CONSIDERATIONS OF LARGE SCALE IMPACT AND THE EARLY EARTH. R. A. F. Grieve and E. M. Parmentier, Department of Geological Sciences, Brown University, Providence, RI 02912.

Bodies which have preserved portions of their earliest crust indicate that large-scale impact cratering was an important process in early surface and upper crustal evolution. For example, large impact basins form the basic topographic, tectonic, and stratigraphic framework of the moon and impact was responsible for the characteristics of the second order gravity field and upper crustal seismic properties (1). The earth's crustal evolution during the first 800 my of its history is conjectural. The lack of a very early crust may indicate that thermal and mechanical instabilities resulting from intense mantle convection and/or bombardment inhibited crustal preservation. Whatever the case, the potential effects of large scale impact have to be considered in models of early earth evolution.

The number of impacts can be estimated from the cratering record in the lunar highlands, corrected for terrestrial impact conditions. With model approach velocities of 6-10 km s^{-1} for the residual planetesimal swarm (2), it is estimated that the earth will collect 2-4 times as many bodies of the same mass as the moon and the resultant craters will be $\sim$ 1.5 times larger, depending on the energy-diameter scaling relationship used. Assuming the size-frequency distribution of primary craters approximates $N \propto D^{-2}$, this translates to a correction factor of 4.5-9 for the number of craters of equivalent diameter D. From estimates of the cratering rate for the lunar highland (3) and the number of large lunar basins (4), the lower factor of 4.5 predicts $\sim$ 200 basins with D > 1000 km on the earth in the period 4.6-3.9 by. This is considerably more than previous minimum estimates of $\sim$ 30-60 basins with D > 1000 km, which were based on relatively fast approach velocities of 15-20 km s^{-1} (4, 5) and may be unrealistic as it requires a significant number of bodies in the residual planetesimal swarm, compared to the one or more of mass 10^{23}g required for basin formation on the moon (2).

Even for minimal estimates of the number of impacts, the values for additional exogenic energy and impact melt production are of the same magnitude as present-day processes of internal heat losses and island-arc volcanism, respectively (5). Among the effects of large-scale impact on an early terrestrial crust will be the formation of topography, enhancement of the sub-impact thermal gradients in the lithosphere and asthenosphere due to uplift of originally deep-seated materials, and the potential for eruption of basalt due to adiabatic decompression. Given that the earth has a relatively thin lithosphere, which is likely to have been even thinner during its early history, it is possible that large basin-sized impacts could bring the asthenosphere to the surface, leading to volcanism over large areas (4, 5).

Following impact, however, a number of processes act to modify large impact basins. These are: thermal contraction and subsidence from the loss of post-shock and uplift heat, topographic degradation by erosion and relaxation, and loading of the basin. Some of these effects, as they apply to basins on the moon, have been examined with the suggestion that lunar basins have appreciable effects, on timescales of 10^6-10^8 y, on the thermal, volcanic, and tectonic history in and around the basins (6, 7).

Preliminary models of the evolution of a large terrestrial impact basin have been undertaken. A 1000 km basin, with the extent of the sub-impact thermal anomaly constrained by uplift data from smaller terrestrial structures (8) and post-shock waste heat calculations (9), was formed in a model lithosphere 80 km thick with a thermal gradient of 20°C km^{-1}, overlying an asthenosphere with a thermal gradient of 0.5°C km^{-1}. Sub-impact temperatures were calculated as a function of time from finite difference solutions of the transient heat conduction equation in a cylindrical geometry (Fig. 1). Substantial conductive cooling occurs on a timescale of $\sim 10^7$ years. Consideration of the initial thermal anomaly and the melting behavior of anhydrous peridotite, suggests that $\sim 3 \times 10^6$ km^3 of partially to totally molten mantle material will be brought to upper crustal and near surface depths. For more conservative gradients of 15°C km^{-1} and 0.1°C km^{-1}, the initial thermal anomaly results in the immediate production of $\sim 8.5 \times 10^5$ km^3 of at least partially melted peridotite. This is equivalent to an $\sim$ 1 km thick layer of basalt within the basin. This is in addition to the production of impact melt, which for an approach velocity of 10 km s^{-1} is estimated to be of the same order of magnitude.

The initial topography of an 1000 km early earth basin, and how it changes with time, is not known. Some general statements, however, can be made. Isostatic adjustment will occur on a timescale of $2\mu/\rho g D$, where ρ and μ are the density and viscosity of mantle material underlying the basin and g is the acceleration of gravity. For $\mu = 10^{22}$ poise, isostatic rebound times are on the order of 10^4 years. On longer timescales the basin floor will subside due to conductive cooling and thermal contraction. Thermal subsidence will be concentrated near the center, where a maximum subsidence of $\sim$ 2 km occurs after $\sim 10^8$ years (Fig. 2). In addition, the basin may be dynamically loaded by volcanics, sediments and water, which will cause further subsidence. For example, a sediment load with density $\rho_s = 2.7$ g cm^{-3} and $\rho = 3.3$ g cm^{-3} will cause a 1000 km basin with a surface topography of $\sim$ 3 km, a maximum based on depth-diameter ratios of terrestrial complex structures (8), to subside by an additional factor of 1.8; i.e., the original floor will subside to a depth of $\sim$ 5.4 km. This is in addition to thermal subsidence, which will contribute significantly to the thickness of sediments, particularly near the center.

These models do not account for several effects which will be important in determining the rate of heat transfer to the surface. First, loading of sediments into the basin will reduce the rate of heat loss and the rate of thermal subsidence. Subsidence due to loading will occur

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on isostatic timescales ($\lesssim 10^4$ y) that are shorter than the time for conductive cooling of hot mantle beneath the basin (Fig. 1). Thus, basin-filling impact melt, breccia, post-impact volcanics, and sediments may be thermally metamorphosed and even partially melted to generate more sialic magmas. Second, solid-state convective upwelling beneath the basin will result in higher rates of heat transfer to the surface and to the basin fill. Convective flow will also cause radial spreading of the thermal anomaly resulting in the thermal subsidence being less concentrated near the basin center. Hotter mantle beneath the basin is less dense than surrounding mantle by an amount $\Delta\rho = -\rho\alpha\Delta T$, where α is the coefficient of thermal expansion and T is temperature which extends to a depth d. The resulting hydrostatic pressure difference $\Delta\rho gd$ must be balanced by a viscous stress $\mu w/d$ where w is the velocity of convective upwelling. Then, the time for upward convection of the thermal anomaly is $d/w = \mu/\rho\alpha\Delta Tgd$, while the conductive cooling time is d^2/κ, κ being the thermal diffusivity. For the initial thermal anomaly (Fig. 1a.), $d \sim 50$ km and an average $\Delta T \sim 500°C$. With $\mu = 10^{22}$ poise and appropriate values of the other parameters, the time for convective upwelling is shorter than that for conductive cooling by a factor of 10^2. This is a minimum estimate because the mantle viscosity of an early, hotter earth is likely to be less than a value typical of the present earth. Thus convective upwelling beneath the basin should significantly enhance heat transfer to the basin fill and influence both the rate and radial distribution of thermal subsidence.

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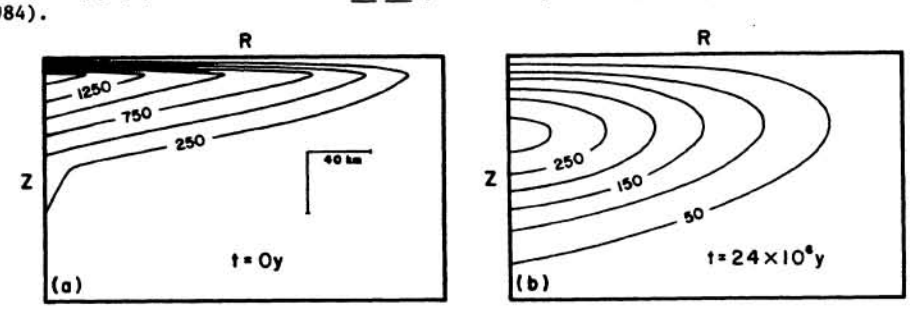

Figure 1. Vertical section of thermal anomaly beneath model 1000 km impact basin. R is radial distance and Z depth. Contours are for temperatures in excess of pre-impact thermal gradient. (a) Initial post-impact anomaly; (b) After 24 x 10^6 y.

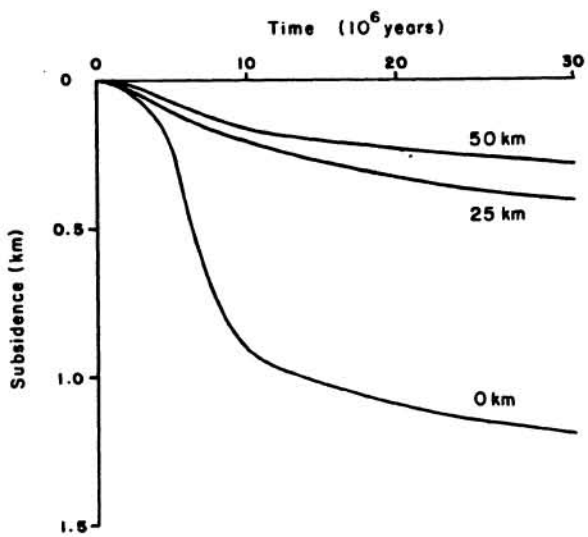

Figure 2. Thermal subsidence at various distances from the center of 1000 km basin, based on post-impact conductive cooling model, as a function of time.