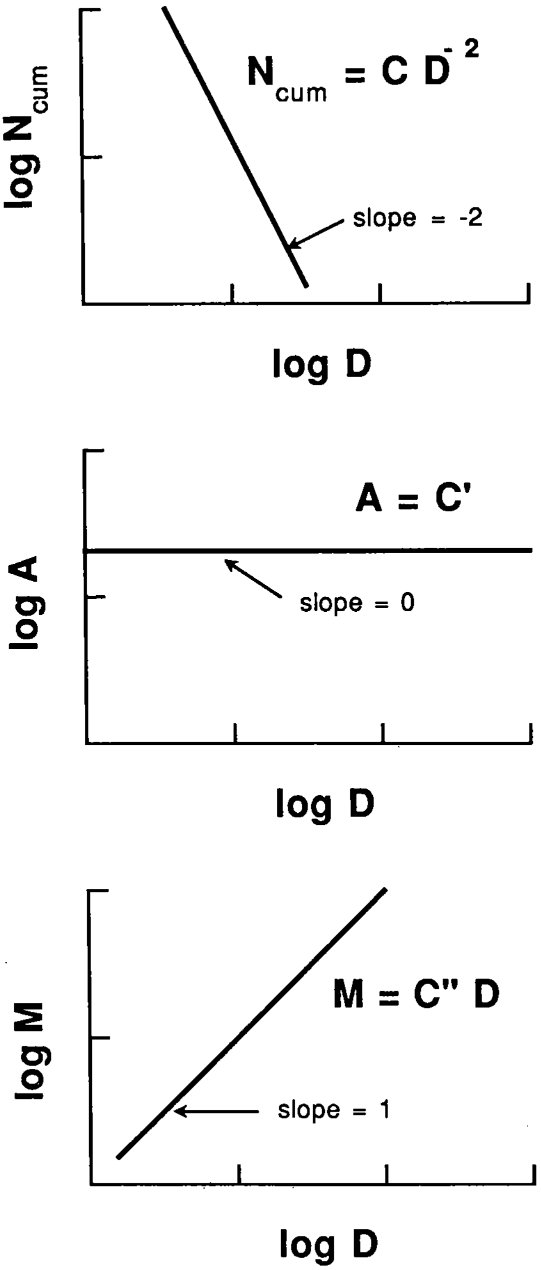


Fig. 4. *Schematic illustration of the distribution of planetesimal sizes in a population described by a $b = 2$ cumulative number distribution. In this distribution the projected area of planetesimals in logarithmically decreasing size intervals is constant.*

than the growing planet fails for the most significant events and a revised thermal analysis must take account of impacts between objects of comparable size. In the following pages I take the first step in such an analysis by examining the effects of a single such "giant impact" between the proto-Earth and an object half its size. Because the direct deposition of heat from a shock wave plays a major role in this process, I begin with a short review of the physics of shock heating.



PHYSICS OF SHOCK HEATING BY IMPACTS

Physics of Shock Compression and Release

During the earliest phases of an impact the incoming projectile meets the target and decelerates, transforming much of its initial kinetic energy into heat through sudden compression. Initially deposited in a region roughly comparable to the projectile in size, the impact energy spreads outward as shock waves advance into the target. Target material engulfed by these shock waves is at first suddenly compressed, then released more slowly to ambient pressure. The three equations describing this sudden compression were first derived by P. H. Hugoniot in 1887, and express the conservation of mass, momentum and energy (respectively) across the shock (*Melosh,* 1989)

$$\rho(U - u_p) = \rho_0 U \tag{3a}$$

$$P - P_0 = \rho_0 u_p U \tag{3b}$$

$$E - E_0 = \frac{1}{2}(P + P_0)\left(\frac{1}{\rho} - \frac{1}{\rho_0}\right) \tag{3c}$$

where ρ is material density behind the shock, ρ_0 is initial density, U is shock velocity, u_p is particle velocity behind the shock, P and P_0 are shock and initial pressures, respectively, and E and E_0 are specific internal energy (per unit

Fig. 5. *The dependence of various quantities on planetesimal diameter D in a population where the cumulative number N_{cum} is a function of D^{-2}. In such a population the number distribution is dominated by the smallest objects, the area (either total surface or cross-sectional) is uniformly distributed, and the mass distribution is dominated by the largest objects.*

mass) behind and before the shock. These equations are often written in terms of the specific volumes $V = 1/\rho$ and $V_0 = 1/\rho_0$ rather than density.

Although shock compression conserves mass, momentum, and energy, it does not conserve entropy; shock compression is an irreversible process that cannot be represented as a continuous path on a thermodynamic diagram, and in which heat is irreversibly deposited in the compressed material. In contrast, the slower release from high pressure is reversible and adiabatic, so that although the temperature declines as the pressure decreases, no net heat is transferred to the material during this release.

The three Hugoniot equations involve five unknowns and so are not sufficient to determine the thermal state of shocked material. One unknown such as P or u_p is established by the boundary conditions. In addition, a thermodynamic equation of state is needed to complete the system. This equation contains all of the material and chemical complexity of the shocked material, and is conventionally written $P = P(V,T)$, where the specific volume $V = 1/\rho$ and temperature T are thermodynamic state variables. For given initial conditions in an impact the equations (3a-3c) and the equation of state determine the thermodynamic state of material behind the shock wave. The release adiabat is computed from a simple relation derived from the first law of thermodynamics

$$\left.\frac{dE}{d\rho}\right|_S = \frac{P}{\rho^2} \qquad (4)$$

where the derivative with respect to density ρ is taken at constant entropy S. Equation (4) can be integrated from the initial shock state to some desired final pressure or density. Most equations of state used in impact computations are written in the convenient form $P = P(\rho,E)$, even though E is not strictly a state variable (*Melosh,* 1989, Appendix II). This form makes integration of equation (4) straightforward using elementary numerical techniques.

The best equation of state currently available for impact computations is generated by an approximately 3000-line FORTRAN computer code called ANEOS (*Thompson and Lauson,* 1972). This code requires 24 input parameters to describe a given substance, from which it computes a numerical approximation to the Helmholtz free energy. Unlike other impact equations of state, thermodynamic quantities such as pressure, temperature, entropy, and internal energy are derived from the free energy by standard thermodynamic relations, so they are guaranteed to be consistent with one another, even through phase transitions. The ANEOS parameters for dunite were previously constructed and the resulting equation of state was found to agree well with the existing data (*Benz et al.,* 1989, Appendix I). Dunite is regarded as a reasonably good approximation to the Earth's mantle.

Three Hugoniot curves derived from this ANEOS equation of state are shown in Fig. 6. This plot shows the shock pressure as a function of particle velocity. This form is particularly useful here because, for the impact of two objects composed of the same material, the particle velocity is precisely half the normal component of the impact velocity. Thus, for an impact between the Earth and a large planetesimal the particle velocity is in the vicinity of 5 km/sec, and the corresponding shock pressure of material starting from surface conditions is a few hundred GPa. The dashed line on this figure shows the Hugoniot curve for material starting at conditions applicable deep in the Earth's mantle, $P = 100$ GPa and $T = 3000°K$. For such material the shock pressures are higher, nearly 500 GPa at the same particle velocity (note that the shock pressure plotted is the *increase* above the ambient pressure). The gray line is for material starting at $P = 100$ GPa and $T = 298°K$. It is clear from this curve that initial temperature has very little effect on the

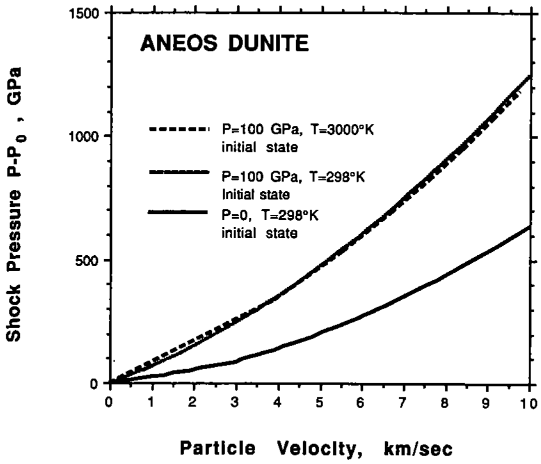


Fig. 6. *Hugoniot curves for dunite computed by the ANEOS equation of state package described in the text. The particle velocity is one half of the impact velocity for a head-on impact between objects composed of the same material. The solid curve is for material that starts at Earth-surface conditions, the dashed curve is the Hugoniot for material whose initial conditions are appropriate for the Earth's deep mantle, and the gray curve is for cold material at deep mantle pressures, illustrating that starting temperature has little effect on the Hugoniot.*

Hugoniot curve; the main variable affecting the curve is initial pressure.

The release curves are shown on a P-T plot in Fig. 7 for material starting at surface conditions, and the residual temperatures are plotted in Fig. 8 as a function of the shock pressure. It is clear that shock pressures in the vicinity of 200 GPa are sufficient to raise mantle material starting near the Earth's surface to the liquid-vapor phase boundary. A weaker shock of 100 GPa suffices to melt it. State changes in the deep mantle are less easy to gauge, as the liquidus is not known at pressures much above 25 GPa (*Ito and Takahashi,* 1987), although recent work on Perovskite, $MgSiO_3$, pushes the melting curve to nearly 100 GPa (*Knittle and Jeanloz,* 1989). Recent estimates of the current temperature at Earth's core-mantle boundary establish a lower bound of 3800°K (*Williams et al.,* 1987), at which temperature the mantle is still evidently in the solid phase. Nevertheless, shock pressures of a few hundred GPa result in temperature increases of 1000°K or more, which I here presume leads to melting of the deep mantle, even if it was not molten to begin with.

Shock Heating in Impacts

The pattern of pressure decline near an impact is a complex function of impact geometry and target material. This pattern is best understood for vertical impacts on a planar target, a geometry that is not particularly relevant to an oblique giant impact between the growing Earth and a large planetesimal of comparable size. Nevertheless, understanding the simple geometry may aid in interpreting the outcome of a more realistic impact geometry.

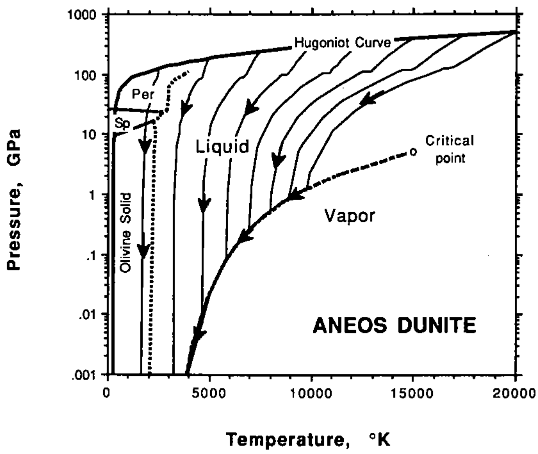


Fig. 7. *The Hugoniot curve and release adiabats for ANEOS dunite as a function of pressure and temperature. Most of the release adiabats intersect the liquid-vapor phase boundary and follow the curve to lower pressure as the vapor condenses. The liquid and vapor phase boundaries and critical point are idealized to approximate those of a simple material. In reality melting and vaporization of silicates is incongruent, although this subtlety is not easily shown on a plot covering such a broad range of P and T. Also shown is the computed critical point and the melting curve of candidate mantle materials up to 25 GPa as determined by Ito and Takahashi (1987). The Spinel-Perovskite phase boundary is from Ito and Takahashi (1989), and the melting curve from 25 to 100 GPa is from Knittle and Jeanloz (1989).*

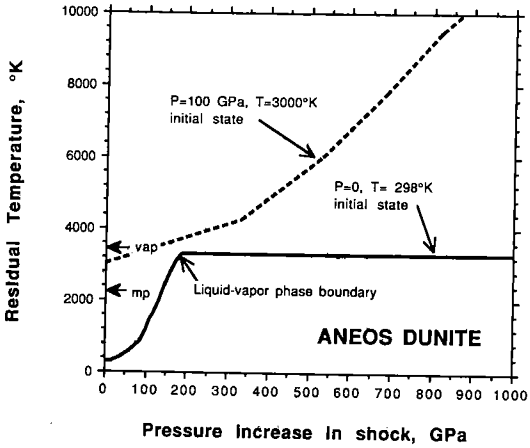


Fig. 8. *Residual temperatures after release from the shock pressure on the horizontal axis to a final pressure of 1 bar (0.1 MPa). The solid curve is for material that starts at Earth surface conditions and the dashed curve is for material that starts at conditions approximating those of the Earth's deep mantle. The plateau in the solid curve develops because both liquid and vapor are present at 1 bar over the indicated range of shock pressures. The temperature of 3380 K is the vaporization temperature of dunite at 1 bar. There is no corresponding plateau at 110 GPa, where melting begins, because the ANEOS is unable to treat both a melt transition and solid state transition simultaneously. In any event this step is small due to the relatively small enthalpy of melting compared to the enthalpy of vaporization.*

Both numerical computations and data from a limited number of experiments show that in a vertical impact a nearly isobaric core of high pressure develops around the site of the impact. This core is comparable in size to the projectile, and the mean pressure in this region is accurately given by the planar impact approximation (*Melosh,* 1989, p. 54 ff). Although islands of higher pressure develop during the earlier jetting phase, most of the projectile mass (and a roughly equal volume of the target) is shocked only to the mean pressure. As the shock waves from the impact spread out they weaken both because their energy is diluted as they engulf progressively more material and because of irreversible energy deposition in the target. The rate of this weakening is a function of maximum pressure and material, but is approximately given by

$$P(r) = P_{core}\left(\frac{a}{r}\right)^n \qquad (5)$$

where P_{core} is the mean pressure in the isobaric core, a is the projectile radius, and r is now the distance from the impact site. The decay power n is typically in the range of 2 to 3, so that the pressure declines by a factor of 10 at a distance of 2 to 3 projectile radii from the impact site. This approximation is similar to the successful "gamma model" previously proposed by *Croft* (1982), except that in the gamma model r is the distance from a point located a distance equal to the projectile radius a below the surface.

Impacts at velocities on the order of 10 km/sec (particle velocity ~5 km/sec) in dunite generate pressures P_{core} on the order of 300 GPa and will thus melt and partially vaporize a few projectile masses of the target, but such melting will be confined to regions within about 2 projectile radii of the impact site. More distant regions of the planet will be warmed by impact heating, but will not necessarily melt unless they are close to melting already (see Fig. 9). Large impacts thus deposit their energy deeper than small impacts, and impacts with objects half the size of the proto-Earth can be expected to melt at least one hemisphere of the Earth's mantle right down to the core. The melt pool following a large impact will undergo further change in shape after it forms, since the melt will have a different (generally smaller) density than surrounding rocks at the same depth. Since the surrounding rocks were likely to be hot, and therefore relatively fluid, subsolidus viscous deformation subsequent to the melting event should close the initial melt-solid crater, producing a global magma ocean of nearly uniform depth overlying hot, more dense solid mantle material. Although the timescale of this relaxation is difficult to estimate, is was probably not shorter than the timescale of post-glacial relaxation in the present Earth; that is, a few thousand years.

Core formation in the context of this model is relatively quick and simple: Metallic iron trapped in the outer regions of a planet quickly sinks toward the bottom of any large melted region immediately following a large impact. The only requirement for core formation is thus the impact of a projectile sufficiently large and fast to melt a substantial portion of the planet. The precise time of core formation thus contains a stochastic element, although it cannot occur too early as encounter velocities must be large enough to generate melting without completely disrupting the planetary embryo.

EFFECT OF A GIANT IMPACT ON EARTH'S THERMAL STATE

The collision between a Mars-sized protoplanet and the proto-Earth adds a truly prodigious amount of energy to the Earth over a time interval measured in hours. The mass m, velocity v, and impact parameter b of the projectile are constrained only by the total angular momentum L of the Earth-Moon system, 3.49×10^{34} kg m^2/sec. Since the angular momentum is a product of all three terms, L = mvb, the value of each individual quantity is uncertain within broad limits. The impact parameter is bounded between zero and the sum of the Earth's and the projectile's radii, while the impact velocity is at least as large as the proto-Earth's escape velocity but, because of the overall geochemical similarity of the Earth and Moon, it was probably not as much as twice the escape velocity (i.e., the projectile had an initial orbit close to that of Earth's). These constraints point to an impact by a body with a mass about 10% of the Earth's mass, hence a diameter approximately half of the Earth's—about the size of Mars.

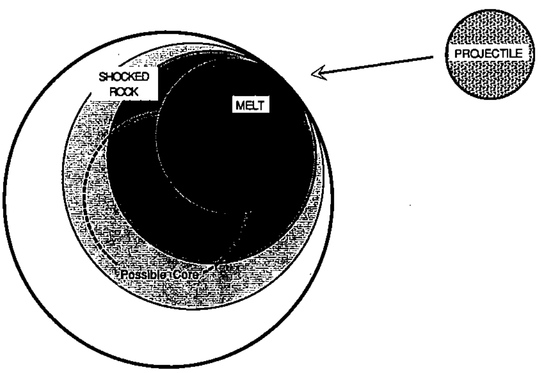


Fig. 9. *Schematic illustration of the pattern of heat deposition in a planet struck by a projectile of comparable size (1/4 its diameter in this figure). Adjacent to a melted region roughly twice the projectile's diameter, the shock level, and thus temperature, falls off steeply with increasing distance from the impact site.*

The energy released in such a collision is the sum of gravitational and kinetic energies, and was probably within the range 2×10^{31} to 5×10^{31} J. Averaged over the Earth's mass this is about 7.5×10^6 J/kg. Several authors (including myself) have used this energy in conjunction with a single value of silicate heat capacity to infer rather high average temperatures for the Earth. Thus, a silicate heat capacity of 10^3 J/kg K yields an average temperature of 7500 K for the Earth. However, such estimates neglect the relatively small latent heat of melting, $\sim 4 \times 10^5$ J/kg, the large latent heat of vaporization, $\sim 5 \times 10^6$ J/kg, and the rapid increase of heat capacity at temperatures above the Debye temperature (taken to be 676 K for dunite in ANEOS). According to the ANEOS equation of state, an internal energy of 7.5×10^6 J/kg corresponds to a temperature rise of about 3200 K for dunite near the Earth's surface and about 7000 K deep in the mantle where vaporization does not limit the temperature rise. Nevertheless, it is clear that melting, even vaporization, should be widespread in a giant collision.

Although the total energy available from the collision of a Mars-sized projectile with the proto-Earth is impressive, the *distribution* of the energy within the Earth is equally important. If, as has been suggested by *Stevenson* (1987), this energy is mainly expended in vaporizing the projectile, the Earth may acquire a transient silicate vapor atmosphere without strongly heating the deeper mantle. The simple considerations of shock wave generation and decay discussed above, however, indicate that only partial vaporization occurs at probable impact velocities, although jetting (*Melosh and Sonnett,* 1986) may enhance the local production of hot vapor, and that deep melting should be widespread in at least the hemisphere that the impact occurs.

An additional factor not previously considered here (but see also *Benz and Cameron,* 1990) is that the projectile might have an iron core that will sink through the Earth's mantle shortly after the collision and merge with the Earth's core, releasing its gravitational potential energy. Assuming a core equal to 30% of the projectile's mass, the energy released by sinking 3000 km through Earth's mantle is of order 3×10^{30} J, which itself will cause strong heating of the mantle through which it sinks and of the Earth's core when it arrives. After the cores have merged, this heat is applied to the bottom of the mantle so that any portion of the mantle that escaped melting by the direct shock wave will likely be melted by this means.

To address the temperature rise in the Earth more exactly, I performed a series of 3-dimensional numerical hydrocode computations in conjunction with M. E. Kipp of Sandia National Laboratory. These computations were designed to simulate the impact between the proto-Earth and a Mars-sized protoplanet. We used the code CTH, implemented on the Cray X/MP supercomputers of the Sandia National Laboratory. This computation uses the ANEOS equations of state for dunite in the mantles and iron in the cores of the two colliding planets. The Earth has a central gravitational field, and is adjusted so that its initial temperature profile is similar to that of the present-day Earth. These models thus start out relatively cold, with mantle temperatures well below the solidus of dunite. We have performed computations at a variety of initial velocities and impact parameters, including pairs that give the Earth-Moon system its present angular momentum.

At the lowest velocity ($v_\infty = 0$), for impact parameters b of 0.88 (Fig. 10a) and 1.25 times the Earth's radius R_e, the strongest heating upon impact is confined to the hemisphere on which the projectile strikes. Shock-induced temperature rises are typically 2000 to 3000 K between the site of the impact and the Earth's core. A crater forms that extends most of the way down to the core. The gravitational energy of this excavation itself is of order 10^{30} J, which appears as heat within an hour or two as mantle material flows inward to fill the crater cavity. Unfortunately, our computations do not extend to long enough times for the entire projectile to merge with the Earth in the high impact parameter runs. This limitation is a result of our finite grid size; by this time a substantial amount of material had left the grid and further computations could not take account of the fallback of this material. In these cases we had to stop the computation while some of the projectile was still falling on portions of the Earth more distant from the impact site. In these cases we believe that more than half the mantle will be strongly heated, so the quoted results must be seen as a lower limit. Figure 10b illustrates the temperature contours for the $b = 0.88R_e$ computation 1800 sec after the impact (this impact parameter and velocity correspond to the angular momentum of the present Earth-Moon system). Note that in this computation a very hot (>4200 K) low-velocity vapor plume is expelled backward from the impact site. This plume eventually spreads over the entire Earth, producing a transient silicate vapor atmosphere.

The results for the higher impact velocity $v_\infty = 7.8$ km/sec are more spectacular. For the impact parameters studied (0.59 and $1.25R_e$) the hemisphere near the impact was heated nearly uniformly by 1000 to 3000 K. The projectile's core was almost entirely vaporized and a much larger crater formed in the proto-

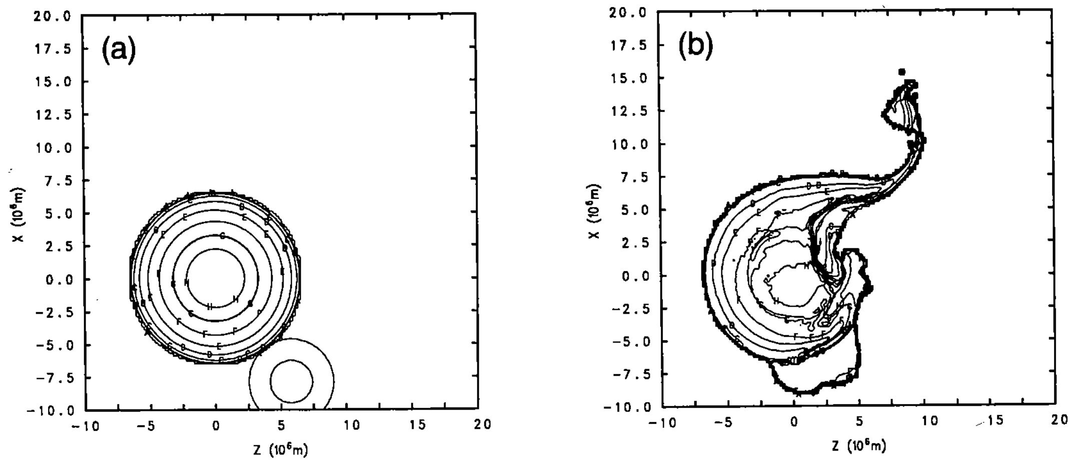


Fig. 10. *Temperature contours in the collision between the proto-Earth and a protoplanet half its diameter. This computation is for $v_\infty = 0$ km/sec (8 km/sec at contact) and an impact parameter of $0.88\,R_e$ at contact.* **(a)** *The initial configuration before impact in which the projectile is traveling upward (positive x-direction) and the proto-Earth is at rest.* **(b)** *The configuration 1802 sec after contact. The contour values are A = 300 K, B = 600 K, C = 1200 K, D = 1800 K, E = 2400 K, F = 3000 K, G = 3600 K, and H = 4200 K. Figures 10 and 11 were computed by M. E. Kipp at Sandia National Laboratory, Albuquerque, NM, using the 3-D hydrocode CTH. These plots are in the symmetry plane of the two colliding spheres.*

Earth. A fast, hot vapor plume also carries several lunar masses of material out along trajectories that eventually take up elliptical orbits about the Earth. Figure 11b illustrates temperature contours for $b = 0.59R_e$, corresponding to the angular momentum of the present Earth-Moon system, at 1200 sec after the impact. Again, a hot low-velocity backward vapor plume is formed that will eventually cover the Earth's surface.

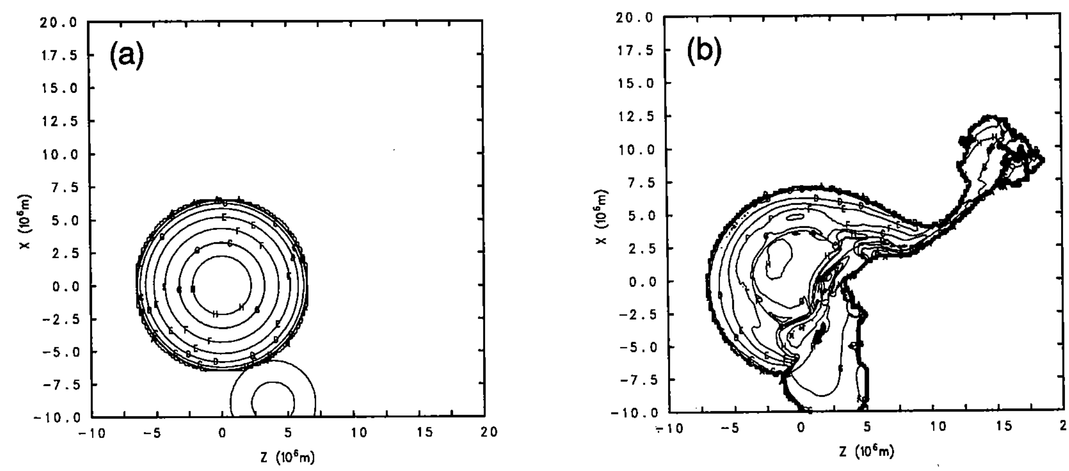


Fig. 11. **(a)** *Initial configuration and* **(b)** *temperature contours 1193 sec after the impact between the proto-Earth and a protoplanet half its diameter moving at $v_\infty = 7.8$ km/sec (12 km/sec at contact) and an impact parameter of $0.59R_e$ at contact. Contour values are the same as in Fig. 10.*

CONCLUSION

The 3-D hydrocode computations, in conjunction with the more general considerations described above, indicate that a Moon-forming impact would have had a profound effect on the Earth's thermal state. The shock produced by the impact would have heated the Earth to great depths, raising at least the hemisphere adjacent to the impact above the melting temperature. Later phenomena, such as the merger of the projectile's and proto-Earth's cores and the collapse of the mantle-deep crater created by the impact, would have added comparable amounts of energy to the Earth. There seems to be no way to avoid the conclusion that a large Moon-forming impact is inevitably accompanied by widespread melting of most or all of the Earth's mantle. This scenario has implications for the geochemistry of the Earth that are dealt with in an accompanying paper (*Tonks and Melosh,* 1990).

The effects of "giant" impacts (i.e., impacts by bodies roughly half the diameter of the primary) on the thermal state of a growing planetary embryo is, however, a more general problem than that of the hypothetical Moon-forming impact on the proto-Earth. If the cumulative spectrum of planetesimal sizes is close to a power law of slope -2, then most of the mass and energy added to a growing planet will be deposited by such "giant" impacts. In this case a simple pattern of temperature vs. radius may never develop, as the thermal state at any given era will depend upon the time, velocity, and obliquity of the last large collision. The process of core formation and differentiation of such a catastrophically growing body may be qualitatively different than that suggested by current gradualist models of planetesimal growth. The implications of such catastrophic growth have yet to be worked out, but it is clear that much more work needs to be done on the effects of large impacts on protoplanets of all sizes.

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