MELTING DUE TO IMPACTS ON GROWING PROTO-PLANETS J. de Vries[1], F. Nimmo[2], H. J. Melosh[3], S. Jacobson[1,4], A. Morbidelli[4], and D. C. Rubie[1], [1]Bayerisches Geoinstitut, University of Bayreuth, Germany, [2] Department of Earth and Planetary Sciences, University of California, Santa Cruz, USA, [3] Department of Earth, Atmospheric, and Planetary Sciences, Purdue University, USA, [4] Observatoire de la Côte d'Azur, Nice, France.

Introduction: The terrestrial planets in our solar system were formed by a series of collisions between smaller bodies - a process that is modelled numerically using N-body accretion simulations, e.g. [1]. Planetary embryos grow by accreting smaller bodies from their so called 'feeding zone' and collide with one another when they enter each other's feeding zones. The energy involved in each collision causes large-scale melting, which allows metal-silicate segregation to occur, thus resulting in an episode of core formation. To develop a model of the compositional evolution of our rocky planets, based on a collisional history predicted by N-body accretion models [2], estimates of the amount of melting and the pressure and temperature conditions at the base of the magma ocean are required. To determine the amount of melting for each of the several hundred to several thousands of collisions, as calculated from an N-body model, a reliable and efficient shock/melting model is needed.

Depth of melting: Full three-dimensional models are computationally too time consuming, whereas two-dimensional models cannot be used for non-vertical impacts due to their assumed symmetry in the third dimension. Therefore, a parameterised model is used which describes the amount and depth of melting based on the energy needed to melt a dunite mantle (and in cases of deep melting an iron core), E_m, and the energy available from the impact [3]. The available energy depends on the impact angle and velocity (θ and v resp.) as well as on the impactor diameter, D_p and the densities of both the projectile and the target (ρ_p and ρ_t resp.).

$$V_{melt} = \frac{\pi}{6} k E_m^{-3\mu/2} \frac{\rho_p}{\rho_t} D_p^3 v^{3\mu} \sin^{2\gamma}\theta \qquad (1)$$

Taking into account the temperature increase with depth, the energy needed to melt the material at depth is corrected as follows:

$$E_m = E_m^0 \left(1 - \frac{C_p(T_s + \frac{dT}{dz} d_m)}{C_p \frac{dT_L}{dP} P(d_m) + L_m}\right), \qquad (2)$$

where C_p is the specific heat, T_s the surface temperature, d_m the depth of melting, T_l the liquidus temperature, P the pressure and L_m the latent heat of melting. Values for the material properties for both core and mantle are give in Table 1.

Parameter	Dunite	Iron
E_m^0	9.0 MJ/kg [4]	10 MJ/kg [4]
C_p	1300 J/kg K [5]	449 J/kg K [6]
L_m	718 kJ/kg [7]	247 kJ/kg [6]
T_{l0}	1950 K [8]	1811 K [9]
$\frac{dTl}{dP}$ (P<60 GPa)	28.3 K/GPa [8]	17.2 K/GPa [9]
$\frac{dTl}{dP}$ (P>60 GPa)	14.0 K/GPa [8]	17.2 K/GPa [9]

Table 1: Material properties.

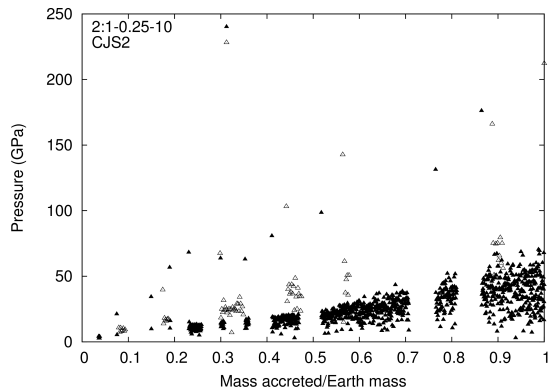

Figure 1: The pressure at the melting depth as a function of the mass of the growing planet for all collisions that finally create an Earth-like planet. Two different N-body accretion models are used, a model with a large number of relatively small bodies (2:1-0.25-10A) and a model with a smaller number of large bodies (CJS2).

Once the depth of melting is known, the pressure and liquidus temperature at that depth can be determined, and our aim is to use such conditions to model core formation and siderophile element partitioning. Ideally, these results should be independent of the specific N-body model used for the collision history. However, calculations using two different N-body accretion models with a different size distribution of the initial bodies show that this is not the case. Figure 1 shows the pressure at the melting depth for all impacts on the body that forms the final Earth-like planet as a function of the mass accreted. It compares a N-body model with a large amount of small initial bodies (2:1-0.25-10A) and a model with a smaller number of larger bodies (CJS2). A more detailed description of these two models is given in Table 2.

	CJS2	2:1-0.25-10A
# of embryo's	25	123
# of planetesimals	1023	5589
Embryo mass	$\sim \mathrm{M}_{Mars}$	$\sim \frac{1}{4}\,\mathrm{M}_{Mars}$
Planetesimal mass	$\frac{1}{40}$ M_{embryo}	$\sim 10^{-8}$ M_{Mars}

Table 2: Description N-body models.

The generally higher pressures in the CJS2 model are caused by the fact that the amount of melting depends on the size of the impactor and since the target is assumed to solidify between impacts, many small impactors will result in shallower melting depths than a small amount of larger impactors in the same time period. The initial size distribution of the bodies in the N-body models therefore influences the results. In an attempt to solve this issue, the assumption of solidification between impacts is investigated.

Magma ocean crystallisation: To investigate the assumption of solidification between impacts, a model is used to estimate the cooling time of each global magma ocean [10]. This model balances the heat flux (F) through the planet's surface with the latent heat released during crystallisation (H) and the heat released through secular cooling:

$$4\pi R^2 F = \frac{dr}{dt}\rho H 4\pi r^2 - \frac{4}{3}\pi\rho C_p \frac{d}{dt}[T(R^3 - r^3)] \quad (3)$$

The heat flux is approximated by a constant value. However, this value is strongly dependent on the amount of atmosphere present. It is likely that large impacts remove most of the atmosphere of the target, thereby increasing the surface heat flux. However, degassing of a magma ocean creates a new, possibly very dense atmosphere, decreasing the heat flux through the surface by as much as 3 orders of magnitude. Calculations have been done for both end members. For the case with no atmosphere, black body radiation gives an estimate for the heat flux and for the case with a dense atmosphere, estimates of the limit of atmospheric radiation are available. From the magma ocean crystallisation times of Lebrun et al. 2013 [11], we calculated heat fluxes of about 200,000 W/m^2 and 475 W/m^2 for these cases respectively, which fits well with the available constraints.

In general, at the start of accretion the impacts are close together. Most of these impacts are small and will cause only small amounts of melting, which will not result in a global magma ocean. However, some of these collisions are collisions between two similar-sized bodies resulting in a deep magma ocean that cannot crystallise before the next impact. Figure 2 shows results for a model with relatively large bodies. With the assumption of no insulating atmosphere, all but one of the magma oceans crystallise before the next impact. With a dense atmosphere present, a more significant number of magma oceans would still be molten when the next impact occurs.

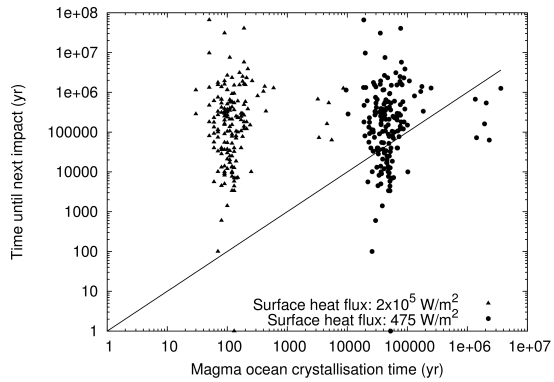

Figure 2: Crystallisation time versus time until the next impact for the CJS2 model. All points above the line represent magma oceans that crystallise before the next impact. In the case where there is no atmosphere (flux=200,000 W/m^2) only one impact happens on a molten surface. When a dense atmosphere is taken into account (flux=475 W/m^2), a large number of impacts would happen on a molten surface.

Conclusions and outlook: Further research will look into the time scales of formation of global magma oceans, compared to the crystallisation time of a melt pool. Furthermore, the differences between impacts on a magma ocean compared with a solid surface will be studied, since at least some of the impacts will occur on a molten surface.

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