THE THERMAL EVOLUTION OF POST-IMPACT PLANETESIMALS. F. J. Ciesla[1], G. S. Collins[2], and T. M. Davison[2], [1]Department of Geophysical Sciences, The University of Chicago, 5734 South Ellis Avenue, Chicago IL 60430, (fciesla@uchicago.edu), [2]Department of Earth Science and Engineering, Imperial College London, London SW7 2AZ, UK.

Introduction: Short-lived radionuclides undoubtedly played an important role in the heating and metamorphism of planetesimals and meteorite parent bodies [e.g. 1]. However, the role that planetesimal impacts and collisions have had on the thermal evolution of small bodies is less certain. Quantifying the amount of heat deposited in an impact and its subsequent transport through a planetesimal is critical for understanding the chemical evolution of primitive bodies as well as the planets that accreted them.

Keil et al. [2] carried out a series of SPH simulations to examine the amount of heat deposited during the collision of planetesimals. It was found that the volume of heated material in a given collision was small compared to the original volumes of the planetesimals. Further the material which was heated to the highest temperatures (and in some cases, melted) was largely ejected from the largest bodies that survived the collisions, and thus lost to space. Keil et al [2] thus concluded that impacts could not have produced enough heat to explain the large-scale metamorphisim amount of chondrite parent bodies or the igneous processes seen in other meteorite parent bodies.

Recently, we demonstrated that the amount of heating that would occur in planetesimal collisons would depend sensitively on the porosity of the planetesimals [3], with higher porosities generally leading to higher post-impact temperatures. Here we explore how shock-heated material in such collisions evolved dynamically (what fraction was retained versus ejected) and how the heat in the material that was retained was redistributed throughout the planetesimal.

Porous Planetesimal Collisions: As a starting point, we have simulated the collision of a 1 km planetesimal into a 10 km planetesimal, each with a porosity of 50% with the iSALE hydrocode. The results of such a collision is shown in Figure 1, with contours of density (left) and temperature (right) indicating the extent to which materials were processed as a result of the collision. The smaller planetesimal excavates a cavity approximately 3 km deep in the larger one. At the bottom of the crater, a layer ~500 m in thickness remains in which the solids are heated and compressed. Maximum temperatures and densities (T~2000 K, ρ~3.3 g/cm^3, φ=0%) are reached at the surface of the crater, falling off with distance into the planetesimal.

Heat Transfer: In order to determine the post-impact thermal evolution of the surviving planetesimal, we calculate how the heat is transferred within the body over time and eventually radiated away from the surface. To do so, we follow the methods used in other studies of planetesimal thermal evolution [eg. 4-6], by solving the heat equation. However, given the nature of the problem, we can no longer assume spherical symmetry. Instead, the heat equation is solved in 2 dimensions, using a finite-volume scheme in cylindrical coordinates. The planetesimal is assumed to be rotationally symmetric around an axis connecting the center of the planetesimal and the center of the crater.

For the starting conditions, we take the results of the iSale simulation, and define the density of material, porosity, and temperature throughout the planetesimal. Only the thermal energy from the impact is considered; heat produced from the decay of shortlived radionuclides such as ^{26}Al are ignored in order to isolate the effects of shock-heating. For thermal parameters, we assume that solid rock has a thermal diffusivity of κ_{whole} =7x10^{-3} cm^2/s [4] which is then modified by the porosity of the rock, φ, by the equation $\kappa_{por} = \kappa_{whole}(1-\varphi)^2$, which accounts for changes in thermal properties due to complexly connected pores [7]. The non-porous density is assumed to be 3.3 g/cm^3, with the density falling off as $\rho_{por}=\rho_{whole}(1-\varphi)$. The heat capacity of the solid rock is taken to be 7x10^6 erg/g/K.

Figure 2 shows how the heat is redistributed into the planetesimal considered above by considering the temperature evolution of a point 500 m below the bottom of the crater as a function of time. Three different cases are shown, the case where the crater accumulates no porous regolith with time (solid line), the case for 24 m of porous (φ=80%) regolith (dot-dashed) and the case for 48 m of porous (φ=80%) regolith (dashed). Note here that these values of regolith tend to be much smaller in thickness and lead to much higher thermal diffusivities than used in previous studies [4-6]. A layer of 48 m with this porosity is approximately what would be expected to be accreted by the planetesimal during a single orbit at 3 AU in a Minimum Mass Solar Nebula [8].

While the particular results of cooling in the planetesimal are dependent on the assumed thickness of the regolith that accumulates ontop of the crater (from further accretion, mass wasting, etc.), in all cases the planetesimalretains the heat produced from impact for >10^5 years. In fact, for the cases considered here, temperatures in excess of 1000 K are maintained in the planetesimal for time periods of 5000-25,000 years. Such time periods may be long enough for localized metamorphism or localized chemical processing (i.e. devolatilization) to have occurred on the planetesimal.

Discussion: The peak temperatures produced during an impact of porous planetesimals (>~1200 K) are great enough to match the temperatures that all equilibrated chondrites are expected to have reached. While sufficient thermal energy is expected to be delivered in this manner, the cooling rates that these chondrites are thought to have experienced (~1-100 K/Myr at 400-800 K) [9] are slower than those found for the specific case shown here by 1-2 orders of magnitude. However, the cooling rate is highly dependent on details of insulation provided by the surviving planetesimal and the regolith that buries the shock heated materials (from mass wasting or subsequent accretion of dust from the surrounding space). Whether temperatures can be maintained to anneal shock damage as proposed by [10] will also depend on such factors. We will report on the feasibility of such scenarios by further investigating a variety of planetesimal collision as well as regolith accumulation scenarios.

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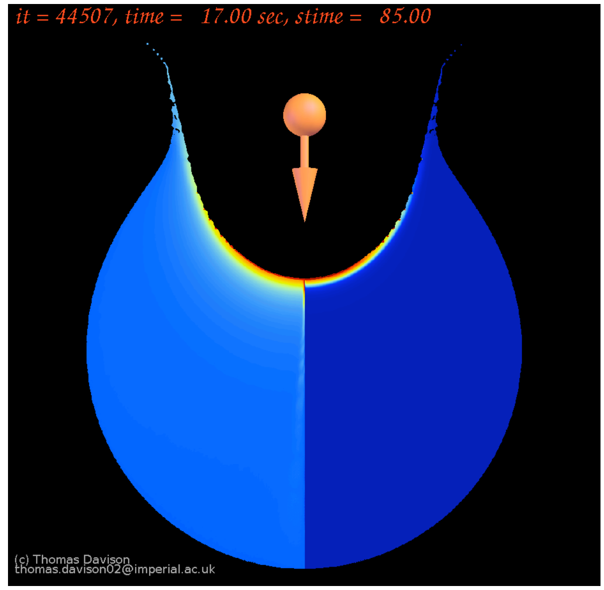

Figure 1: Density (left) and temperature (right) contours of a 10 km planetesimal after being impacted by a 1 km planetesimal at 5 km/s. Both planetesimals began with porosities of 50%. Peak densities are 3.3 g/cm^3 (red), with undisturbed densities of 1.65 g/cm^3. Peak temperatures are ~1500 K (red), with initial temperatures being 293 K (dark blue). The impact creates a heated layer approximately 500 m in thickness which then loses heat by radiation from the surface and conduction into the interior of the planetesimal.

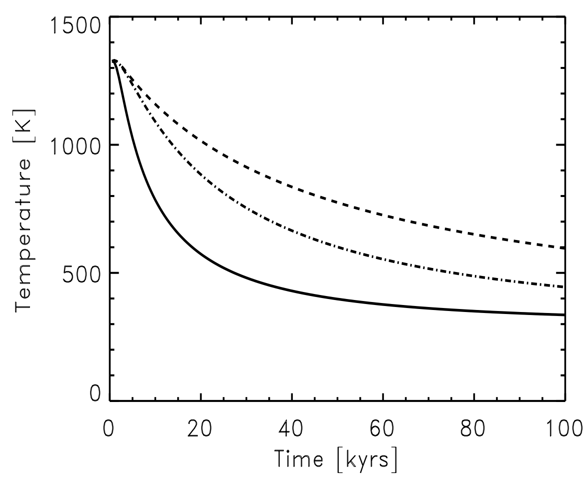

Figure 2: Thermal evolution of a point 500 m below the bottom of the crater that formed in Figure 1. The different curves represent the cooling history for different assumed layers of regolith: none (solid), 24 meters of 80% porous regolith (dash-dot), and 48 meters of 80% porous regolith (dashed).