NUMERICAL MODELING OF BASIN-FORMING IMPACTS ON MARS: IMPLICATIONS FOR THE HEAT BUDGET OF PLANETARY INTERIOR M. Bierhaus[1], K. Wünnemann[1], D. Elbeshausen[1] and G. S. Collins[2], [1]Museum für Naturkunde Leibniz-Institut an der Humboldt-Universität zu Berlin, Invalidenstr.43, 10115 Berlin Germany, (Michael.Bierhaus@mfn-berlin.de) [2]Impacts and Astromaterials Research Centre, Department of Earth Science and Engineering, Imperial College London, London SW7 2AZ, UK

Introduction: The lack of a global magnetic field on Mars suggests that no active geodynamo exists on Mars, presently. Measurements of the magnetization of the crust, on the other hand, show widespread weak remanent magnetization indicating that a strong magnetic field was present in the past, most likely generated by convection processes in a fluid core [1]. Various reasons have been proposed why the Martian geodynamo ceased sometime early in the evolution of the planet [4]. The absence of remanent magnetization in and around the large impact basins Hellas and Argyre indicate that at the time of their formation Mars did not posses an intrinsic magnetic field [1]. It has been suggested that giant impact events forming these basins may have caused the cessation of the Martian dynamo or significantly perturbed the generation of the magnetic field [3-5]. Additionally, impacts may strongly affect mantle convection processes [5]. Basin-forming impacts cause shock waves sufficiently strong to travel through the entire planet and deposit a substantial amount of heat deep in the Martian interior. We present dynamic numerical (hydrocode) models of giant collision events to quantify the amount of heat that is deposited into a planet by the impact process. We focus on scenarios representing impacts on a Mars-like planet with impactor sizes in the range required to form the Hellas basin. We use the iSALE [6-8] code in 2D/3D for this study.

Scaling laws: In most hydrocode models peak pressures are used to work out post-shock temperatures. The temperature rise as a result of shock compression and subsequent unloading can be calculated from the peak shock pressure the target was exposed to and specific material parameters [e.g. 2,3]. Fig.1 shows the temperature increase after an impact of a 200km projectile at 15km/s: (a) for a planar surface with a conductive temperature gradient of 0.8K/km and a surface temperature of 220K (Fig.1c, green line), (b) for a 6800-km diameter sphere of uniform composition with a two layer temperature profile consiting of a conductive zone starting at 220K at the surface with a temperature gradient of 2.2K/km followed by a convecting zone where temperture increases gradually to 2300K at the centre (Fig.1c, red line). In case (a) the results are consistent with other modeling and experimental studies [e.g. 9,10], showing a zone with almost constant post-shock temperature, the so-called isothermal core, and a decrease of post-shock temperature according to a power law outside of this core (see also Fig. 2). In case (b) the effect of of the curved planetary surface is visible, leading to significantly different results. The apparent layering in Fig.1b is partially the result of the two different geotherms used for the spherical. Interferences of the shock wave with rarefaction waves propagating from the curved free-surface of the sphere also significantly affect post shock temperature distribution. Apparently, the shock wave decay and distribution of peak shock pressure cannot be estimated by using scaling laws derived for impacts on planar surfaces.

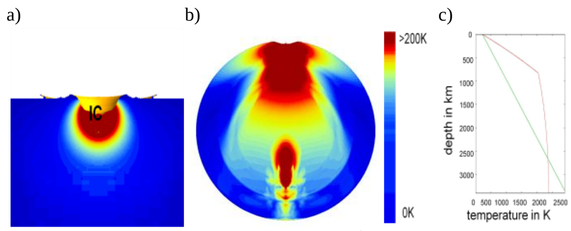

Figure 1: Two dimensional simulation of a 200 km diameter projectile impacting head-on (vertical) at 15km/s on a homogeneous (a) half space (with a planar surface), (b) uniformly composed sphere (with a curved surface). (c) starting temperature fields. green: half space; red: sphere

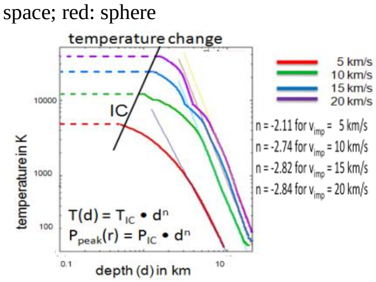

Figure 2: Impact into a planar surface; temperature vs. depth profiles for different velocities. Scaling laws [9,10] for temperature/pressure decrease outside the isobaric core (IC) and size of the IC.

Giant impacts on Mars:

Interior structure of Mars: To represent Mars in our models we assume a planet radius of 3400km, a dense iron core of 1700km radius, a 1660-km thick dunite mantle, and a 40-km thick granite crust. To calculate the thermodynamic state of the planetary interior pre-impact and during shock wave propagation we use ANEOS [11] for the different materials. The model is initialized with a self-consistent gravity, pressure, density, and temperature distribution inside the planet, implying a gradual decrease of gravity with depth from $3.7 m/s^2$ at the surface to zero at the center. We assumed a surface temperature of 220K and a core temperature of 2000K. The temperature gradient with depth was chosen to represent heat transport by conduction in the crust and convection in the mantle and core. (see [12])

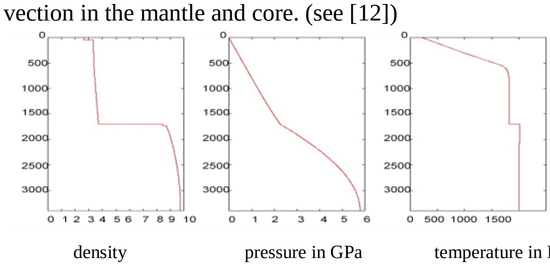

Figure 3: Setup of the assumed Mars model: pressure (a), temperature (b) and density (c) as a function of depth.

Fig.3 summarizes the initial setup of our Mars model showing pressure, density and temperature profiles from the surface to the center. Since partial or complete melting strongly affects the rheological behavior of matter we consider two different solidus curves. We assume on the one hand a solid core representing a more-or-less pure iron composition and on the other hand a liquid core, representing an iron core with some impurities [16,17].

Impactor properties: The impactor properties chosen for our models are based on model requirements and some basic assumptions. The mean impact velocity on Mars is about 10km/s [13]. For simplicity we chose the impactor material to be the same as the crust. So far we have modeled only vertical impact scenarios (head-on collisions) using a 2D cylindrically symmetric grid. Full 3D simulations taking oblique impact angles into account are currently in progress. We model impact scenarios forming crater structures similar in size to the Hellas basin. However, simulations of the entire impact process, including the formation of the final crater structure, are computationally expensive. To quantify the deposition of heat to the planetary interior it is sufficient to model only the shock wave propagation through the entire planet. Therefore, our models provide only the dimensions of the transient crater not the final crater. We use scaling laws [15] to estimate the size of the transient crater of about 1000km from the given size of the Hellas basin (2070km). In our models, projectiles 200-400 km in diameter with an impact velocity of 10 km/s produced transient craters of about 1000km. Note, the scaling laws to calculate transient crater size from final craters size [15] are based on observations of lunar complex crater structures much smaller in size than the Hellas basin. The relation between transient crater size and final crater size depends strongly on material properties (rheology) and is also affected by curvature [14] for basin structures such as Hellas.

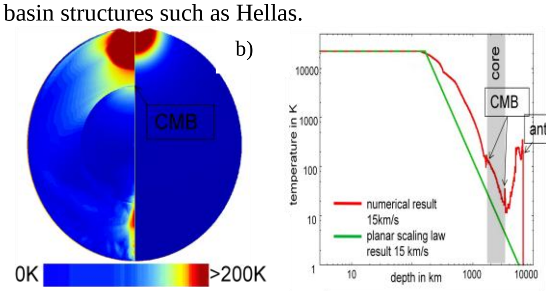

Figure 4: distribution of shock temperature after an impact of a 200km projectile at 15km/s. a) temperature difference. The left hand side shows the numerical modeling results, on the right hand side the temperature distribution was calculated with scaling laws (see Fig. 2). b) profile of temperature increase along the centerline (symmetrie axis) through the planet resulting from numerical models (red) and scaling laws (green). The shaded zone indicates the core. Note the logarithmic representation of the depth.

Impact-induced heat transfer: Fig.4 shows a comparison of the distribution of post-shock temperature increase (temperature above the initial temperature) between our 2D numerical models of a giant impact on Mars and scaling laws [3] for impacts on a planar surface (compare Fig.2). Apparently, the heating of the planetary interior is significantly underestimated by the use of simple scaling laws. In particular, a large increase in temperature can be observed antipodal to the point of impact that presumably results from superposition of shock waves traveling in different directions around the sphere.

Conclusions: Our numerical models of giant impacts on a Mars-like planet underline the differences between impacts on planar and curved sufaces. Apart from a slightly different post-shock temperature distribution in the near field, the effect of shock waves at the rear side (antipodal to the point of impact) of the planet is most striking. The models demonstrate that in previous studies of the consequences of giant impacts on the thermal budget of planetary interiors [e.g. 3,4] the deposition of heat was underestimated. In our models we observe a temperature increase of more than 100K at the core-mantle-boundary for a 200km impactor at 15km/s. This temperature increase is sufficient to change the heat flux through the core-mantle-boundary and thus may be enough to cease a geodynamo [4]. Our results feed directly into numerical models of mantle and core convection. First results show that the high temperature area flattens rapidly and spreads out along the upper boundary by convection processes displacing downwellings and plumes in its path. The flow field reorganizes after about 1200Ma, showing different convection patterns for calculations with and without antipodal heating[18]. Whether the impact-induced thermal input can sustainably pertubate geodynamic processes in the mantle and core depends on how rapidly the impact-generated temperature field is neutralized by convection processes. Further studies of the effect of impact heating on mantle convection and the geodynamo are in progress.

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