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Ejecta range: A simulation study of terrestrial impacts

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Abstract

In this work the meteorite and target mass partition into high-speed ejecta during the formation of terrestrial impact craters is investigated. Multi-material hydrocode calculations are carried out through the entire excavation phase, and the mass of each material moving upwards with velocities inside a range of intervals is obtained. Impact of a 10 km diameter stony asteroid with 20 km s^{-1} into the continental crust is compared for the cases of a single layer of granite, taken to be representative for the crust, and of a two-layer crust with a 3 km thick sedimentary cover of limestone on top of granite basement, more appropriate for the Chicxulub crater. The proportion of meteorite and crustal material in high-speed ejecta is found as a function of velocity and time, and maximum distances to the crater can be estimated. The resulting distal ($>$7000 km) ejecta mass for vertical impact is less than a percent of the impactor mass, assuming ballistic transport. Simulations of oceanic impact of a 1 km-sized stony asteroid into 5.5 km deep sea are also presented. Here, ejection of meteorite material initially is delayed, but finally it leaves the ocean in a cloud of steam and water. The velocities of meteorite material are much lower compared with the continental impact, insufficient to reach large distances on ballistic trajectories.

Keywords: Impact cratering; Hydrocode; Modeling; Terrestrial craters

1. Introduction

Presently, more than 70 terrestrial impact craters with diameters larger than 10 km are known (Earth Impact Database, 2006, http://www.unb.ca/passc/ImpactDatabase). While small craters possess localized ejecta blankets, highly obliterated in most cases, larger ones are probably associated with global ejecta layers containing tektite, magnesioferrite spinel, shocked quartz and other relics of a high-pressure or condensate phase (Melosh, 1989). Extra-terrestrial material is mostly dispersed in the form of condensate or melt droplets. Sometimes it can be identified due to an unusual high abundance of siderophile and platinum-group elements (PGE) (Alvarez et al., 1980) or due to isotopic ratios of certain elements uncommon in the crust and the Earth mantle (in particular the ^{187}Os/^{188}Os and ^{187}Re/^{188}Os ratios, see Koeberl and Shirey, 1997). Its clear distinction from material of the Earth mantle is still difficult (Montanari and Koeberl, 2000). Chondrite and iron meteorite impactors may create an undisputed indication, whereas other types of meteorites as well as comets cannot so easily be identified after a few millions of years. CI, CV, or CO chondrites carry 450–770 ng g^{-1} of iridium and iron meteorites up to about 16 000 ng g^{-1} (Kring et al., 1996) what is a large concentration in comparison to the average in the Earth's continental crust of 22 pg g^{-1} (Peucker-Ehrenbrink and Jahn, 2001). In most parts of the K–T boundary layer the PGE abundance exceeds the crustal one by more than a factor 50–100 (Alvarez et al., 1980; Claeys et al., 2002). However, some K–T boundary sites show an extraordinary high abundance corresponding to up to 10% of meteoritic material admixture (Evans et al., 1994). At early times of crater formation, when ejecta velocities are large enough to make long-range ballistic transport possible, one would expect that the meteorite mass fraction within these ejecta is high. Later in the process of transient crater growth an increasing mass of target material is launched, and the mass ratio should decrease. Thus the meteorite-to-target

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mass ratio in ejecta should increase with the distance to the impact site, as it is actually observed (Claeys et al., 2002; Tagle et al., 2004). Measured PGE and siderophile element enhancement over crustal values displays variations (Claeys et al., 2002), which have not been completely understood.

There is a growing number of examples of numerical impact crater simulations in planetary science using hydrocodes. Multi-material calculations make it possible to distinguish between material of the meteorite and target contained in any computational cell in an unambigous way. A discussion of such numerical methods is given by Benson (1992), see also de Niem et al. (2007). Previous simulations of the Chicxulub impact by Pierazzo et al. (1997, 1998) did not distinguish the material of tracer particles when displaying their velocity. Furthermore, their results obtained with mass-less tracer particles do not make it possible to give numbers for the actual ejected mass. Here we only remark that the use of tracers is not without problems. To resolve the high-velocity tail of ejecta of impactor origin, representing as few as a permil by mass, will require thousands of tracer particles for the meteorite, at least. About three orders of magnitude more target material is ejected, and to maintain the same accuracy there have to be millions of tracers then. Cylindrical coordinates add another difficulty: mass in a computational cell is growing radially outward, and the number of tracers has to grow in proportion to mass. This is counter-productive because the innermost part has to be resolved properly. Nevertheless most authors except Stoeffler et al. (2002) have used a rather small number of tracer particles in related work.

Simulations of initial phases of the Chicxulub crater formation by Pierazzo et al. (1997, 1998) showed a relatively long-time containment effect of the evaporated impactor, such that the expansion of the vapor cloud took appreciably longer than expected for a hemispherical free expansion of highly compressed gas into vacuum (see Zel'dovich and Raizer, 1967; Melosh, 1989). This was attributed to peculiarities of their equation of state ANEOS (analytical equation of state, Thompson and Lauson, 1972). Subsequent investigations of the Chicxulub impact (Pierazzo and Melosh, 2000) using CTH (chart-D to three halves, improvement and 3D extension of Chart-D, an earlier 2D hydrocode) cover only the initial high-pressure phase (5 s). Although the highest resolutions for the meteorite have been obtained, so far, the simulation time is not long enough to reach the end of the excavation phase. Even earlier simulations with CSQ (chart-D squared) by Pierazzo et al. (1997, 1998) covering about 33 s did not yet reach the end of the excavation phase either. Furthermore, they used the so-called simple line interface construction (SLIC) technique (Noh and Woodward, 1976), a less accurate representation of material boundaries than the piecewise-linear technique implemented in CTH (McGlaun et al., 1990; Benson, 2002) and our hydrocode (de Niem et al., 2007). Applications of an Eulerian multi-material extension of SALE (simplified arbitrary Lagrangean–Eulerian, Hirt et al., 1997; see references in Wünnemann and Lange, 2002; Ivanov, 2005 for multi-material extensions of SALE) to the Chicxulub event do not show any detail and resolution for the impactor. They are tailored to test models of elastic–plastic flow with a transient weakening of shear strength during the crater collapse and implement a continuum damage model (Wünnemann and Ivanov, 2003; Collins et al., 2004).

The consequences of a terrestrial impact depend on the impact site. There are strong differences between impacts into continental crust and into the deep ocean (Shuvalov and Trubetskaya, 2001; Artem'eva and Shuvalov, 2002) resulting in compositional dissimilarity of ejecta material sampled from a "fireball layer" for both scenarios. If no crater is created on the ocean floor or most of deceleration of the impactor is due to water, genuine meteorite material is ejected, only subsequent diagenetic alteration can diminish the meteorite component. This assumes that mixing with sediments on the ocean floor is less important, and the dynamics of ejection of this material is fast enough, as is obtained here. The effect tends to enhance siderophile element abundances for any oceanic impact occurring in the deep sea (Kyte and Brownlee, 1985; Kyte, 2002a, b). Furthermore, peak pressures can be lower for oceanic impacts at comparable velocity and, therefore, evaporation of the meteorite is less probable in deep sea impact (Shuvalov, 2003). Crawford and Mader (1998) simulated oceanic impact of 250 m-, 500 m- and 1 km-sized stony asteroids into 5 km deep ocean using CTH and subsequently investigated tsunami propagation with an incompressible Navier–Stokes solver. Shuvalov and Trubetskaya (2001) considered three different marine impacts: Mjølnir, Lockne, and Eltanin using their SOVA (solid air vapor) hydrocode in 2D geometry. Without taking special care of the impactor they characterized different regimes of flow at vertical impact to better understand crater morphology in the oceanic floor, the collapse of the overlying water cavity and the tsunami generation. Spatial resolution, boundary conditions and other details are neither given by Shuvalov and Trubetskaya (2001) nor by Artem'eva and Shuvalov (2002), where additionally oblique impacts were simulated for 1 s in 3D. Wünnemann and Lange (2002), with a two-material extension of SALE, omitting the atmosphere, performed similar 2D and very long-lasting simulations of Eltanin, using reflective boundary conditions except at the symmetry axis, with the purpose to study interaction of currents with material displaced during crater formation. All authors except Crawford and Mader (1998) used an identical Tillotson EOS for water. Gisler et al. (2003) simulated 3D oceanic impacts of several hundred meters to 1.5 km-sized pure iron ($7.81\,g\,cm^{-3}$) as well as dunite ($3.32\,g\,cm^{-3}$) spheres using a tabulated EOS for water to estimate Tsunami amplitudes. As in the papers by previous authors their domain size was considerably smaller than the range of a compressible wave during the whole