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# Impacts onto $H_2O$ ice: Scaling laws for melting, vaporization, excavation, and final crater size

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ABSTRACT

Shock-induced melting and vaporization of $H_2O$ ice during planetary impact events are widespread phenomena. Here, we investigate the mass of shock-produced liquid water remaining within impact craters for the wide range of impact conditions and target properties encountered in the Solar System. Using the CTH shock physics code and the new 5-phase model equation of state for $H_2O$, we calculate the shock pressure field generated by an impact and fit scaling laws for melting and vaporization as a function of projectile mass, impact velocity, impact angle, initial temperature, and porosity. Melt production nearly scales with impact energy, and natural variations in impact parameters result in only a factor of two change in the predicted mass of melt. A fit to the $\pi$-scaling law for the transient cavity and transient-to-final crater diameter scaling are determined from recent simulations of the entire cratering process in ice. Combining melt production with $\pi$-scaling and the modified Maxwell Z-model for excavation, less than half of the melt is ejected during formation of the transient crater. For impact energies less than about $2 \times 10^{20}$ J and impact velocities less than about 5 km s$^{-1}$, the remaining melt lines the final crater floor. However, for larger impact energies and higher impact velocities, the phenomenon of discontinuous excavation in $H_2O$ ice concentrates the impact melt into a small plug in the center of the crater floor.


## 1. Introduction

Impact craters are the most common geologic feature on planetary surfaces. Bolides impacting at typical velocities of a few to several 10's of km s$^{-1}$ (Zahnle et al., 2003) achieve shock pressures capable of melting and vaporizing $H_2O$ ice (Stewart et al., 2008). In some cases, impact-generated crater lakes may persist for geologically interesting timescales (Thompson and Sagan, 1992; Artemieva and Lunine, 2005; O'Brien et al., 2005). Giant impact events, which dominated the late stages of planetary accretion, may generate oceans of melt (Tonks and Melosh, 1993). Finally, an intense period of impact events could also lead to differentiation of a planet or satellite (e.g., Tonks and Melosh, 1992; Tonks, W.B., Pierazzo, E., Melosh, H.J., unpublished manuscript, 1993; Monteux et al., 2009; Barr and Canup, 2010).

Impact events have also shaped planetary atmospheres. The production of vapor during accretionary impacts contributes shock-released volatiles to the growth of terrestrial atmospheres (e.g., Benlow and Meadows, 1977; Lange and Ahrens, 1982). However, when the mass of vaporized material becomes sufficiently large, rapid expansion may lead to partial loss of a pre-existing atmosphere (e.g., Melosh and Vickery, 1989; Ahrens, 1993; Shuvalov, 2009).

At present, it is difficult to predict the mass of melted and vaporized material associated with a particular size impact crater on ice-rich surfaces. Laboratory scale experiments do not produce significant amounts of melt, and unlike cratering on rocky planets, ground truth data does not exist for planetary scale craters in ice. Observations of crater melt sheets on the icy bodies of the outer Solar System are limited (e.g., Schenk and Turtle, 2009) and complicated by the negative buoyancy of liquid water over ice.

Recent developments in shock physics models of the equation of state (EOS) and rheology of $H_2O$ ice have led to simulations of full crater formation that reproduce much of the diversity of morphological features observed on icy bodies (Senft and Stewart, 2008, 2011). In particular, the inclusion of high-pressure solid polymorphs in the EOS leads to an unusual phenomenon during crater formation in ice called discontinuous excavation (Senft and Stewart, 2011). Discontinuous excavation causes a concentration of impact melt in a small plug in the crater floor of similar dimensions to observed central pit features on the largest icy satellites.

In this work, we present calculations of shock-induced melting and vaporization of $H_2O$ ice for the range of impact and target conditions found in the Solar System. Our parameter space spans low velocity accretionary impacts to the highest velocity cometary

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impacts and considers the relatively warm solid ice of Earth to cold porous ice of Kuiper Belt objects. This work considers more impact and target parameters for impact induced melting and vaporization than included in previous studies. We also determine the fraction of shock-induced melt ejected from the crater using the modified Maxwell Z model (Maxwell, 1977; Croft, 1980). Then using crater size scaling relationships, the amount of melt produced during an impact event is related to the final crater size.

Section 2 provides a description of the numerical model, the equation of state, and the entropy method used to calculate the mass of material that was melted or vaporized. In Section 3, the effects of impact angle, temperature, impactor composition, and porosity are presented, and two scaling equations are derived to predict the amount of melt and vapor produced for any impact condition. Section 4 discusses aspects of the calculations in more detail, compares the results to previous work, and addresses more complicated scenarios including icy mixtures. To estimate the mass of melt and vapor for a given size crater, Section 5 presents scaling relationships for transient craters in ice, the mass of ejected melt during transient crater formation, and crater collapse.

### 1.1. The shock pressure field from an impact event

When an impactor strikes a target, a strong shock wave propagates into the target and back into the impactor. The highest shock pressures are generated below the impact point in a nearly isobaric volume of target material called the isobaric core. The hemispherical shock front propagates into the target, decaying in amplitude with distance and depositing internal and kinetic energy. Above material-dependent critical shock pressures, the deposited internal energy is sufficient to melt or vaporize the target upon release to ambient pressure.

Croft (1982) developed a semi-analytic model for shock wave decay and found that the shock pressure $P$ at a radial distance $R$ is given by

$$P(R) = P_0\left(\frac{R_{ic}}{R}\right)^n, \qquad (1)$$

where $P_0$ is the pressure at the radius $R_{ic}$ of the isobaric core. The decay exponent $n$ depends on both the material and impact velocity (Ahrens and O'Keefe, 1977; Pierazzo et al., 1997), and ranges from 1 to 3 for a wide range of nonporous materials.

The residual kinetic energy of the target after passage of the shock wave creates a divergent flow of material centered on the isobaric core, which leads to formation of the transient crater cavity (Maxwell, 1977; Croft, 1980). The excavated volume overlaps with the volume of most highly shocked material; hence, a significant fraction of melt is ejected from the crater. The melt that is not ejected lines the walls and floor of the transient crater. The shape of the transient crater is gravitationally unstable and must collapse to some degree. Smaller, simple craters undergo relatively little collapse. The walls of the transient crater slump, and a melt-rich breccia lines the bottom of the final bowl-shaped crater. The greater gravitational instability in larger transient craters leads to a more significant collapse and the formation of complex crater morphologies. The spatial distribution of the melt sheet in complex craters depends on the final morphology (e.g., a central peak or peak-ring crater). For a more complete description of the dynamics of the impact cratering process see Melosh (1989).

### 1.2. Previous work

Previous researchers have estimated the total amount of melt generated in an impact event by calculating the shock pressure decay field. We acknowledge the semi-analytic estimates for vaporization of water and melting of ice by Croft (1982, 1983). However, such semi-analytic approaches have been superceded by the use of a shock physics code coupled with an EOS model to calculate the pressure decay field in the target with much higher accuracy (e.g., Ahrens and O'Keefe, 1977; Pierazzo et al., 1997; Pierazzo and Melosh, 2000; Artemieva and Lunine, 2003, 2005). The mass of melt plus vapor is derived by calculating the volume of material that exceeded the critical shock pressure for melting. In particular, the comprehensive study by Pierazzo et al. (1997) calculated the volume of melt plus vapor produced for vertical impacts onto a wide range of materials. They found that most materials follow a single scaling law:

$$\log\left(\frac{V_m}{V_i}\right) = a + \frac{3}{2}\mu\log\left(\frac{U^2}{E_M}\right). \qquad (2)$$

The form of Eq. (2) comes from an analytical analysis of melt volumes by Bjorkman and Holsapple (1987), where $V_M$ is the volume of melt, $V_i$ is the impactor volume, $U$ is the impactor velocity, and $E_M$ is the specific internal energy required to reach complete melting upon decompression from the shocked state on each material's principal Hugoniot. The nondimensional $U^2/E_M$ is called the melt number. The values for the intercept, $a$, and the velocity exponent of the coupling parameter of Holsapple and Schmidt (1982), $\mu$, were fitted to the combined simulation results for dunite, granite, aluminum, and iron to obtain $a = -0.80 \pm 0.14$ and $\mu = 0.709 \pm 0.041$ for $U^2/E_M > 30$ (Pierazzo et al., 1997).

In Pierazzo et al. (1997), only $H_2O$ ice deviated from Eq. (2), with about an order of magnitude greater melt volume. Artemieva and Lunine (2003, 2005) also calculated the volume of melt produced for impacts onto Titan using a revised water EOS (Turtle and Pierazzo, 2001) and found about an order of magnitude less melt than Pierazzo et al. (1997). However, a direct comparison of the two studies is imperfect as Pierazzo et al. (1997) considered vertical impacts onto nonporous ice and Artemieva and Lunine (2003, 2005) investigated impacts onto much colder ice with an 800-m porous surface layer. Barr and Citron (2011) studied impact melt production in a variety of materials including nonporous $H_2O$ ice; they also found about an order of magnitude less melt compared to Pierazzo et al. (1997).

## 2. Methods

As discussed briefly in the introduction, calculating the amount of melt and vapor produced from an impact event is qualitatively very simple: (i) determine the maximum shock pressure achieved as a function of initial position within the target and (ii) compare the pressure field in the target to the critical shock pressures for melting and vaporization.

### 2.1. Numerical model

The shock pressure field was calculated with the widely used Eulerian finite-volume shock physics code CTH (McGlaun et al., 1990). We considered impact velocities from 1 to 80 km s$^{-1}$, initial temperatures from 50 to 300 K, porosities of 0–60%, impactor compositions of a porous comet to a rocky asteroid, and impact angles from vertical to 30° from the horizontal. To keep the melting and vaporization results general, gravity and thermal gradients were not included. The projectile diameter was 1 km in all simulations, and as it is the only length scale in the calculation, the results can be scaled to other projectile sizes. The simulations were run until the shock wave decayed to pressures well below that required for melting (typically several seconds for the 1 km diameter projectile considered in this study). A separate set of full crater formation simulations was used to determine scaling laws for the