Modeling the formation of the K-Pg boundary layer.

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ABSTRACT

In this paper we investigate the formation of the Cretaceous-Paleogene (K-Pg) boundary layer through numerical modeling. The K-Pg layer is widely agreed to be composed of meteoritic material and target rock from the Chicxulub impact site, that has been ejected around the globe and mixed with local material during final deposition. The observed composition and thickness of the K-Pg boundary layer changes with azimuth and distance from the impact site. We have run a suite of numerical simulations to investigate whether we can replicate the observational data, with a focus on the distal K-Pg layer and the impact glasses at proximal sites such as Beloc, Haiti. Previous models of the K-Pg ejecta have assumed an initial velocity distribution and tracked the ejecta to its final destination. Here, we attempt to model the entire process, from impact to the arrival of the ejecta around the globe. Our models replicate the observed ejecta thickness at proximal sites, and the modeled ejecta is composed of sediments and silicate basement rocks, in agreement with observational data. Models that use a 45° impact angle are able to replicate the total ejecta and iridium volume at distal sites, and the majority of the ejecta is composed of meteorite and target sediments. Sub-vertical impacts generate too little iridium, and oblique impacts of ≤ 30 degrees generate too much. However, in contrast to observations, models that involve ballistic transport of ejecta lead to ejecta thickness decreasing with increasing distance, and are unable to transport shocked minerals (quartz and zircon) from the Chicxulub basement rocks around the globe. We suggest that much of the K-Pg ejecta is transported non-ballistically, and that the most plausible mechanism is through re-distribution from a hot, expanding atmosphere. The results are important for future investigations of the environmental effects of the Chicxulub impact.

Key words: Chicxulub, K-Pg boundary, impact, crater, shocked quartz

1. Introduction

The K-Pg boundary is widely recognized as a global ejecta layer formed by a large meteorite impact 65 million years ago. The discovery of an iridium anomaly (Alvarez et al., 1980) and shocked quartz grains (Bohor et al., 1984) within the K-Pg boundary layer provided the strongest confirmation of the impact hypothesis. The Chicxulub crater, historically referred to as the Cretaceous-Tertiary (K-T) impact crater or ‘smoking gun’ (Hildebrand and Boyton, 1990; Hildebrand et al., 1991; Swisher et al., 1992), was discovered following a decade of studies of these worldwide ejecta deposits. The thickness and composition of the K-Pg layer varies with distance from the Chicxulub structure (see Smit, 1999 for a review). Distal sites (> 4000 km from Chicxulub) are characterized by an ejecta layer of no more than 2-4 mm thick that is enriched in PGEs (platinum-group elements) and contains abundant impact spherules – microkrystites (Alvarez et al., 1980; Hildebrand 1993; Smit et al., 1992; Smit, 1999). The ejecta layer at sites in North America (2000-4000 km from Chicxulub) is 0.5-2.0 cm thick, and has a dual-layer stratigraphy. The upper layer (historically called the fireball layer) is compositionally comparable to the distal ejecta layer, and the lower layer is comparable to spherule beds at sites closer to Chicxulub. The K-Pg layer at sites < 2500 km from Chicxulub is quite variable, and appears to depend on whether the site was in the Gulf of Mexico, Caribbean or Atlantic, and whether the site was in shallow or deep water, or was continental (Smit, 1999). Although we have good constraints on the composition and thickness of the K-Pg ejecta, we do not have a clear understanding of the mechanism for the transport of these ejecta from the impact site to their final location.

The ballistic distribution of ejecta is well studied around craters on the Moon, but the Moon is not an appropriate analogue for ejection processes on Earth as it has no atmosphere. We also cannot use experimental data, as plume formation and its expansion in the Earth’s atmosphere cannot be reproduced in the laboratory or in small-scale field experiments. The only available

tools for modeling ejection processes, and post-impact transport through an atmosphere, are 3D numerical modeling codes. There have been a number of previous attempts to model the ejection of material from Chicxulub, and these have either assumed an initial velocity distribution of ejecta as it leaves the impact site (e.g. Alvarez et al., 1995; Kring and Durda, 2002; Croskell et al., 2002; Goldin and Melosh, 2007) or have modeled the early stages of impact only (O’Keefe et al., 2001). Here we attempt to model the entire process, from impact to arrival of the ejecta around the globe. The focus of this paper is to determine the total volume and composition of materials ejected around the globe using 3D numerical modeling codes, and assess whether results from these models adequately match observational data. We have modeled ejection processes for all materials involved and then used ballistic continuation to reproduce the deposition of: 1) impact glasses at distances of 800-2000 km from the impact site and 2) the global distal ejecta. At this stage, we do not attempt to model the dual layer in North America, or any non-ballistic mechanisms of ejecta re-distribution, such as within atmospheric flows, ground surges, or tsunamis. However, in the discussion, we consider possible additional mechanisms of ejecta re-distribution.

There are good constraints on the size of this impact (Morgan et al., 1997; Hildebrand et al., 1998) and lithology of the target rocks (e.g. Lopez Ramos, 1975; Sharpton et al., 1992; Kring, 1995), but estimates of its environmental consequences and causes for the mass extinction are still debated (Ivanov et al., 1996; Pierazzo et al., 1998; Pope, 2002; Kring, 2007). We have determined the total mass and energy of global ejecta as well as its velocity, physical state and precursor target rocks, as these are important input parameters for modeling the post-impact climate. Our results are important for estimating the long-term consequences of the K-Pg impact and assessing the causes of the K-Pg mass extinction.

2. Observational data

Alvarez et al. (1980) identified anomalously high iridium levels within the K-Pg layer and first suggested the catastrophic impact hypothesis. They estimated a projectile size of between 6.6 and 14 km diameter, based upon the Ir-abundance at sites in Gubbio, Italy and Stevn's Klint, Denmark, which varied by an order of magnitude. Numerous studies of K-Pg samples during the subsequent 20 years have shown that: (1) Ir-concentration does not correlate with distance from Chicxulub; (2) Peak Ir-concentration changes substantially even within the samples from the same site (Smit, 1999; Claeys et al., 2002). The average Ir fluence in the K-Pg fireball layer is $(40 - 55) \times 10^{-9} \text{ g/cm}^2$ (Orth et al., 1987; Hildebrand et al., 1998; Kyte, 2004), and the total mass of Ir worldwide is $(2.0 - 2.8) \times 10^8 \text{ kg}$. An undifferentiated impactor was indicated by the near chondritic distribution of the siderophile trace elements in the K-Pg layer (Kyte et al., 1985), while the chromium isotopic composition pointed to carbonaceous chondrite (Shukolyukov and Lugmair, 1998). Using an average Ir mass fractional abundance for stony meteorites of about 500 ng/g (Ehmann et al., 1970) and an asteroid mass fraction globally dispersed after the impact of between 0.22 (Alvarez et al., 1980) and 0.5 (Vickery and Melosh, 1990), the K-Pg projectile mass should be $0.8 \times 10^{15} - 2.5 \times 10^{15} \text{ kg}$. These values suggest a projectile diameter of about 8-12 km (assuming asteroid density of 2600 kg/m^3), but even the upper estimate is too small to produce the 180-200 km diameter Chicxulub crater (Schmidt and Housen, 1987). If the K-Pg projectile was a carbonaceous chondrite then the Ir-abundance could be as high as 799 ng/g (Kallemeyn and Wasson, 1981), and the impactor would be even smaller.

The fact that the total mass of iridium in the K-Pg layer is less than expected, given the size of the Chicxulub crater, suggests that some aspect of these previous calculations is incorrect. Either the impactor had a lower than average iridium concentration, or the volume of iridium that remained close to the impact site or escaped the Earth is greater than suggested. For example, a

high-velocity cometary impact, as suggested by Vickery and Melosh (1990) and Hildebrand et al. (1998), would produce a larger crater for the same projectile size and a lower Ir-abundance.

2.1 Distal K-Pg boundary layer

The distal K-Pg layer has a fairly uniform thickness of 2-3 mm, and contains spherules and shocked minerals in a clay matrix. Impact spherules are the largest component, and constitute more than half of the global K-Pg boundary clay layer: 850 km³ (Smit, 1999) in a total volume of between 1000 and 1500 km³. Smit and Klaver (1981) linked microscopic mineral spherules from the K-Pg boundary clay in Spain to the distal fallout of an extraterrestrial impact cloud, even though they were almost entirely replaced by K-feldspar. The origin of the clay that surrounds the spherules is unclear, and may be either locally derived or part of the fallout. The total Ir deposited is about 55 ng per cm² in K-Pg sediments (Kyte, 2004). If the K-Pg layer is formed entirely from ejecta, this implies a ratio of ~1:5 between meteoritic material with an iridium concentration of 500 ng/g and target rocks.

Impact spherules have subsequently been found in every K-Pg boundary clay fingerprinted by the Ir anomaly, and have been interpreted as diagenetically altered quenched melt droplets from an impact cloud (Montanari et al., 1983; Montanari, 1991; Ebel and Grossman, 2005). It appears that different groups of microkrystites coexist in individual ejecta layers. The most notable difference is that some microkrystites contain skeletal Ni-rich magnesioferrite spinels (Kyte and Bostwick, 1995; Montanari et al., 1983). The spinel-bearing microkrystite group is more mafic in composition and significantly more enriched in Ir (Montanari et al., 1983) when compared with the glauconite and K-feldspar rich microkrystites. Ebel and Grossman (2005) used thermodynamic calculations to determine the composition of spinels condensed from the K-Pg plume, and concluded that their observed differences in chemistry could be related to different equilibration

temperatures, or to an original chemical heterogeneity within the plume. An alternative hypothesis of projectile ablation during atmospheric entry has been proposed by Robin et al. (1992) for the origin of the spinel, but is difficult to reconcile with the short entry time, and the observed global systematic pattern in spinel chemistry (Kyte and Bostwick, 1995).

[Table 1]

The distal K-Pg boundary contains shocked minerals that form around 1% of the layer (Hildebrand, 1993). Using observational data from Morgan et al. (2006) the total volume of shocked quartz within the global K-Pg layer is estimated to lie between 0.044 and 0.141 km³ (Table 1), and within the distal ejecta layer to be 0.017-0.053 km³ (> 3000 km from Chicxulub) . Shocked quartz grains from the K-Pg boundary show an inverse relationship between size and distance from the impact site (Bohor et al., 1987; Bostwick and Kyte, 1996). The largest shocked grains (500 to 1250 µm in diameter) are found in North America, Gulf of Mexico and Caribbean; much smaller (diameter < 190 µm) and less abundant grains are found at European sites, in the Pacific, and in New Zealand (Alvarez et al., 1995; Pope, 2002). The decrease in particle size (in both the spherules and shocked minerals) with paleodistance from Chicxulub is consistent with a single impact at Chicxulub. Shocked quartz and zircon within the distal K-Pg layer are thought to originate from Yucatán basement rocks that are buried beneath ~3 km of platform carbonates and evaporites at Chicxulub (Hildebrand et al., 1991).

Studies of basement clasts in impact breccias from the Chicxulub crater suggest that the uppermost basement rock is formed from silicate-rich rocks (Kettrup and Deutsch, 2003; Sharpton et al., 1996) with an average quartz content of ~25 % (Vermeesch, 2006). No quartz has been observed in drill core from the Cretaceous sediments, hence it is the basement rocks that are the most likely source of the shocked minerals within the K-Pg layer. This is supported by the observation that shocked zircon within the K-Pg layer in North America and Canada have

comparable age ranges to those in the Chicxulub impact breccias (Krogh et al., 1993; Kamo and Krogh, 1995).

2.2 Impact glasses at proximal sites

Spherule deposits containing glass have been found at the K-Pg boundary in deep-water carbonate sediments in the Beloc region of Haiti (Hildebrand and Boyton, 1990; Izett et al., 1991; Sigurdsson et al., 1991) and in NE Mexico (Schulte et al., 2003). The K-Pg section at Beloc, which was 860-910 km from Chicxulub at the time of impact (Hildebrand et al., 1991), consists of two distinctive spherule beds, which may be attributed to more than one depositional process. The basal unit (20-30 cm thick) is considered to be formed from ballistic fallout of glass spherules, based on normal grading, good sorting, and low carbonate content; while the upper unit, 20-40 cm thick, poorly sorted, is thought to be deposited from particle gravity flow such as ground surge from the impact itself or secondary, earthquakes/tsunami generated flows (Carey et al., 1993). The glass spherules range from 1 - 6 mm in diameter and constitute ~25% of the layer (up to 50% if altered spherules are included). The size and abundance of these glass spherules point to the close proximity of the impact site.

The chemical composition of the glass suggests a mixture of mafic rocks and Ca-rich sediments. These glasses resemble tektites as they are crystal-free, possess vesicles, and are volatile poor. The gas pressure in the vesicles is almost zero, indicating that the glass was formed from melt droplets that solidified in ballistic flight in the upper atmosphere. The chemical composition of black silica-rich (66-68% silica) glasses is similar to upper continental crust and correlates well with impact melt from the Chicxulub crater (Smit, 1999, fig. 14), while the Ca-rich yellow glasses represent a mixture of basement rocks with sediments (Sigurdsson et al., 1991). The proportion of yellow relative to dark glasses was initially estimated as 2-5% (Sigurdsson et

al., 1991; Lyons and Officer, 1992), but was considered to be higher by Mauresse and Sen (1991) who noted that all the least weathered glasses were high in Ca at some sites.

2.3 Summary

The first-order observational data that we consider to be reasonably well constrained, are: the volume and thickness of ejecta at proximal and distal sites, the volume and thickness of silica-rich glasses at proximal sites, and the volume of iridium and shocked quartz at distal sites. Hence, we focus on comparing our model outputs with these observations in section 4.

3. Numerical methods and initial conditions

We model the impactor colliding with the earth and high-velocity impact ejecta motion using the 3D hydrocode SOVA (Shuvalov, 1999) and the ANEOS equations of state for geological materials (Thompson and Lauson, 1972). SOVA is a two-step Eulerian code that can model multidimensional, multi-material, large deformation, strong shock wave physics. It is based on the same principles utilized in the well-known code CTH (McGlaun et al., 1990). A suite of benchmark/validation tests of impact hydrocodes (including the SOVA) have been performed recently and show that SOVA performs well and produces comparable result to other hydrocodes (Pierazzo et al., 2008; also Pierazzo et al. Meteoritics and Planetary Science, accepted for publication in August 2008). We use a tracer (massless) particle technique to reconstruct dynamic (trajectories, velocities), thermodynamic (pressure, temperature) and disruption (strain, strain rate) histories in all parts of the flow. Most of the results presented below are obtained by the analysis of these tracer particles. To model collapse of the plume and ejecta deposition we introduce real particles into the flow. We do not model the fragmentation process itself (i.e. crack growth). However, all solid/molten ejected materials subjected to tension (i.e. with density below normal

for a given temperature) are disrupted into fragments with a size-frequency distribution defined by the maximum shock compression (Shuvalov 2002; Artemieva and Ivanov, 2004).

Immediately after fragmentation, the velocity of these particles is the same as their original source material velocity. Subsequently, however, the velocity of individual particles may change relative to each other. For example, large fragments will move on ballistic trajectories that are independent of, and at a different velocity to, the expanding plume. In contrast, small fragments will travel as a passive component within the plume, and at plume velocity. The motion of particles within a gas/liquid mixture can be fully described using two- or multi-phase hydrodynamics. It should be noted that the meaning of “phase” here is not equivalent to “material”. Whereas, in one computational cell, phases may have different velocities, “materials” have different EOSs but must have the same velocity. In the modeling presented here we use the “dusty flow” approach to model individual “phases” (Boothroyd, 1971), which was implemented into the SOVA code by Shuvalov (1999). “Dusty flow” is, on one hand, a simplification of two-phase hydrodynamics (Harlow and Amsden, 1975; Valentine and Wohletz, 1989) as the direct particle-particle interaction and interphase mass exchange are neglected. On the other hand, it is a better approximation of the real flow as it allows particles of various sizes to be modeled, whereas two-phase hydrodynamics only deals with two phases, i.e. gas and one sort (size) of particles. In the “dusty flow” model, each fragment is characterized by its individual parameters (mass, density, position, and velocity) and is subjected to gravity and drag. Two types of drag forces are important: so-called high-velocity drag, $D_{hv}=0.5C_d\rho_a(\pi r^2/4)(V_g-V_d)^2$, where ρ_a is gas density, r is particle radius, V_d and V_g are particle and gas velocities, respectively, and C_d is drag coefficient (we assumed a typical value for spherical particles of ~ 1); and Stokes drag $D_{lv}=6\pi r\mu(V_g-V_d)$, where μ is the dynamic viscosity of gas. We can see from these equations that drag depends on the difference between particle and gas velocities, not on particle velocity alone. Particles also

exchange momentum and energy with the surrounding vapor-air mixture; gas parameters at each time step are initially defined by standard hydrodynamics and then are “corrected” according to their momentum and energy exchange with surrounding particles.

In our model large-scale turbulence is modeled directly, while sub-cell turbulence is taken into account as a turbulent viscosity in order to mimic particles’ transfer by eddies. Turbulent viscosity is proportional to eddy size (usually 10-20% of the cell size) and gas velocity within the cell (see Valentine and Wohletz, 1989). This viscosity may be of a similar value to a standard artificial viscosity that is used in modern hydrocodes to smooth gas dynamical parameters across a shock front. However, the physical meaning and applicability of these two viscosities is quite different.

Standard sedimentation process is a special case of dusty flow, in which gas may be treated as immobile and particles’ mass is negligible in comparison with gas mass. At the opposite limit, when total particles’ mass is substantially larger than gas mass, “dust” behaves almost like a solid piston. For example, massive ejecta re-entry may severely perturb the Earth’s atmosphere. In our models presented in section 4 we have ignored this effect, but we have explored ejecta re-entry further in the discussion in section 5.5. No-drag regime is another special case, when the particles and gas are moving with the same velocities (this is likely to be most important in the initial stages of plume expansion).

In summary, the "dusty flow" approach allows us to model the interaction between particles and gas, moving with different velocities, in a wide range of gas/particle mass ratios, from diluted flows, in which particles are a passive component, to quite dense particle flows, in which the particles may act as gas accelerator.

Routinely we use a constant spatial resolution of 12-15 particles per projectile radius, i.e. 500-600 m, within the “future crater” zone. The cell size increases towards the boundaries of the computational mesh, which is usually located at a distance of > 300 km from the crater center. This resolution is, as usual, a compromise between the accuracy of calculations and computer time/memory available. This compromise still exists in 3D models. We use “outflow” boundary conditions to minimize disturbances of the inner part of the flow from boundary related disturbances. However, the mass and velocity of all materials that reach the model boundaries are retained and used later to determine their final site of deposition (proximal or distal) or their escape from the Earth (using ballistic continuation above the Earth’s atmosphere).

We start with a simplified model of the Chicxulub target that is similar to previous models (Stöffler et al., 2004; Ivanov, 2005), and is comprised of a 3-km-thick layer of sediments with zero porosity (calcite EOS), a 30-km-thick crystalline basement (granite EOS), and mantle (dunite EOS). First, we have modeled a 45 degree impact (the most likely impact angle) into this target and refer to this as our standard model. We have chosen an impact velocity and impactor diameter that produces a transient cavity diameter of 90-100 km, in order that the modeled crater matches geological and geophysical observations (Morgan et al., 1997; Hildebrand et al., 1998; Collins et al., 2002). Then, we have run a number of other simulations with different impact angles and higher impact velocities, as well as 45°-impact into water-saturated sediments, to investigate the effect of changing these parameters on the ejection process.

For the wet sediments, we have used a model with 30% porosity. Physical properties of Cretaceous sediments at the Chicxulub impact site have been measured and estimated at Yax-1 borehole (Mayr et al., 2008). Anhydrite samples within Yax-1 typically have a porosity of a few percent, while dolomite and calcarenite have more variable porosities that range from a few to up to 20%. The Cretaceous sediments in Yax-1 were almost certainly fractured on impact, and then

subsequently affected by chemical alteration and crack closure through compaction. Porosity may have been higher at the time of impact than is currently observed, and we use a 30% wet porosity value for the entire sedimentary section, in order to investigate the maximum possible effect that porosity may have on vapor expansion. Part of the rationale for using a wet porosity is to investigate whether it leads to a significant increase in ejection velocity, as suggested by Kieffer and Simonds, 1980 (Artemieva, 2007). Specific strength properties of porous rocks were not taken into account, however special EOS, assuming a total thermodynamic equilibrium between water and calcite, has been constructed similar to a basalt-water mixture (Pierazzo et al., 2005). We also ran one impact simulation into a target with a 3-km-thick water layer (instead of sediments), as a recent seismic survey indicates a deep basin, with > 2 km water, existed in the north and northeastern quadrant of the impact site (Gulick et al., 2008).

For a vertical impact we determine the size of the transient crater using the efficiency parameter $D_{pr}V_{pr}^{0.58}$ (Dienes and Walsh, 1970), and use units of km and km/s for projectile diameter and velocity. We use similar parameters to Ivanov (2005), that led to a good fit between modeling and observations at Chicxulub, with an impact velocity V_{pr} of 18 km/s (slightly higher and more realistic) and projectile diameter D_{pr} of 12 km instead of 14 km. We have changed the projectile size with impact angle so as to maintain a constant transient crater size of 90-100 km diameter, and this equates to a 14.4-km-diameter projectile for a 45° impact, and 16 km for a 30° impact. The laboratory-derived scaling (Chapman and McKinnon 1986) used here assumes that only the vertical component of impact velocity is efficient in crater excavation, whereas numerical models suggest that this may be incorrect for natural impacts (Ivanov and Artemieva, 2002). Thus, we may overestimate the projectile size (and hence, amount of ejecta) in our modeling of oblique impacts. We have also modeled one high velocity impact (36 km/s), and use standard scaling

(Schmidt and Housen, 1987). In all runs we use a projectile density of 2600 kg/m³ as this is typical for large asteroids with a porosity of 20-30% (Britt et al., 2003).

Below we summarize the results for five of our simulations: 1) 45° impact angle, velocity 18 km/s (the standard model), 2) 30° impact angle, velocity 18 km/s, 3) 45° impact angle, velocity 36 km/s, 4) wet target, 45°, velocity 18 km/s, 5) wet target, 90°, velocity 18 km/s.

4. Results

The maximum pressure recorded by tracers in the projectile and target rocks during our simulations is used to estimate the total volume of melt and vapor produced for each material during impact (Table 2). According to the EOSs in use, basement rocks are molten after shock compression above 55 GPa, while vaporization starts at 140 GPa (and wt% of the vapor may be defined for each value of shock compression). Sediments (we do not separate them into anhydrite and calcite) are decomposed at 60 GPa and pore water is vaporized at 10 GPa. If any target/projectile tracer reaches a position above the pre-impact surface, it is treated as “ejecta” from the crater. We have also monitored particle velocity, and use this to estimate the total volume of ejecta that either: escapes Earth’s gravity (i.e. ejecta velocity is above 11 km/s), reaches distal and proximal sites, or remains close to or within the crater (Table 3).

[Table 2] [Table 3]

We note that, in all impact scenarios, > 80% of decomposed sedimentary materials leave the crater, while only 5-12% of molten basement is ejected. The projectile is almost entirely molten and, on average, 75% of this melt is ejected from the crater. At the highest impact velocity of 36 km/s nearly all projectile material leaves the crater.

4.1 Modeled distal ejecta

In our standard modeling simulation, we noted a surprising result: basement (both molten and solid) travels at fairly low velocities immediately after ejection (Fig.1) and ten seconds after impact, the maximum velocity of basement material is 2.7 km/s in the downrange direction. In addition, only a small volume of basement material is ejected in the uprange direction and at very low velocity (Fig. 1). More highly compressed material (melt) is ejected more asymmetrically than solid material, with more ejecta traveling in the downrange direction.

[Fig.1] [Fig.2]

Figure 2a shows the material (projectile, sediments, basement) in the expanding plume for the standard model, 35 seconds after impact and the ejected mass and velocity of these materials (Fig. 2b) at various altitudes. The high-velocity ejecta material in the upper part of the plume is composed of vaporized projectile and target sediments, while the lower part of the plume is composed of shocked and molten basement, and solid sediments, traveling at relatively low velocities. After 35 seconds, all the ejected basement material is below an altitude of 70 km and has a maximum velocity of 2.5 km/s (Fig. 2). The maximum distance reached by the basement rocks traveling on ballistic trajectories with these velocities is about 700-1000 km. This result (relatively low ejection velocity of the basement material) appears to be in accordance with the observation that high-velocity ejecta, such as tektites (Koeberl, 1994; Stöffler et al., 2002) or meteorites from other planets (Melosh, 1984; Nyquist et al., 2001; Artemieva and Ivanov, 2004), originate exclusively from the uppermost layers of the target (spallation zone) and not deeper than 1/10 of the projectile diameter. At Chicxulub, onshore wells show that crystalline basement lies ~3 km below surface (as in our model) and may be even deeper offshore (Bell et al., 2004; Gulick et al., 2008). This corresponds to a depth of $\sim 0.2D_{pr}$, and confirms that we should not expect shocked basement rocks to be expelled from the impact site at high velocity.

Assuming ballistic trajectories, we have used our mass-velocity distribution of ejecta to determine global ejecta thicknesses (Fig. 3). The total amount and estimated thickness of ejecta is comparable with observations, with a few cm at intermediate distances of 1800-2500 km (e.g. in North America), and a few mm thick worldwide (at distances > 4000 km). However, a model of pure ballistic continuation inevitably leads to ejecta thickness decreasing with distance from the crater (Fig. 3). Hence, one of the first-order observations – the almost constant 2-3 mm thickness of the global K-Pg layer, is not replicated in our modeling.

[Fig. 3]

In the standard model (Table 3, first row) 5% of the projectile material escapes the Earth, 31% remains within the final crater (not ejected or ejected with velocity below 1 km/s), and another 30% is deposited within the first thousand kilometers from the crater (ejection velocity between 1 and 3 km/s). Around 14% of the projectile has a velocity between 5 km/s and 11 km/s, and may be distributed worldwide, which is equivalent to $\sim 210 \text{ km}^3$ of projectile material. If sediments and projectile with velocities between 5 and 11 km/s (13.6% and 35.4% of the projectile volume, respectively, see Table 3) are distributed around the globe in a similar way, then the distal ejecta would be composed of sediments (72%) and projectile (28%), and the total volume of material is $\sim 770 \text{ km}^3$. This is slightly less than 850 km^3 of K-Pg spherules world-wide estimated by Smit. (1999). However, we expect some back reactions to occur (and hence, additional energy release) in the plume, as suggested by the calcite-rich spherules in proximal sites like Belize (Smit, 1999), so our modeled volume of 770 km^3 for the distal ejecta is considered to be a minimum estimate. There is no basement material in the modeled distal ejecta. In this model 0.14 of the projectile may reach distal sites, and this is significantly lower than estimates of 0.22 by Alvarez et al. (1980) and 0.5 by Vickery and Melosh (1990). This is an important result as it means that the volume of iridium within the global ejecta layer is estimated to be $2.77 \times 10^8 \text{ kg}$ (assuming an Ir

concentration of 500 ng/g) and is similar to the observed volume ($2\text{--}2.8 \times 10^8$ kg). Thus, in our standard model, if we assume the observational data is accurate, we have a small volume deficiency (10%), but not an Ir-deficiency in the fireball layer. Our results suggest that distal spherules are composed from projectile and sediments in a ratio of $\sim 1:3$. If the clay that surrounds these spherules was dominantly locally derived, or if the clay and some spherules originate from low-velocity (projectile-poor) ejecta that has travelled non-ballistically (see section 5.5), then the meteoritic ratio would be $\sim 1:6$, which is in a good agreement with observations (Kyte, 2004).

One other consequence of ballistic ejection, is that sites between 1000 and 4000 km (ejecta velocities > 3 km/s and < 5 km/s), have an excess in Ir of 4.1×10^8 kg, and a total volume of ~ 1500 km³. Iridium concentrations are not particularly high at these distances, and the observed 1-2 cm thick K-Pg layer in North America integrates to a volume of only 500-1000 km³. Hence, two more first-order properties of the K-Pg layer are not modeled well: 1) there are no shocked minerals from the Chicxulub basement at distal sites, and 2) the volume of ejecta in North America is too high (by a factor of ~ 2).

Part of the rationale for simulating a number of impact scenarios was to investigate the sensitivity of the ejection process to impact angle and target structure. The results for Model 5, a vertical impact (last row in Table 3), are distinctly different from the others: there is no high-velocity projectile material and the volume of high-velocity sediments is extremely low. In all models, basement material is ejected at velocities of less than 5 km/s, and would therefore not reach distal sites on ballistic trajectories. The total volume of material ejected at velocities of > 5 km/s is between 770 and 1630 km³, with a maximum for the most oblique impact. The ratio between projectile and sediments varies from 0 to 1, and the maximum volume of projectile material with velocities > 5 km/s is 840 km³, obtained with the 30° impact. It is difficult to establish a general trend from 5 models but: low impact angles of $\leq 30^\circ$ appear to disperse too

much projectile world-wide, while sub-vertical impacts appear to disperse too little. Impacts into a water-saturated target are similar to the dry target, with only a slight increase in volume of ejected projectile and sediments, and high-velocity ejecta. The three models with 45° impact angles (1, 3, and 4) give similar world-wide Ir masses of $2.8\text{--}3.5 \times 10^8$ kg, and are reasonably close to the observed estimate of $2\text{--}2.8 \times 10^8$ kg. The projectile to sediment ratios in the ejecta are between 1:3 and 1:4.

Alvarez et al. (1995) used analytical estimates and O’Keefe et al. (2001) used early stage modeling to suggest that basement material is accelerated to high velocity within the expanding plume. In order to investigate this, we created three hemispheres (with radii of 100, 200, and 300 km) around the growing crater, and examined material flow between spheres to document any particle acceleration. Analytical continuation of ejecta mass-velocity distribution from a 100-km-radius sphere to a 200-km-radius sphere, assuming pure gravitational deceleration, is confirmed by our numerical modeling (Fig. 4a,b).

[Fig. 4]

Our models show that there is no substantial acceleration within the lower part of the plume. Essentially, acceleration is only possible if at least one of three conditions is fulfilled in the plume: (1) solid ejecta are surrounded by hot expanding vapor; (2) turbulent mixing of slow and fast parts is fast and efficient; (3) the plume itself is buoyant (as it was in the case of Shoemaker-Levy 9 collision with Jupiter). None of these conditions operate in the case of the Chicxulub plume: (1) solid/molten basement ejecta are surrounded by sediments that were compressed to even lower pressures (as sediments are closer to the surface), hence vaporization in the lower part of the plume is minimal; (2) turbulence does not operate efficiently on a short time scale and on a large spatial scale; moreover, velocity distribution within the plume is linear in agreement with the standard analytical solution for gas expansion into vacuum (Zeldovich and Raizer, 1966); (3) the

lower part of the plume, containing basement ejecta, is not hot and rarefied, and hence, buoyant (moreover, it is not surrounded by atmosphere, which has been removed from the impact site earlier by the expanding plume). Thus, acceleration within the plume can not be invoked to explain the ejection of shocked quartz from basement rocks worldwide. In the discussion, we explore alternative explanations for the presence of shocked quartz within the distal K-Pg layer.

4.2 Modeled proximal ejecta

Our estimates of the total melt production from the basement (18,000-28,000 km³, depending on impact angle and impact velocity) are in a good agreement with simple scaling-based estimates (Kring, 1995; Pope et al., 2004), geophysical studies (Ebbing et al., 2001; Morgan et al., 2002; Hildebrand et al., 2003) and modeling results using different hydrocodes (Pierazzo and Crawford, 1998; Ivanov, 2005).

Using pure ballistic continuation for the basement tracers, subjected to shock compression above 50-60 GPa (i.e. molten after decompression) we estimate melt deposition at proximal sites (Table 2). The majority of the melt (~80%) lies within or near the crater even at late time moments, and would be a component of melt rocks or melt rich (suevitic) impact breccias. Around ~3% of the total melt produced is ejected from the crater with velocities above 1 km/s and deposited outside the crater rim. In Fig. 5 we show two examples (Model 1: standard model, and Model 4: wet sediments) of the distribution of basement melt around Chicxulub. The thickness of the melt deposit is about 10 cm - 1 m at a distance of 700-1000 km for the wet target, and this is comparable with the observed volume of Si-rich impact glass deposits.

[Fig. 5]

5. Discussion

There are three first-order observations that are not well-modeled in our simulations: shocked minerals from the Chicxulub basement rocks have velocities that are too low to travel to distal K-Pg sites on ballistic trajectories, the volume of ejecta in North American sites is slightly too high, and the thickness of ejecta at distal sites decreases with distance from the crater. Here we explore some possible explanations for these discrepancies: models with a different impact angle and/or velocity, models with a different target stratigraphy, errors in the code and non-ballistic mechanisms of transport.

5.1. *Models with different impact angles and velocities*

Highly oblique impacts lead to higher ejection velocities, but also to substantially shallower excavation depths (Gault and Wedekind, 1978; Schultz, 1996; Pierazzo and Melosh, 2000). Impacts close to vertical have maximum excavation depths, but lowest ejection velocities. Models 1, 2, 4, and 5, with various impact angles and a velocity of 18 km/s (Table 3) do not produce high-velocity (> 5 km/s) basement ejecta. A high-velocity impact of 36 km/s (similar to cometary impacts) leads to the ejection of 13 km^3 of basement material with velocities above 3 km/s (but still below 5 km/s). This is enough material to produce the global shocked quartz volume (Table 1), but the velocity is too low for the quartz to reach distal sites through ballistic transportation.

5.2 *Models with a different target stratigraphy*

As already discussed, water content in sediments has only a minimal effect on the basement ejection velocities. In our tests, the only scenario that led to high-velocity basement ejecta, was an impact into a 3-km water layer directly above basement. This is compatible with an impact into oceanic crust, a scenario that was discussed intensively in the 80-ies after the K-T bolide hypothesis had been widely accepted, but before Chicxulub was discovered (Alvarez et al., 1980; Montanari et al., 1983; Smit and Kyte, 1984; Hildebrand and Boyton, 1990). However, the lack of quartz in oceanic crust, the Yucatán continental basement age of zircon in K-Pg ejecta (Krogh et al., 2003), and the observation that the Cretaceous sediments around the Chicxulub site were 2.5-3 km thick onshore (Lopez Ramos, 1975; Ward et al., 1995) and > 3 km thick offshore (Bell et al., 2004; Gulick et al., 2008; Collins et al., 2008) all suggest that this scenario is unlikely.

The presence of shocked quartz in the distal ejecta layer could also be explained if quartz were a minor component of Upper Cretaceous sediments. This also seems unlikely, as 1) no quartz has been reported in Cretaceous sediments in Yucatán drill core to date (Koeberl, 1993; Ward et al., 1995) and 2) PDF in quartz from sedimentary rocks have different characteristics to PDF in crystalline rocks (Grieve and Theriault, 1995). However, we cannot dismiss this as a possible explanation, as we only require a quartz abundance of less than 1% to create the known shocked quartz distribution, and the sampling of the Upper Cretaceous is imperfect – the holes are sparsely distributed, old holes are only partially cored, and the uppermost Cretaceous is missing in some wells (Urrutia-Fucugauchi et al., 1996).

5.3 Errors in the code and EOS

There are unlikely to be any fundamental errors in the SOVA code: the hydrocode has been thoroughly tested; mass, momentum and energy conservations are under strict control during the runs; the same code is now routinely used to model high-velocity ejecta such as tektites (Stöffler et

al., 2002) and Martian meteorites (Artemieva and Ivanov, 2004). Following an investigation into the effect of resolution on model results, a spatial resolution of 12-15 cells per projectile radius (as used here) was considered sufficiently high for 3D impact models. A model that was run at twice this resolution, for the initial stages of impact, did not lead to any substantial differences in either maximum shock compression or ejection velocity. One possible limitation is that it may lead to a slight underestimate in the amount of molten material (Pierazzo et al., 1997) but not in ejection velocities (Artemieva and Ivanov, 2004).

It is well-known that ANEOS does not accurately reproduce vaporization in geological materials, and that this might lead to an overestimate of velocities with the expanding vapor plume. Recently Melosh (2008) substantially improved the existing EOS for quartz by introducing molecular clusters. Tests did show a substantial increase in the amount of vapor (from 10% to 30% for a shock compression of 242 GPa), but, at the same time, plume expansion velocities were essentially unchanged (Melosh, 2008, fig. 11). Moreover, the plume without molecules expanded even faster at the beginning because of slower pressure decay. Hence, there also does not appear to be a fundamental problem with the EOS.

One final plausible reason for a discrepancy is that we have not taken into account back reactions in the plume (and energy release), and there is observational data that suggests calcite is being formed within the plume and being distributed locally. The thick deposits of calcite-rich spherules in Belize are evidence for the occurrence of back reactions (e.g. Smit, 1999). However, this is unlikely to affect the velocity of shocked quartz as these materials are substantially separated in the plume.

5. 4 Small particles suspended in the plume and slow non-ballistic transport

One other possible mechanism for ejecting basement minerals at high velocity is through small particles being turbulently incorporated into the plume. While standard hydrodynamics describe a continuum media, i.e. all materials ejected at the same place and at the same time move along the same trajectories, the actual situation may be different: vaporized rocks expand separately from solid fragments; large boulders follow ballistic trajectories, while small ones and molten droplets may be turbulently incorporated into the plume. Hence, in addition to our standard modeling (presented in Tables 2 and 3) we attempt to model this “non-ballistic” transport using “dusty flow” hydrodynamics by creating particles instead of solid-molten continuous medium. In this late-stage modeling (up to 15 minutes), when all ballistic ejecta with velocity of 2-3 km/s should have reached the upper atmosphere close to its final destination, about 0.05 km³ of grains from the basement eventually reach the boundaries of our computational box (located at a distance of 1000 km from the impact site and at an altitude of 500 km) and escape it. There are another 1.2 km³ of basement material at lower altitudes (< 200 km), at distances of 1000 km from the impact site in the downrange direction. The size of these particles ranges between 1 μm (the lower limit in modeled SFD) to 3 mm, and the maximum shock compression of the particles is from 20 GPa to 55 GPa (melting point), i.e. pressures that can produce PDFs in quartz. As all these droplets have been subjected to turbulent mixing, we do not see any correlation between velocity and shock pressure, or velocity and particle size. In addition, the spatial distribution is more symmetric than the initial one. The total amount of this “low-velocity turbulent” ejecta is much lower than would be required to create a worldwide ejecta layer with an average thickness of 2-3 mm, but is enough to produce the total volume of shocked quartz distributed worldwide (Table 1).

These suspended particles (including shocked quartz grains) could be dispersed around the globe by atmospheric flows and slow gravitational settling, similar to volcanic ashes such as in the El Chichon eruption in 1982 (Rampino and Self, 1984). This mechanism was proposed by Toon

et al. (1997) and Covey et al. (1990), and modeled by Pope (2002) who assumed 10^{16} - 10^{17} g of clastic material at stratospheric altitudes near the Chicxulub crater. This variant looks attractive because, such non-ballistic transport of quartz grains at low velocities and low altitudes, would allow PDF to survive upper atmosphere heating during re-entry. If the shocked quartz had traveled ballistically at high-velocity, the shocked quartz should have annealed (Croskell et al., 2002). This raises the question as to whether we can distinguish between this ballistic or non-ballistic transport on the basis of grain-size distribution. Unfortunately we cannot, because both would lead to an inverse relationship between maximum grain size d_{max} and distance from the crater D.. Modeling results give a similar inverse relationship, but with a different exponent. In the case of ballistic transport grain size decreases with distance because, on average, larger grains are ejected with lower velocity (Grady and Kipp, 1980; Melosh, 1989, p.107) and $d_{max} \sim D^{-1/3}$. In the case of non-ballistic distribution by atmospheric flows, this distance is defined by settling time, i.e. by particle size, and $d_{max} \sim D^{-1/2}$. The observed maximum grain size decreases more rapidly than predicted by either transport mechanism ($d_{max} \sim D^{-0.6}$, Morgan et al., 2006).

5.5 Floating of impact debris and fast non-ballistic transport

Colgate and Petschek (1985) proposed a mechanism for lateral transport of ejecta termed “floating of impact debris in the atmosphere”. Their idea arose from the observation that large volumes of ejecta slid sub-horizontally around Jupiter hours after the Shoemaker-Levy impact (McGregor et al., 1996). The principal idea is quite simple: the re-entering of debris heats the atmosphere, which then expands upwards and outwards, and redistributes the debris. Unlike for atmospheric flows, which take several weeks to transport volcanic ashes globally, this mechanism may allow global re-distribution of the ejecta in several hours. However, Colgate and Petschek (1985) presented the concept only, and did not attempt to model the phenomena.

Here we present a simplified model of ejecta re-entry using SOVA complemented by dusty flow procedure, to investigate whether the floating debris mechanism could lead to a re-distribution of ejecta around the globe. We assume that the ejecta re-enters an undisturbed atmosphere (this is true for the initial arriving ejecta, but not for later arriving ejecta). In Fig. 6a a beam of ejecta is shown arriving at an altitude of 100 km with a velocity of 2 km/s. The beam of ejecta is 25-km in diameter and, if not re-distributed, would have formed a 25-km wide layer that was 14 cm thick at surface. The velocity and volume of ejecta is approximately equal to that of ejecta arriving ~1000 km from the Chicxulub impact site. We observe that, in the upper atmosphere, the ejecta moves ballistically and transfers its energy to the gas, creating shock waves (visible as darker gray in color in comparison with stratified atmosphere). At lower altitudes, atmospheric disturbances become strong enough to disperse ejecta – first uprange (Fig. 6c) and later downrange (Fig. 6e-f).. In Fig. 7 we show the resulting deposit thickness for the model in Fig. 6. Two gray rectangles show the fate of the ejecta if it had passed straight through the atmosphere by: pure ballistic travel without substantial deceleration (B) and quick deceleration in the upper atmosphere followed by slow sedimentation (S). In the case of ejecta being re-distributed by the expansion of the atmosphere, the ejecta thickness is ~2 cm in the center and more or less constant at 1-mm-thick at distances of up to 400 km downrange and uprange. Approximately 1/3 of the ejecta are deposited at greater distances, outside the area shown in the figure.

[Fig. 6] [Fig. 7]

The real re-distribution, and final ejecta thickness, would depend on the ejecta mass, velocity, and particle-size distribution. Hence, the picture is more complicated than suggested in Figs. 6 and 7, as in reality we would have ejecta arriving at all distances, traveling with a range of particle velocities, and different masses per unit area. Of importance, smaller particles are

dispersed more intensively, and are thus more likely to end up on the other side of the globe. Hence, this mechanism is consistent with the observation that the size of shocked quartz gradually decreases with distance from Chicxulub. We also note that, the dispersion process should actually be even more intensive than shown, as the later arriving ejecta will enter an atmosphere that has already been disturbed by the initial, fast-arriving, ejecta. This implies that particles suspended in the expanding plume, as well as proximal ejecta with velocities of 2-3 km/s, may be efficiently redistributed by this mechanism to much larger distances, and within a comparable time interval, compared with pure ballistic transport.

If we assume that all high-velocity ejecta (above 4-5 km/s) is distributed non-ballistically and evenly by this method, then the global thickness is roughly equal to 2-3 mm (see square and triangle in Fig. 3). In addition, this transport mechanism would: reduce the total volume of ejecta arriving at North American sites (which is slightly too large in our models), increase the volume of material arriving at distal sites (which is slightly too small in some of our models), as well as enable shocked quartz and zircon to reach distal sites without being annealed. These preliminary results are quite encouraging, as they address all the discrepancies between observational data and models with purely ballistic transport.

5.6 Deposition time

Thus, we suggest that some ejecta material are deposited ballistically (spherules) and other ejecta non-ballistically (probably some spherules, shocked quartz and zircon, and fine particles). This leads to the question – why do they jointly form a single thin layer instead of two distinct layers? Ballistic flight is shorter than half an hour, while non-ballistic transport by the floating debris mechanism could take several hours to reach the most distal sites. However, the time of ballistic flight above the atmosphere is much shorter than the settling time through a dense

atmospheric and water layer (for marine sites). Even in the case of massive ejecta re-entry (Goldin and Melosh, 2007), settling velocity remains high only in the upper atmosphere and later, at an altitude of 50 km and below, is defined by standard Stokes's law. Thus, settling time for the finest particles (50 μm) in the atmosphere or in deep water exceeds 10 days. Most probably, ballistic and non-ballistic ejecta from the Chicxulub crater were finally deposited together to form one single layer within a couple of weeks of the impact.

6. Conclusions

We have modeled some first-order observations well, and have performed a simple pilot study on a mechanism for non-ballistic transport that appears to explain the most significant mismatches between models and observational data. We conclude that:

1. Spherules within the distal K-Pg boundary layer were formed from vaporized/molten projectile and sediment in a ratio from 1:3 to 1:4, and were principally emplaced ballistically;
2. Impact glasses from proximal sites (~900 km from Chicxulub) originate from porous wet sediments mixed with crystalline basement in our model, as indicated by observations;
3. Models with impact angles of around 45° can reproduce the observed world-wide iridium volume, whereas sub-vertical impacts and oblique impact angles of ≤ 30 degrees cannot;
4. According to our numerical modeling and analysis of various mechanisms of ejecta emplacement, shocked quartz and zircon within the K-Pg layer were most likely emplaced non-ballistically by winds (slow transport) or by a mechanism termed

“floating of impact debris in the atmosphere” (fast transport); the latter looks more attractive as it does not involve extremely small particles, and particles arrive at their final destination at roughly the same time as the ballistic ejecta;

5. The non-ballistic mechanisms may explain why the K-Pg distal ejecta thickness is close to constant.

6. All components of K-Pg layer were deposited within a few days (up to a couple of weeks for the finest fractions) after the impact.

Our modeling allows us to establish a relationship between the Chicxulub target composition, and ejecta thickness/composition at various distances from the impact site. The results are important for future investigations of the environmental effects of the Chicxulub impact.

Acknowledgements. IARC contribution number 2008-1056. This work was supported by PRARC grant PP/E001513/1. We are grateful to our colleagues for interesting discussions. The paper was substantially improved with the help of two anonymous reviewers.

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Figure captions

Fig. 1 Distribution of ejecta from the crystalline basement versus azimuthal angle (0 is downrange, 180 – uprange) at 15 seconds after the impact for a) ejected mass and b) ejection velocity. This distribution is symmetrical over the Y-axis, i.e. is the same for negative angles. Gray is for melt, black – for solid ejecta. Solid lines on velocity graph show average velocity, dashed lines - maximum velocity for a given azimuth. Highly compressed ejecta (impact melt) are distributed more asymmetrically and have higher velocity than solid ejecta.

Fig.2 a) The Chicxulub crater and its plume 35 seconds after impact. Green is basement, gray is projectile and sediments, light blue is atmosphere, dark blue is mantle. b) Mass of ejecta (with 20 km altitude step) and average ejecta velocity (black line) as a function of altitude 35 s after impact. While upper part of the plume, containing projectile and sediments has velocity above 6 km/s, its lower part (mixture of sediments and basement) expands with velocity of 2 km/s or even less.

Fig. 3 Cumulative mass of ejecta (left axis) versus ejection velocity (bottom axis) is shown by black lines. Corresponding ballistic thickness of ejecta (right axis) as a function of distance (top axis) is shown with gray lines. Solid lines are for dry target, dashed ones - for water-saturated sediments. Squares (dry target) and triangles (water saturated sediments) represent worldwide ejecta thickness if high-velocity ejecta (with velocity higher than on velocity-axis) are deposited non-ballistically and evenly.

Fig. 4a Basement ejecta volume passed a 100-km-radius hemisphere (bars and left axis) and its average (solid lines, right axis) and maximum (dashed line, right axis) velocities as a function of azimuthal angle. Gray colors represent all basement material, black – molten material. Left graph is for the dry target, right graph – for the target with water-saturated sedimentary layer. While mass distributions are similar, ejection velocities are substantially higher in the latter case (water in the target).

Fig. 4b. Comparison of the real molten ejecta volume passing 200-km-radius hemisphere (black bars) assuming ballistic continuation from 100-km-radius hemisphere (shaded bars). There is no acceleration of basement ejecta in the plume between these two radii.

Fig. 5_ Thickness of molten basement ejecta (Haitian glasses) near the crater (shown by circle at position 0,0) for impact into a) dry target, and b) water-saturated target. Only high-velocity ejecta (which reached the 100-km-diameter hemisphere) is shown, i.e. thickness of ejecta near the crater rim is underestimated. The impact into a wet target produces roughly the observed thickness of proximal ejecta in Haiti.

Fig. 6. Shock waves in atmosphere and dispersion ejecta as it re-enters the atmosphere. Atmosphere thickness (Y-axis) is 100 km, numbers along the X-axis show distance from the re-entry point in km. Y-scale shows 100 km of the Earth's atmosphere. Numbers in the upper right corner of each snapshots are time in seconds.

'Fig. 7 Solid line shows the final ejecta distribution for the model shown in Fig. 6, in which ejecta is re-distributed by the floating debris mechanism. The two rectangles show the location and

969 thickness of ejecta that would have been deposited if the ejecta had not been re-distributed, and
970 had traveled through the atmosphere by either ballistic continuation (B) or by deceleration in the
971 upper atmosphere followed by vertical Stoke's sedimentation (S).'

972

973 **Table 1**

974 Estimated amount of shocked quartz

Distance, 10 ³ km	Mean size, μm	Nos/cm ²	Volume, km ³
< 3	60-80	800-1100	0.027-0.088
3 – 6	45-55	300-400	0.011-0.028
> 6	35-45	70-130	0.006-0.025
Total			0.044-0.141

975

976 **Table 2** Melting and vaporization of target materials and projectile (total volume; ejected volume
977 – in parenthesis)

978

979

	Dry, 14.4 km, 45°, 18km/s	Dry, 16 km, 30°, 18 km/s	Dry, 10 km, 45°, 36 km/s	Wet, 14.4 km, 45°, 18 km/s	Wet, 12 km, 90°, 18 km/s
Sediments, km ³ > 60 GPa > 10 GPa	1924 (1679) 11186 (10941)	2619 (2139) 10647 (10116)	2376 (2363) 14899 (11532)	1823 (1648) 14899 (11532)	544 (448) 5694 (5598)
Basement, km ³ Melt Vapor	24905 (3304) 224	21092 (946)	34866 (2624) 1980	25255 (3086) 226	21858 (1984) 250
Projectile, melt (%)	97.6 (77.6)	92.5 (88.4)	100 (96.6)	96.5 (76.9)	100. (11.4)

980

981

982 **Table 3.** Cumulative material distribution over ejection velocity (in % of projectile volume).

	Escape, 11 km/s	> 8 km/s	> 5 km/s (global)	> 3 km/s (proximal)	> 1 km/s (outside crater)
Dry, 14.4 km, 45°, 18km/s	5.6 (5.2) * 4.5 (4.5) -	9.4 (9.1) 12.9 (12.1) -	19.2 (18.9) 39.9 (29.5) -	39.6 (38.8) 118. (62.4) 0.44 (0.44)	68.8 (66.8) 566.(99.0) 134. (83.2)
Dry, 16 km, 30°, 18 km/s	27.5 (27.5) 6.9 (6.9)	40.2 (36.4) 14.8 (12.9)	66.7 (61.1) 43.8 (24.6)	81.8 (75.4) 104. (40.8)	94.5 (87.3) 420. (85.7)

	-	-	-	-	51.9 (15.1)
Dry, 10 km, 45°, 36 km/s	12.4 (12.4) 32.2 (31.5) -	24.5 (24.5) 72.7 (71.6) -	53.5 (53.5) 199. (174.) -	79.8 (79.8) 446. (315) 13.0 (13.0)	96.5 (96.5) 2066 (449) 452. (406.)
Wet, 14.4 km, 45°, 18 km/s	3.0 (2.5) 6.1 (6.1) -	7.2 (6,4) 19.2 (15.9) -	20.4 (19.5) 62.2 (40.2) -	41.8 (40.4) 145. (69.7) 5.4 (5.4)	72.8 (69.9) 637. (99.6) 191. (87.)
Wet, 12 km, 90°, 18 km/s	- - -	- 0.07 (0.07) -	- 6.1 (3.2) -	- 54.0 (18.3) -	5.0 (5.0) 593 (48.0) 97.6 (86.0)

* In parenthesis – molten or degassed material. In each cell the first line is for the projectile, the second - for the sediments, the third – for the basement.

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