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Scaling laws of impact induced shock pressure and particle velocity in planetary mantle

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ABSTRACT

While major impacting bodies during accretion of a Mars type planet have very low velocities (<10 km/s), the characteristics of the shockwave propagation and, hence, the derived scaling laws are poorly known for these low velocity impacts. Here, we use iSALE-2D hydrocode simulations to calculate shock pressure and particle velocity in a Mars type body for impact velocities ranging from 4 to 10 km/s. Large impactors of 100–400 km in diameter, comparable to those impacted on Mars and created giant impact basins, are examined. To better represent the power law distribution of shock pressure and particle velocity as functions of distance from the impact site at the surface, we propose three distinct regions in the mantle: a near field regime, which extends to 1–3 times the projectile radius into the target, where the peak shock pressure and particle velocity decay very slowly with increasing distance, a mid field region, which extends to ~4.5 times the impactor radius, where the pressure and particle velocity decay exponentially but moderately, and a more distant far field region where the pressure and particle velocity decay strongly with distance. These scaling laws are useful to determine impact heating of a growing protoplanet by numerous accreting bodies.

1. Introduction

Small planets are formed by accreting a huge number of planetesimals, a few km to a few tens of km in size, in the solar nebula (e.g. Wetherill and Stewart, 1989; Matsui, 1993; Chambers and Wetherill, 1998; Kokubo and Ida, 1995, 1996, 1998, 2000; Wetherill and Inaba, 2000; Rafikov, 2003; Chambers, 2004; Raymond et al., 2006). An accreting body may generate shock wave if the impact-induced pressure in the target exceeds the elastic Hugoniot pressure, ~3 GPa, implying that collision of a planetesimal with a growing planetary embryo can generate shock waves when the embryo's radius exceeds 150 km, assuming that impact occurs at the escape velocity of the embryo and taking the mean density of the embryo and projectile to be 3000 kg/m^3. Hundreds of thousands of collisions must have occurred during the formation of small planets such as Mercury and Mars when they were orbiting the Sun inside a dense population of planetesimal. Such was also the case during the formation of embryos that later were accreted to produce Venus and Earth. Terrestrial planets have also experienced large high velocity impacts after their formation. Over 20 giant impact basins on Mars with diameters larger than 1000 km (Frey, 2008), the Caloris basin on Mercury with a 1550 km diameter, and the South Pole Aitken basin on Moon with a 2400 km diameter are likely created during catastrophic bombardment period at around 4 Ga. The overlapping Rheasilvia and Veneneia basins on 4-Vesta are probably created by projectiles with an impact velocity of about 5 km/s within the last 1–2 Gyr (Keil et al., 1997; Schenk et al., 2012).

The shock wave produced by an impact when the embryo is undifferentiated and completely solid propagates as a spherical wave centered at the impact site until it reaches the surface of the embryo in the opposite side. Each impact increases the temperature of the embryo within a region near the impact site. Because impacts during accretion occur from different directions, the mean temperature in the upper parts of the embryo increases almost globally. On the other hand, the shock wave produced by a large impact during the heavy bombardment period must have increased the temperature in the mantle and the core of the planets directly beneath the impact site, enhancing mantle convection (e.g. Watters et al., 2009; Roberts and Arkani-Hamed, 2012, 2014), modifying the CMB heat flux which could in turn favor a hemispheric dynamo on Mars (Monteux et al., 2015), or crippling the core dynamo (e.g. Arkani-Hamed and Olson, 2010a).

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The impact-induced shock pressure inside a planet has been investigated by numerically solving the shock dynamic equations using hydrocode simulations (e.g. Pierazzo et al., 1997; Wünnemann and Ivanov, 2003; Wünnemann et al., 2006; Barr and Citron, 2011; Kraus et al., 2011; Ivanov et al., 2010; Bierhaus et al., 2012) or finite difference techniques (e.g. Ahrens and O'Keefe, 1987; Mitani, 2003). However, these numerical solutions demand considerable computer capacity and time and are not practical for investigating the huge number of impacts that occur during the growth of a planet. Hence, the scaling laws derived from field experiments (e.g. Perret and Bass, 1975; Melosh, 1989) or especially from hydrocode simulations (Pierazzo et al., 1997) are of great interest when considering the full accretionary history of a planetary objects (e.g. Senshu et al., 2002; Monteux et al., 2014) or when measuring the influence of a single large impact on the long-term thermal evolution of deep planetary interiors (e.g. Monteux et al., 2007, 2009, 2013; Ricard et al., 2009; Roberts et al., 2009; Arkani-Hamed and Olson, 2010a; Arkani-Hamed and Ghods, 2011). Although the scaling laws provide approximate estimates of the shock pressure distribution, their simplicity and the small differences between their results and those obtained by the hydrocode simulations of the shock dynamic equations (that are likely within the numerical errors that could have been introduced due to the uncertainty of the physical parameters used in the hydrocode models) make them a powerful tool that can be combined with other geophysical approaches such as dynamo models (e.g. Monteux et al., 2015) or convection models (e.g. Watters et al., 2009; Roberts and Arkani-Hamed, 2012, 2014).

During the decompression of shocked material much of the internal energy of the shock state is converted into heat leading to a temperature increase below the impact site. The present study focuses on deriving scaling laws of shock pressure and particle velocity distributions in silicate mantle of a planet on the basis of several hydrocode simulations. The scaling laws of Pierazzo et al. (1997) were derived using impact velocities of 10–80 km/s, hence may not be viable at low impact velocities. For example, at an impact velocity of 5 km/s, comparable to the escape velocity of Mars, the shock pressure scaling law provides an unrealistic shock pressure that increases with depth. Here we model shock pressure and particle velocity distributions in the mantle using hydrocode simulations for impact velocities of 4–10 km/s and projectile diameters ranging from 100 to 400 km, as an attempt to extend Pierazzo et al.'s (1997) scaling laws to low impact velocities and reasonable impactor radii occurring during the formation of terrestrial planets. Hence, on the basis of our scaling laws it is possible to estimate the temperature increase as a function of depth below the impact site for impact velocities compatible with the accretionary conditions of terrestrial protoplanets. These scaling laws can easily be implemented in a multi-impact approach (e.g. Senshu et al., 2002; Monteux et al., 2014) to monitor the temperature evolution inside a growing protoplanet whereas it is not yet possible to adopt hydrocode simulations for that purpose.

The hydrocode models we have calculated are described in the first section, while the second section presents the scaling laws derived from the hydrocode models. The concluding remarks are relegated to the third section.

2. Hydrocode models of shock pressure distribution

The huge number of impacts during accretion makes it impractical to consider oblique impacts. Not only it requires formidable computer time, but more importantly because of the lack of information about the impact direction, i.e. the impact angle relative to vertical and azimuth relative to north. Therefore, we consider only head-on collisions (vertical impact) to model the thermo-mechanical evolution during an impact between a differentiated Mars size body and a large impactor. We use the iSALE-2D axisymmetric hydrocode, which is a multi-rheology, multi-material hydrocode, specifically developed to model impact crater formation on a planetary scale (Collins et al., 2004; Davison et al., 2010). To limit computation time, we use a 2 km spatial resolution (i.e. more than 25 cells per projectile radius, *cppr*) and a maximum time step of 0.05 s which is sufficient to describe the shockwave propagation through the entire mantle. The minimum post impact monitoring time is set to the time needed by the shockwave to reach the core–mantle boundary ($\approx$5 min for the impact velocities studied here).

We investigate the shock pressure and particle velocity distributions inside a Mars size model planet for impact velocities $V_{\rm imp}$ of 4–10 km/s and impactors of 100–400 km in diameter. Such impactors are capable of creating basins of 1000–2500 km in diameter according to Schmidt and Housen (1987) and Holsapple (1993) scaling relationships between the impactor diameter and the resulting basin diameter. These basins are comparable with the giant impact basins of Mars created during the heavy bombardment period at around 4 Ga (Frey, 2008).

In our models, the impactor was simplified to a spherical body of radius $R_{\rm imp}$ with uniform composition while the target was simplified to a two layers spherical body of radius R and an iron core radius of $R_{\rm core}$. The silicate mantle has a thickness of δ_m (see Table 1). We adopt physical properties of silicates (dunite or peridotite) for both the mantle and the impactor to monitor the shock pressure and the particle velocity in a Mars type body. We approximate the thermodynamic response of both the iron and silicate material using the ANEOS equation of state (Thompson and Lauson, 1972; Benz et al., 1989). To make our models as simple as possible we do not consider here the effects of porosity, thermal softening or low density weakening. However, as a first step towards more realistic models, we investigate the influence of acoustic fluidization and damage. All these effects can be accounted for in iSALE-2D and we will consider each effect in a separate study in near future.

Table 1
Typical parameter values for numerical hydrocode models.

Target radius	R	3400 km
Target core radius	$R_{\rm core}$	1700 km
Silicate mantle thickness	δ_m	1700 km
Impactor radius	$R_{\rm imp}$	50–200 km
Impact velocity	$V_{\rm imp}$	4–10 km/s
Mantle properties (silicates)		
Initial density	ρ_m	3314 kg/m^3
Equation of state type		ANEOS
Poisson		0.25
Strength Model		Rock
(iSALE parameters)		(Y_{i0} = 10 MPa, μ_i = 1.2, Y_{im} = 3.5 GPa)
Acoustic Fluidization Model		Block
(iSALE parameters)		($t_{\rm off}$ = 16 s, $c_{\rm vib}$ = 0.1 m/s, vib$_{\rm max}$ = 200 m/s)
Damage Model		Ivanov
(iSALE parameters)		($\varepsilon_{\rm fb}$ = 10^{-4}, B = 10^{-11}, p_c = 3×10^8 Pa)
Thermal softening and porosity models		None
Core properties (iron)		
Initial density	ρ_c	7840 kg/m^3
Equation of state type		ANEOS
Poisson		0.3
Strength Model		Von Mises
(iSALE parameters)		(Y_0 = 100 MPa)
Acoustic Fluidization Model		Block
(iSALE parameters)		($t_{\rm off}$ = 16 s, $c_{\rm vib}$ = 0.1 m/s, vib$_{\rm max}$ = 200 m/s)
Damage, thermal softening and porosity models		None