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Impact-induced melting by giant collision events – Implications for the formation of magma oceans on terrestrial planets

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Abstract

We revisit the longstanding problem how to estimate melt production during giant collision events, which may not be accurately addressed by classical scaling-laws. Therefor we carried out a series of numerical models of impact scenarios considering the initial temperature, the solidus, and the layered structure of planetary bodies.

1. Introduction

Giant collisions such as the Moon-forming event on the young Earth or basin-forming impacts are known to have influenced the chemical and thermal evolution of the terrestrial plants [4]. Besides the material that is delivered by such impacts, a significant amount of energy is transferred to the planet resulting in heating up its interior. Existing scaling-laws can predict the amount of shock melting [1,2,6] generated by impact events smaller than basin-forming events. On the scale of giant collisions such scaling laws may not be accurate as they do not account for the initial temperature and lithostatic pressure of planets interior. This can be particularly problematic for younger planets, where the initial temperatures are close to the given solidus. To better understand and quantify the mechanism of heat production and melting during large-scale impact events we conducted a series of numerical models and determined the volume of melt production.

2. Methods

To model hypervelocity collisions we use the iSALE Eulerian shock physics code [3,8] (Version *Dellen*). In iSALE the thermodynamic state (EoS) is calculated by look-up tables derived from ANEOS [eg. 5] for dunite, basalt, granite, and iron representing the Planets crust, mantle and core, respectively.

2.1 Melt calculation

To determine the distribution and volume of impact-induced melting we calculate the local increase of temperature $\Delta T(x,z)$ via the peak shock pressure method assuming that the temperature gain is directly proportional to the peak shock pressure [6]: Therefor massless Lagrangian tracers record the peak shock pressure and track their displacement during crater formation. Each tracer is associated with the mass of material in the cell, where the tracer was initially located in. The final Temperature of a tracer T_f is then determined by adding the shock-induced temperature increase $\Delta T(x,z)$ to the initial Temperature T_i: $T_f = T_i + \Delta T(x,z)$. By comparing T_f with the solidus temperature T_s=f(P_f), where f(P_f) is given by the Simon approximation [7] and P_f is the post-impact pressure at the final position of each tracer, we determine whether a “tracer” (the associated mass) is molten or not. By this approach we also account for decompression melting as a consequence of displacement of matter in the course of crater formation.

2.2 Model

For the different terrestrial planets we assume individual initial conditions regarding composition and temperature distributions, typical impact velocities, and gravity. Our planetary models include a dunitic mantle, and an iron core. For the impactor we also assume a dunitic composition. Differentiated bodies are neglected at this stage. In case of the Earth we use a setup according to the model described by Marchi et al. [2] that includes a granitic crust, while in the model of the Mars, the crust consists of basalt. We consider different thermal profiles (T_i=f(P)) representing different stages of the thermal evolution of a given planet. In all models the projectile radius is resolved by 50 cells (50 CPPR). For very large impacts we account for the curvature of the target.

3. Parameter study for impact-induced melting

In our systematic parameter study we vary impactor diameter L and velocity v_i for different temperature conditions on each planet. A reference model M_{REF} without the effects of gravity, material strength, and depth-dependent temperature, but with an earth-like layered target has been calculated, which can be directly compared with classical scaling-laws. Figure 1 shows that the reference model agrees with the scaling-laws. Depending on the impactor diameter L crust or mantle melting dominates. The more realistic model M_{Earth} (as discribed in 2.1) is approximately in agreement with classic scaling for smaller impacts; however, larger events are not well represented. It can be shown, that the different melt volumes of both models at lager impactor diameters primarily resulting from additional decompression melting in areas, where the solidus and the initial temperatures are close.

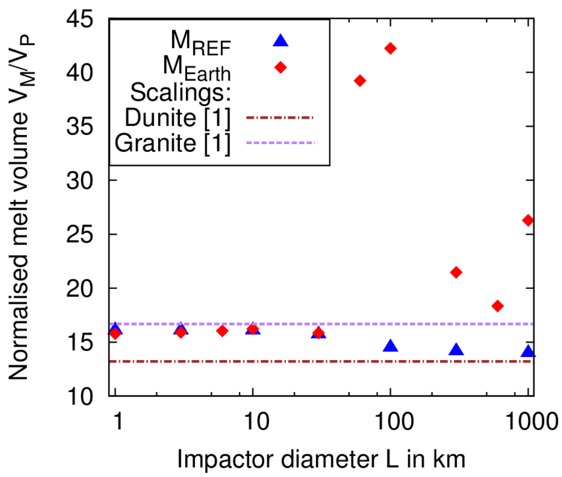

Figure 1: Normalised melt volume V_M by the projectile volume V_P as a function of impactor diameter L on an earth like planar target (red). The reference model (cf.Text) is indicated by a blue dots. The Impact velocity in both models v_i is 17 km/s. Similar behavioer can be observed for impact generated melting on Mars.

The goal is to derive a parameterization for the volume of impact-induced melting V_m as a function of impact parameters $V_m = f(L, v_i, T_i, P_f)$ for collision events arbitrary in scale.

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