THE ROLE OF IMPACTS IN THE THERMAL EVOLUTION OF PLANETESIMALS. T. M. Davison[1], F. J. Ciesla[1] , G. S. Collins[2] and D. P. O'Brien[3], [1]Department of the Geophysical Sciences, University of Chicago, 5734 South Ellis Avenue, Chicago IL 60637. [2]Department of Earth Science and Engineering, Imperial College London, London SW7 2AZ, UK. [3]Planetary Science Institute, 1700 E. Ft. Lowell, Suite 106, Tucson, AZ 85719.

Introduction: Short-lived radionuclides, namely ^{26}Al, undoubtedly played an important role in the thermal evolution of planetesimals in the early solar systems. Indeed, numerical models of the thermal evolution of asteroidal sized bodies heated by the decay of ^{26}Al are able to roughly reproduce the peak temperatures and cooling rates that have been inferred from thermochronometry studies of meteorites [1-3]. Impacts and collisions between planetesimals have also been discussed as a potentially important heat source for planetesimals [4-6], as they are expected to be frequent and energetic events in the early evolution of the solar system. Further, such collisions are thought to explain the discrepancy between the range of cooling rates predicted in short-lived radionuclide thermal models and the inferred evolution from studies of meteorites.

While Keil et al [7] discounted impacts as important heat sources on planetesimals because the amount of energy deposited in a single collision would only produce a small (<100 K) change in temperature when averaged over the entire target body, Davison et al. [8] demonstrated that impacts could locally heat temperatures to higher values than previously realized if planetesimals were porous. Indeed, given that planetesimals are expected to form through low-velocity collisions of smaller bodies, possibly in self-gravitating clumps [9-11], first generation planetesimals are likely to have significant pore space, which must be taken account in thermal models.

Here we extend the work of Davison et al. [8], who determined the peak temperatures that would be produced in impacts between porous planetesimals, by calculating the post-impact thermal evolution of the target body. By calculating how heat is redistributed throughout the target body with time, we can, for the first time, determine the volume of planetesimal that is heated by the impact, and the resulting peak temperatures and cooling rates of that volume of material.

Impact Simulations: Davison et al [8] performed a series of impact simulations between porous planetesimals using the iSALE hydrocode. The planetesimals were assumed to be composed of porous dunite, with porosities ranging from φ=0 to 80%. A range of impact conditions were considered, with target:impactor mass ratios ranging from 0.01 to 1, and impact velocities from 0.5 to 8 km/s. For each case considered, the simulation was run until the shock wave had reached the back of the target planetesimal. Further simulations have been performed considering the gravity of the planetesimals, which were run until after complete collapse of the crater [12].

Planetesimal Thermal Model: In order to determine the post-impact thermal evolution of the surviving planetesimal, we solve the heat equation, allowing for the diffusion of energy within the planetesimal to the surface, where it is then radiated away. This is a similar approach to those models that investigated heating due to short-lived radionuclides [1-3], however, given that the impact occurs on one side of the planetesimal, we cannot assume spherical symmetry as was done in those studies. 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.

The starting conditions for a given model run are taken directly from the output of an iSALE simulation, and thus the density of material, porosity, and temperature are defined from the hydrodynamic simulation of the impact. For material parameters, we assume that solid rock has a thermal diffusivity of κ_{whole} =7x10^{-3} cm^2/s [13] which is then modified by the porosity of the rock, φ, by the equation $\kappa_{por} = \kappa_{whole}(1-\varphi)^2$, therefore accounting for changes in thermal properties due to complexly connected pores [14]. The heat capacity of the solid rock is 7x10^6 erg/g/K, a typical value used in thermal models of chondritic meteorite parent bodies [1-3]. A radiative boundary condition is used.

Model Simulations: Figure 1 shows the physical structure of and temperature distribution within a target planetesimal that originally measured 100 km in diameter after it was struck by an impactor 10 km in diameter at 5 km/s. Both bodies were assumed to have an initial porosity of 50%. An initial temperature of 293 K was assumed everywhere. Currently, we ignore radiogenic heating to isolate the thermal effects of impacts. The impact occurred vertically in this image, with the impactor coming from the top of the Figure.

The maximum temperature in the target body, ~1900 K, is reached at a depth of roughly 25 km below the final surface of the surving body. Temperatures are more or less constant in the 25 km of material above this point, though the surface rapidly begins to cool as it radiates away energy. The hot material that is buried loses heat by conduction. The highly porous (uncompacted) material surrounding the compacted and heated zone is a poor conductor and acts as an effective buffer to heat transport. As a result, heat losses from the hot region occur predominantly via conduction toward the surface, through the compacted zone, and the hottest region for much of the simulation is the base of the compacted zone surrounded by porous material unaffected by the impact.

Figure 2 shows the cooling rates of different grid cells in the simulation as a function of the peak temperature it experiences. Such numbers are essentially what is measured in thermochronometry and can be compared directly with metallographic cooling rates. While a range of behavior is seen, there is significant overlap with the numbers inferred from studies of chondritic meteorites [1-3]. Further, the range of cooling rates is greater than the range predicted by thermal models based purely on ^{26}Al decay, which might explain anomalous cooling rates inferred from meteorites [1]. In addition, the localized heating associated with impacts may be important in driving hydrothermal reactions in parts of an ice-bearing planetesimal, while avoiding global aqueous alteration.

Discussion: We are now exploring the range of thermal histories produced by planetesimal impacts, considering different planetesimal porosities, impact velocities, and relative sizes. Further, we are adding radiogenic heating to our thermal model to investigate under what conditions signatures of impacts are erased or preserverd due to the thermal evolution that arises from the decay of ^{26}Al. These results will be discussed.

References: [1] Kleine T. et al (2007) *EPSL, 270,* 106-118. [2] Trieloff M. (2003) *Nature, 422,* 502-506. [3] Bennett, M. E. III and McSween, H. Y. Jr. (1996) *Meteoritics & Planet. Sci., 31,* 783-792. [4] Rubin A. E. (1995) *Icarus, 113,* 156-167. [5] Wasson J. T. et al. (1987) *Meteoritics*, *22,* 525-526. [6] Cameron A. G. W. et al. *LPS XXI,* 155-156. [7] Keil K. et al. (1997) *Meteoritics & Planet. Sci., 32*, 349-363. [8] Davison T. M. et al. (2010) *Icarus, 208,* 468-481. [9] Cuzzi J. N. et al. (2008) *ApJ, 687,* 1432-1447. [10] Johansen A. et al. (2007) *Nature, 448,* 1022-1025. [11] Teiser J. and Wurm G. (2009) *MNRAS, 393, 1584-1594.* [12] Collins G. S. et al (2011) *LPS XLII* (This meeting). [13] Ghosh A., Weidenschilling S. J. and McSween H. Y., Jr. (2003) *Meteoritics & Planet. Sci*., *38*, 711-724. [14] Sumirat I. et al. (2006) *J. Porous Mater.*, *13*, 439-443.

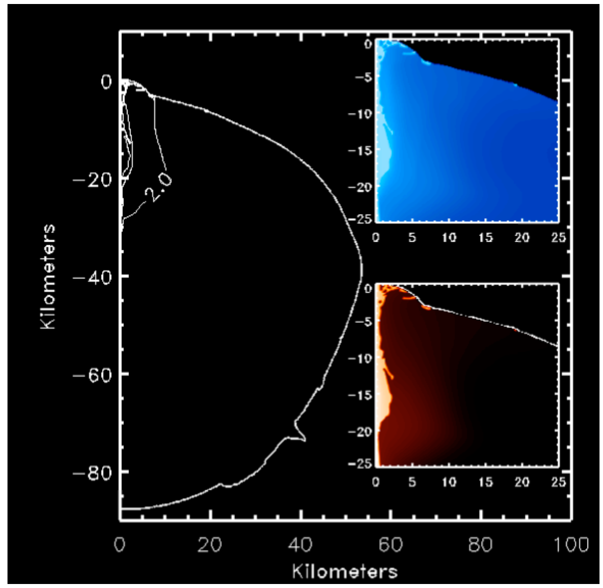

Figure 1: Post-impact structure of the planetesimal described in the text. The large image shows the shape and density structure of the planetesimal. The top inset shows the density variation around the point of impact, with dark blue representing the initial density of ~1.65 g/cm^3, increasing to a max of 3.3 g/cm^3 at the axis of symmetry. The bottom inset shows the post-impact temperature stucture of the materials, with black being the ambient temperature of 293 K, increasing to a peak T of ~1900 K near the axis of symmetry.

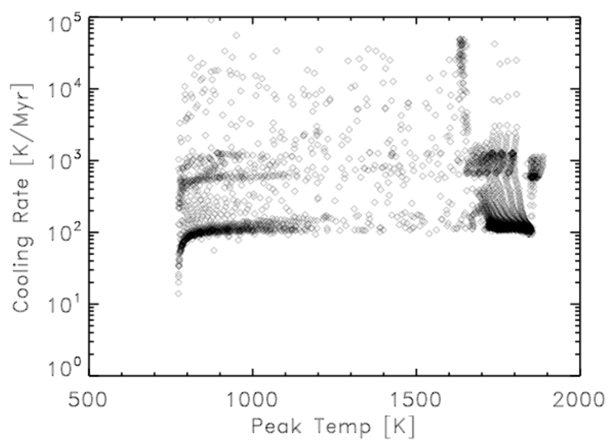

Figure 2: Cooling rates (at 773 K) plotted as a function of peak temperature for each cell in our simulation that is heated to T>773 K. (either by shock heating or thermal conduction). These cooling rates and peak temperatures overlap those inferred for thermally metamorphosed chondrites [1-3].