Impacts do not initiate volcanic eruptions: Eruptions close to the crater

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Abstract:

A theme that runs through many papers on meteorite impact is the idea that large impacts can induce volcanic eruptions through decompression melting of the underlying rocks. We perform numerical simulations of the impact of an asteroid with a diameter of 20 km striking at 15 km s^{-1} into a target with a near surface temperature gradient of 13 K km^{-1} ("cold" case) or 30 K km^{-1} ("hot" case). The impact ultimately creates a 250 to 300 km diameter crater with approximately 10,000 km^3 of impact melt. However, the crater collapses almost flat and the pressure field returns almost to the initial lithostat. Even an impact this large is insufficient to raise mantle material above the peridotite solidus due to decompression only. We also discuss the probability that such an impact coincides with the much more frequent occurrence of Large Igneous Provinces and conclude that it is unlikely that a large impact struck such a province any time in post-heavy bombardment Earth history.

Introduction:

A theme that runs through many papers on meteorite impact is the idea that large impacts can induce volcanic eruptions (one of the first was Ronca, 1966). This idea probably got its start in pre-Apollo days when early observers of the moon noted the common occurrence of dark material—usually supposed to be basaltic lava—filling the nearside impact basins (Baldwin, 1963). A logical inference is that this is a genetic association: The impacts caused lava to upwell in the biggest craters after they had formed, eventually filling them. This view should have collapsed in 1965, when the Russian spacecraft Zond 3 made good photos of the lunar farside showing that the farside basins are not filled with basalt. Moreover, the samples returned from the moon by the Apollo missions showed that the mare basalts are considerably younger (up to about 1 Gyr) than the basins in which they lie (Wilhelms, 1987). Any presumption of a genetic association of impacts with volcanism on the moon must thus be deemed questionable. It seems more likely that the large nearside basins were merely the lowest spots on the moon's surface at the time that the lunar interior warmed to the point where basaltic partial melts formed in its mantle, and that the rising lava simply flowed to the lowest points. At most, this rising lava might have flowed up impact-induced fractures in the lunar crust.

Nevertheless, the idea that impacts may initiate volcanic eruptions on Earth or even the Moon has been vetted many times in recent years (Elkins-Tanton et al., 2002; Glickson, 1999; Jones et al., 1998; Jones et al., 2001; Rampino, 1987; Rogers, 1982; Ryder, 1994). Most of these proposals reiterate the original idea of Ronca (1966) that ascribes impact-induced volcanism to decompression melting beneath the impact crater.

Stratigraphic Uplift beneath Craters:

The naive idea that an impact crater unloads layers deep under the crater needs careful examination. It is true that crater excavation removes and redistributes large volumes of target material. However, the final crater differs dramatically in shape from the first-formed "transient crater". The transient crater is briefly established when the growing crater reaches its maximum volume. The freshly shattered rock near the impact is too weak to retain the transient crater at its maximum depth. The crater collapses under gravity, transforming it to the final complex form as material is both uplifted beneath the crater floor and slides centripetally inward from the rim of the transient crater (see Collins et al., 2002; Melosh and Ivanov, 1999). The final crater's depth is much shallower than the transient cavity. Consequently, the pressure release under the crater is much smaller than some recently published estimates (Jones et al., 2001).

Impact Melt

Melt is very common near large impact craters. This melt is created by the strong shock waves that emanate from the site of the impact. These shocks first compress the underlying target rock, doing irreversible work on this material, and then release it adiabatically to low pressure. If the shock is strong enough (typically 50 GPa or more) the released material may be molten, or even vaporized. However, since the shock waves are strong only very close to the impact site, the melt lines the interior of the crater as it opens and later collapses. With the possible exception of a few deeply injected dikes, it does not mix into the target rocks below the crater floor. The melt thus forms a near-surface sheet throughout the crater formation and collapse processes. Recent hydrocode studies (Pierazzo et al., 1997) are in excellent agreement with earlier estimates of the volume of melt produced in an impact (Ahrens and O'Keefe, 1977; Grieve and Cintala, 1992). In terms of the projectile volume V_p, the volume of melt V_m produced in a vertical impact is given approximately by:

$$\frac{V_m}{V_p} = 0.25 \frac{U^2}{E_m} \tag{1}$$

Where U is the impact velocity and E_m is the energy of melting. More precisely, E_m is the specific internal energy of the Rankine-Hugoniot state from which isentropic release ends at 1-bar pressure on the liquidus (Pierazzo et al., 1997).

Preheated rocks may be melted at lower shock pressures than near-surface rocks. This effectively means that the energy E_m is smaller for rocks with high initial temperatures. Our estimates based on ANEOS for dunite show, for example, that dunite with initial temperature of 293 K appears at the solidus line (1373 K) after 103 GPa shock loading. For dunite, preheated to 1293 K (1000K closer to the melting point), only half of this pressure (~50 GPa) is needed to reach solidus temperature after release. However, naturally "preheated" rocks are situated at some depth (50 to 100 km depending on the local geotherm) where the melting temperature is also larger than on the surface. Hence, the increase of the impact melt volume due to hotter source rocks depends strongly on the assumed local thermal gradient. Because of the complex interaction of different factors that may either enhance or reduce shock melt production, the best way to make reliable estimates is direct numerical modeling. We discuss several such models below.

Numerical modeling of a large impact event

To illustrate the potential for melting during the excavation and modification of a large impact crater, we performed numerical simulations that include the effects of decompression melting and thermal gradients in the target. For simplicity, we assume a vertical impact by a dunite asteroid into a dunite target. We use dunite because it is described by a reasonably reliable equation of state (Benz et al., 1989). Target strength properties are derived from triaxial laboratory tests with the addition of thermal softening--initial cohesion and internal friction decrease gradually as the temperature approaches the melting (solidus) temperature. The Simon melting relationship is used to approximate the near-surface peridotite solidus (figure 3 in Herzberg, 1995)

$$T_m(p) = T_m(p{=}0)\ (p/a + 1)^{1/c} \tag{2}$$

where T_m is solidus temperature in K, p is ambient pressure in GPa. Coefficients a = 1.4 GPa, and c = 5 give a good fit to the experimental data below $p \sim 30$ GPa. The strength and internal friction are determined from laboratory triaxial rock tests (Stesky et al., 1974). Thermal softening is treated as a decrease of strength and friction close to the melting point. Following (Ohnaka, 1995) we use the coefficient

$$k_T = tanh\ \{a\ (\square_m/T - 1)\} \tag{3}$$

where the parameter a acts as an activation energy ($E/R\square_m$) that depends on the an empirical constant E, estimated by fitting of friction and failure data for dunite during triaxial compression at high temperature. The limiting strength at any pressure is the product of factor k_T and the strength at the reference (room) temperature.

The acoustic fluidization model of post-impact strength degradation (see for example, Melosh and Ivanov, 1999) was used with parameters linearly scaled from smaller craters (Ivanov and Artemieva, 2002). According to this scaling, the effective viscosity in the fluidized region is approximately 1 GPa sec with vibrations decaying exponentially over 200 sec. To accurately simulate the structure of a very large crater, more model tuning would be required to apply acoustic fluidization properly. However, such precision is not necessary to resolve the overall details of shock and decompression melting in the simulations reported here.

The calculation begins by bringing the target into equilibrium with the pressure imposed by the Earth's gravity field. We use two values for the near-surface geothermal gradient: (i) a "cold" model with 13 K km^{-1} near the surface, and (ii) a "hot" model with 30 K km^{-1} near the surface. In both cases the thermal gradient gradually decreases to ~ 1 K km^{-1} for depth below ~90 to 200 km. The melting (solidus) curve in our model approaches the geotherm near a depth of 200 km ("cold" model) or 60 to 70 km ("hot" model), resembling the asthenospheric position in general plate tectonic models (Turcotte and Schubert, 1982). Comparing this to a thermal model of the oceanic crust (see Fig. 8 in Jeanloz and Morris, 1986), the "hot" case corresponds approximately to 20 Myr old oceanic floor while the "cold" resembles 100 to 150 Myr old oceanic crust. Assuming a steady-state sea floor age of 200 Myr, our "hot" case is typical of about 10% of the oceanic crust at any given time. According to Jeanloz and Morris (1986), partial melt is present underneath the more restricted area of 1 to 10 Myr old sea floor at depth of 50 km. Hence, a hypothetical impact into such an area could potentially excavate some already partially molten material.

A real episode of "impact triggered volcanism" implies the production of partially molten material due to the uplift of hot interior rocks and consequent pressure release. Fig.1 illustrates positions of the solidus and liquidus curves in comparison with the geotherms used in our numerical models. A set of isentrops shows the thermodynamic paths of material undergoing pressure release. The depth of ~55 km is a critical one for the "hot" geotherm used in this work: Below this level the mantle material is hot enough to just melt as it reaches the surface (i.e. the isentrope reaches the solidus at zero depth). For the "cold" geotherm, material must rise from the greater depth of ~125 km to just reach the solidus at the surface. These depths permit us to estimate the minimum size of an impact crater that can trigger real igneous activity. Supposing that the maximum stratigraphic uplift is about 1/10 of the final crater diameter (eg. Melosh and Ivanov, 1999), one finds that the final crater rim diameter must be 500 or 1200 km in "hot" and "cold" cases respectively. Impact basins of these sizes are not known on either the Earth or Venus. On the moon, the few basins in this size range are very old -- more than 3 Gyr.

What about impact heating in addition to the natural geothermal heat of mantle? Naturally, shock waves heat the mantle below the impact point and facilitate melting. However this is the case of a standard "impact melt", where the main heat source is the impact kinetic energy. Hence the term "impact triggered volcanism" should not be used without an explicit statement of which is the more important factor: shock heating or pressure release due to the interior's uplift.

We have constructed a simple numerical model to estimate the relative contribution of these two factors for a reasonably large impact: the projectile is a 20 km diameter asteroid that strikes the Earth at a velocity of 15 km s^{-1}, producing a transient crater with a diameter of about 150 km. After collapse, the final crater is about 250 km in diameter, comparable to, or larger than, any yet recognized on the Earth (Grieve and Therriault, 2000).

The numerical simulation uses the Eulerian mode of the SALE hydrocode (Amsden et al., 1980). These computations reproduce the main phases of a crater growth: transient cavity excavation, gravity-driven collapse with central uplift formation, and a final flat crater. The numerical model resolution (20 cells per projectile radius, implying 0.5x0.5 km cells) is high enough to give a reliable estimate of impact melt production (Pierazzo et al., 1997). The high-resolution (0.5 km) computational zone covers the area of the entire transient crater (~120 km from the point of impact in horizontal direction and 80 km in depth). A gradual increase in cell size beyond the high-resolution zone allows us to put rigid computational zone boundaries at ~600 km in horizontal and vertical directions, where they will not influence the crater computation. Initially horizontal rows of Lagrangian tracer particles display the target material displacement. Figure 2 shows an intermediate stage of transient crater collapse. Tracer layers at the base of the melting zone are shown with thicker lines. Grey shading shows 0-50%, 50-100%, and 100% of melting of rocks at this moment. This figure illustrates that in the "cold" case the greatest depth of melting is ~40 km, much less than the 125 km needed for "impact triggered volcanism". In the "hot case" melted rocks originate from a depth of ~50 km. However, the base of the melted zone never actually reaches the surface and, hence, does not experience the full release to zero pressure.

Figure 3 shows the positions of the tracer rows at the moment of maximum transient crater depth (~40 sec) and later, close to the time of final stage crater formation (720 sec). Below the maximum transient crater depth of 50 km, target layers oscillate up and down with an amplitude of a few km. Layers with an initial depth about equal to the transient cavity maximum depth are involved in the final structural uplift and irreversibly deliver material from depth to near the surface. This material is the most susceptible to pressure release melting. The numerical model yields estimates of the volume of material displaced from one depth interval to another. The model shows that most of melting is due to the "normal" shock compression/adiabatic release cycle, enhanced by a higher initial temperature at a depth. Without the shock heating, pressure-release melting is impossible for this size impact. The total volume of rock suffering incipient or complete melting is estimated as ~30,000 km^3 and 60,000 km^3 for the "cold" and "hot" cases, respectively. The net volume of melt (combining completely melted rocks--above the liquidus--and melt from partially melted rocks--between the solidus and liquidus) is 2 to 3 times less: 15,000 to 20,000 km^3.

An interesting effect occurring at the large scale of our "hot" case is that the transient cavity collapse traps some melted rock in a "neck" around the axis of symmetry (Figure 2, right panel). One could say that an impact into a "hot" target produces an impact "hot spot", although this is far less voluminous than the volcanic hot spots recognized in the geologic record. However, the geometry of this "hot spot" must be verified with 3D numerical modeling, as 2D modeling often produces artifacts close to the symmetry axis.

What about steeper thermal gradients? A gradient much higher than our “hot” value of 30 K/km already implies the presence of “naturally” melted mantle material close to the surface, making an impact simply a melt excavation event, not a volcanic "trigger". A region with such a high thermal gradient is already an igneous center whether an impact occurs or not.

What about much bigger impacts? Consider an impact that creates a transient cavity approximately twice as deep as in our numerical simulation (depth ~100 km). Such an impact is, indeed, big enough to lift hot mantle rocks close to surface. This impact corresponds to a final crater diameter of 400 to 500 km--a very rare event in the current post-heavy bombardment period. Such a huge event is possible, but the lunar cratering record indicates it is highly improbable that an event of this magnitude occurred in the past 3.3 Gyr of terrestrial (and terrestrial planet) geologic history.

Frequency of large-scale impacts vs. magmatic events:

It is easy to show that impacts cannot be responsible for more than a tiny fraction of the magmatic centers on either Earth or Venus. This argument is statistical and compares the rate of formation of large impact craters to the rate at which magmatic centers have appeared throughout geologic time.

The lunar crater chronology, based on the radiometric dating of returned lunar rocks (Stöffler and Ryder, 2001) can be extended to other planets, including the Earth (Ivanov, 2001; Neukum et al., 2001). Terrestrial impact craters larger than 200 km in diameter form at the rate of approximately 4 to 8 per Gyr for the whole of Earth's surface (~510 x 10^6 km^2). Estimates for larger craters give 2 to 4 craters with D > 250 km and 1 to 2 craters with D > 300 km per Gyr. These estimates roughly agree with the impact cratering record on Venus: the largest impact crater Mead has a diameter of 280 km, while the average age of the observed Venusian surface is estimated as 300 to 500 Myr (Schaber et al., 1992).

In contrast, terrestrial records for the most prominent outbreaks of Phanerozoic magmatism (or *Large Igneous Provinces* – LIPs) give approximately 30 events during the last 0.25 Gyr (Coffin and Eldholm, 2001; Eldholm and Coffin, 2000). The total area covered by erupted material is about 25 x 10^6 km^2 or ~5 % of Earth's total surface. Approximately 70% of this "new" material volume was delivered in 6 major events (Coffin and Eldholm, 2001). Hence, in 1 Gyr at least 4 times more LIPs originated (~ 25) than large impacts occurred. On Venus, Magee and Head (2001) counted 208 large lava fields which cover 40 x 10^6 km^2 (~9 % of the total surface). The 81 largest fields cover 25 x 10^6 km^2 or ~ 5.4 % of the planet's surface. Again, the largest, 280 km, crater on Venus combined with 7 craters with D > 100 km cannot account for the much larger number of magmatic events. Thus, most of the observed igneous provinces must have a purely endogenic origin.

Some authors, recognizing the difficulty of provoking volcanic eruptions by impacts on normal terrestrial lithosphere, suggest that impacts may induce a volcanic outburst if the impact happens to strike a pre-existing hot spot (Jones et al., 2001). However, statistical reasoning, coupled with the lifetimes of hot spots, indicates that such a coincidence is extremely improbable.

As described above, prominent igneous areas covered 5 to 10% of either Earth or Venus in 0.5 Gyr. From the lunar/terrestrial crater chronology, 13 impact craters with $D > 100$ km should form on Earth or Venus in the same time. The probability that an impact randomly hits this small area is thus between 0.7 and 1.3 events per 0.5 Ga. However, hot spots are temporary features. According to Turcotte and Schubert (1982, Section 4.16) at terrestrial mid-oceanic ridges the 1300 K isotherm sinks from the surface to 100 km depth in 100 Myr as the oceanic crust spreads and cools. If hotspots last no longer than this, the probability of hitting an active hotspot in 0.5 Gyr declines by a factor of 5, to between 0.14 and 0.26.

Isolated hot plumes such as that assumed to exist under Iceland are also limited in space and time. For example, the Iceland model by Ribe et al. (1995) predicts a plume head at a temperature close to the peridotite solidus that extends to a depth of 50 km and a radius of about 300 km from its axis. The "hot" area around the plume is thus about $3\ 10^5$ km^2. This is 1/1800 of the Earth's surface area. Current estimates suggest that there are either 46 (Crumpler, 1994) or 40 to 100 (Stefanick and Jurdy, 1984) active hot spots of various kinds on the present Earth. If they are all equal in size to the Icelandic hot spot and persist for the unlikely period of 0.5 Gyr, the probability of one being struck by an impact with $D > 100$ km during this time is about 0.4. As we showed above, a 100-km crater is not actually big enough to trigger a volcanic event. We thus conclude that it is unlikely that even one large impact has occurred on an active hot spot in the past 0.5 Gyr. Such an event is possible, but it is not a normal occurrence in Earth history and cannot be invoked to explain typical volcanic eruptions.

Conclusions:

General geophysical considerations, along with a numerical study of the formation of an impact crater 250 to 300 km in diameter, demonstrate that an impact even of this scale can barely provoke an igneous event in normal lithosphere. Although such an impact right on a pre-existing hot spot might produce more melt, "young" hot spots are limited in both space and time. The probability that such a large impact struck a young hot spot any time in post-heavy bombardment history is small. Such a large impact does create an impact "hot spot" in the upper mantle. However, most of the melted rocks are heated by the "normal" shock compression/pressure release cycle, not by decompression melting. A steep thermal gradient enhances the volume of partially melted rock, although the net melt production is similar to the case of a low thermal gradient. Net impact melt production (~10^4 km^3) is 1 to 3 orders of magnitude smaller than the typical estimated volumes of Large Igneous Provinces (10^5 to 10^7 km^3 according to Coffin and Eldholm, 2001).

We conclude that the role of large-scale impacts in "triggering" volcanism has been small, if not negligible, for the last 3 to 3.3 Gyr. Impacts large enough to create a melt-rich hot spot in the crust and upper mantle are rare (1 or 2 craters per Gyr for the whole Earth). To really trigger a volcanic eruption, a large impact must strike a young hot spot, such as a newly arrived plume head, that is already on the verge of erupting. If such an event has ever happened, it cannot be treated as a regular geologic process.

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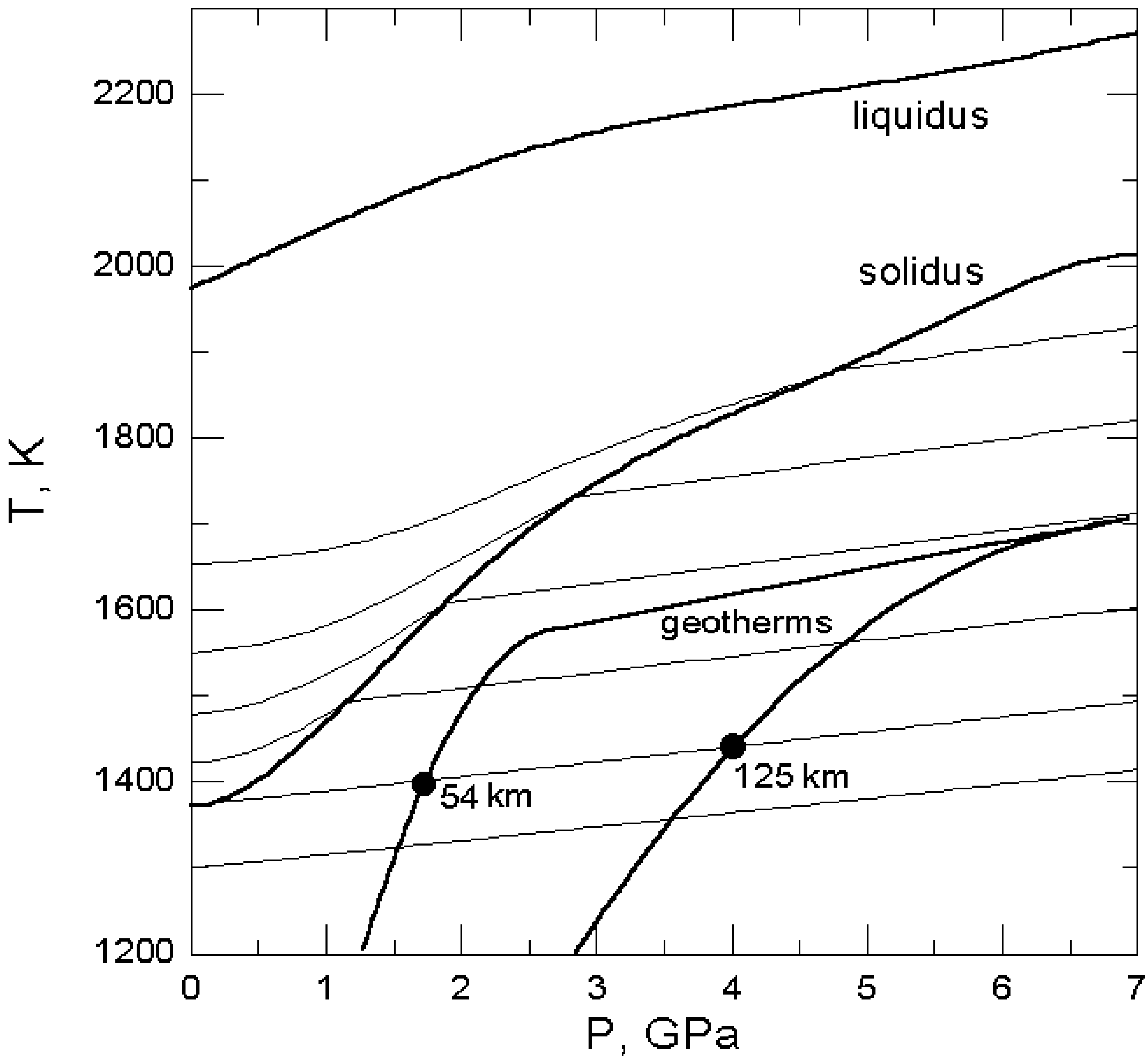

Figure 1. Simplified pressure-temperature diagram (solidus and liquidus curves) for mantle peridotite adapted from McKenzie and Bickle (1988). Adiabats (thin lines) bend where they cross the solidus. This is due to the latent heat of fusion of the melt phase. Black dots show where the adiabat of incipient melting at zero pressure crosses the assumed "hot" and "cold" geotherms. Numbers show the depths of origin of rocks that just reach the solidus after adiabatic release to zero pressure.

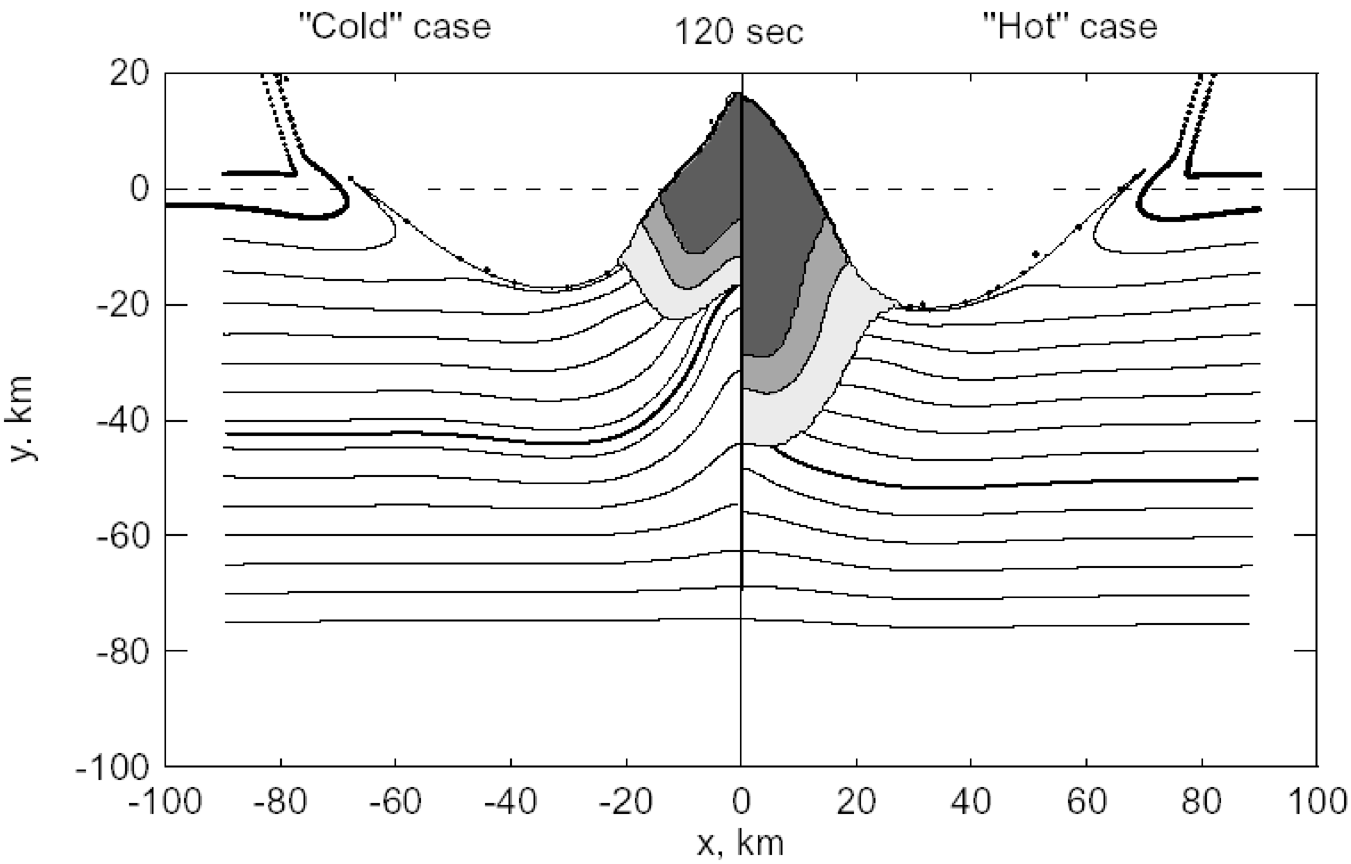

Figure 2. Displacements of initially horizontal layers of Lagrangian tracer particles in the „cold“ (left panel) and „hot“ target (right panel) at an intermediate stage of a transient cavity collapse (120 sec after impact). Thick lines show positions of layers buried initially at the base of a melt zone (~42 and 55 km). In the "hot" case (right panel) the transient crater's maximum depth of 57 km is comparable to the depth of melting. This results in a less effective rebound of the melt zone bottom and to trapping of part of the molten rock below the surface (see Figure 3). Gray shading of the molten zone corresponds (from light to dark gray) to 0 to 50 %, 50 to 100%, and 100% melt content, approximately.

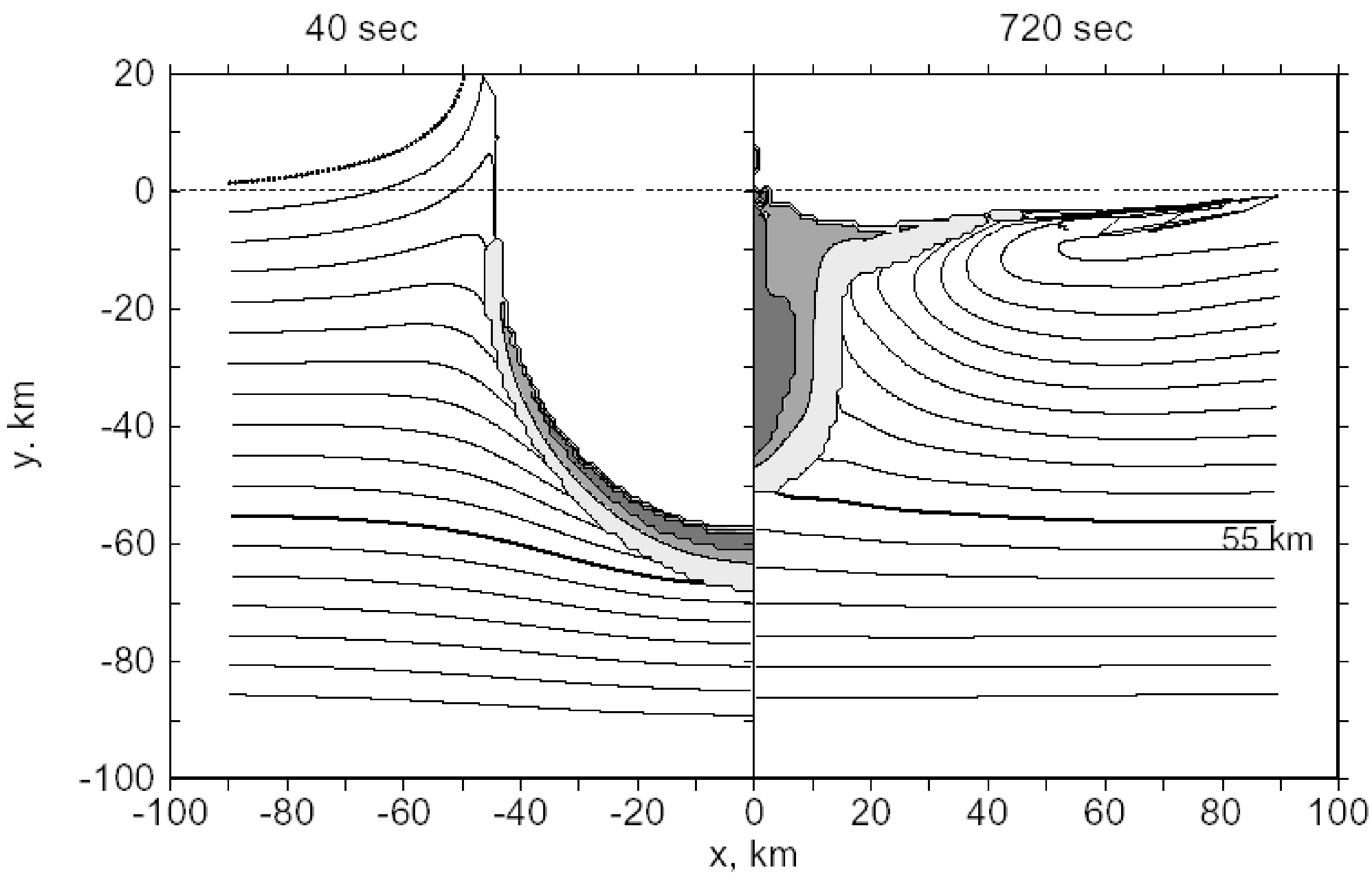

Figure 3. Displacements of initially horizontal layers of Lagrangian tracer particles in the "hot" target near the moment when the transient cavity reaches its maximum depth (left panel) and at a late stage when only small vertical oscillations of the molten material take place. Gray shading corresponds to the same partial melt percentages as in Figure 2. Thicker lines show the positions of layers initially buried close to the base of the melt zone (~55 km). The mantle at this depth is enough hot to just melt upon release even without shock heating (see Figure 1). However, the final position of this level is at a depth of about 5 to 10 km. This implies a pressure release of 10 to 20% of the initial value, which is not enough to reach the solidus at the final pressure. Hence melting of the mantle at this depth is mostly due to the "normal" shock compression/shock pressure release. The impact of a 20 km asteroid may create a deep zone of partially molten mantle material (an "impact hot spot"), but without any significant input from "pressure release by crater formation".