Forum

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

COMMENT

A.Y. Glikson
Research School of Earth Science, Australian National University, Canberra, ACT 0200, Australia

Ivanov and Melosh (2003) define constraints on the likelihood of asteroid impact-triggered volcanic activity. I agree with the authors that to date there is no evidence for an impact-triggered origin of oceanic and continental large igneous provinces, possibly with the exception of the Cretaceous-Tertiary (K-T) boundary Decan basalts (Alt et al., 1988), and that impact reactivation of hot spot loci (Abbott and Isley, 2002) must represent a rare coincidence in Earth history. However, here I point out (1) crustal and petrologic factors that greatly increase the probability of impact-triggered volcanism in geothermally active regions of oceanic basins, and (2) Archean to early Proterozoic field and geochemical evidence of large impact events with likely volcanic consequences.

For a flux of impact craters (D_s > 20 km, where D_s = outer structural diameter) on the order of 4.3–6.3 × 10^{-15} km^{-2} yr^{-1} (Shoemaker and Shoemaker, 1996) on a post–Late Heavy Bombardment (LHB, post–3.8 Ga) Earth occupied by >80% time-integrated oceanic crust (McCulloch and Bennett, 1996), with a cumulative asteroid and crater size/frequency distribution $N_D \propto D\ 10^{-1.8}$ (N_D = number of craters of diameter D), some 360 craters with D >100 km and some 40 craters with D >300 km would form in oceanic basins post-LHB (Glikson, 2001). With a present-day high-geotherm (30 K km^{-1}) oceanic crust occupying ~10% of oceanic crust (Ivanov and Melosh, 2003), assuming higher Archean geothermal gradients and smaller-scale convection cells and plate dimensions (Lambert, 1983), thin (<5 km) oceanic crustal spreading regions overlying shallow asthenosphere (<50 km) can be expected to have occupied >25% of the Archean oceanic basins. In this estimate some ~90 craters with D_s ~ 100 km and ~10 craters with D_s ~ 300 km impacted thermally active oceanic crust post-LHB.

For a 300 km impact structure, using morphometric estimates after Grieve and Pilkington (1996), the stratigraphic uplift $SU = 0.086D_s^{1.03}$ is ~30 km. Alternatively, assuming a D_s/D_t ratio of ~2 (D_t = diameter of transient crater), the transient crater depth, $dt = 0.28D_t^{1.02}$, would be ~45 km. Under geothermal gradients of ~30 K km^{-1} near-solidus asthenosphere would occur at depths of ~40–50 km. The close agreement between the excavation and near-solidus parameters for craters D_s ~ 300 km suggests that impacts on this scale would result in intersection of the peridotite solidus by impact-rebounded asthenosphere, with consequent partial melting. The assumption that catastrophic mantle melting took place during the Archean is supported by the occurrence of peridotitic komatiites (>30% MgO) (Green, 1972, 1981).

Ivanov and Melosh (2003, p. 872) state: "The role of large-scale impacts in triggering volcanism has been small, if not negligible, for the past 3–3.3 b.y." The identification in Archean impact fallout units of high iridium fluxes and large spherule radii (<2 mm) (3.26–3.24 Ga impact fallout deposits in the eastern Transvaal—Lowe et al., 2003 and Kyte et al., 2003; 2.63 Ga and 2.47–2.50 impact fallout units in the Hamersley Basin, Western Australia—Glikson and Vickers, 2003), coupled with the mafic geochemistry and absence of planar deformation features–bearing quartz in these units (Simonson et al., 1998), suggest very large impact events in the contemporaneous oceanic basins. The ca. 3.26–3.24 Ga impact cluster in the Barberton greenstone belt (Lowe et al., 2003) broadly correlates with ca. 3.2 Ga peak impact events on the Moon deduced from Ar-Ar ages of lunar impact spherules (Culler et al., 2000) and with ca. 3.2 Ga Rb-Sr and Ar-Ar ages of volcanics in some of the lunar maria (Glikson, 2001). That some of the Archean impact fallout deposits are accompanied by volcanic tuffs and are succeeded by banded iron formations hints at contemporaneous volcanic and hydrothermal activity.

Finally, I note the title of the article refers to "eruptions close to the crater." Further studies are required to test whether impact-generation and propagation of deep crustal fractures in distal crustal sectors may have taken place. I suggest that the jury is still out regarding the question of the role of large asteroid impacts (D_s >> 100 km) in potentially triggering and/or enhancing mantle fusion events in thermally active oceanic crustal regions (Stothers and Rampino, 1990).

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REPLY

B.A. Ivanov
Institute for Dynamics of Geospheres, Russian Academy of Sciences, Moscow, Russia

H.J. Melosh
Lunar and Planetary Laboratory, University of Arizona, Tucson, Arizona 85721, USA

We appreciate the comment by A.Y. Glikson on our paper (Ivanov and Melosh, 2003a). He raises, we believe, five major points, to which we reply below:

1. Glikson suggests that the eruption of the Deccan traps at the Cretaceous-Tertiary (K-T) boundary is a possible exception to the lack of evidence for impact-triggered volcanism. We disagree. Two recent papers have shown stratigraphically (Bhandari et al., 1995) and by Os isotopes (Ravizza and Peucker-Ehrenbrink, 2003) that the onset of Deccan volcanism preceded the K-T impact by ~500,000 yr. Thus, there is no evidence for an impact-volcanism connection at this boundary.

2. Glikson uses a simple power-law size frequency distribution (SFD) to estimate that 360 craters with diameter $D > 100$ km and 40 craters with $D > 300$ km have formed in the ocean basins since the end of the Late Heavy Bombardment (LHB, which we place at ca. 3, not 3.8 Ga). We believe that his $N_D \sim D^{-1.8}$ power law greatly overestimates the number of large craters expected from extrapolation of the lunar cratering record. The well-known lunar SFD from Neukum and Hartmann (e.g., Neukum et al., 2001) shows a steeper dependence for large lunar craters ($N_D \sim D^{-2.2}$) for lunar craters $D > 64$ km. The lunar SFD (McEwen et al., 1997; Grier et al., 2001) may be translated to Earth (and to other planets, e.g., Ivanov, 2001; Ivanov et al., 1997) with a good fit (within a factor of 2) to the known terrestrial cratering rate for $D \sim 20$–40 km (e.g., Neukum et al., 2001). It predicts 100–200 $D > 100$ km craters and 5–10 $D > 300$ km craters during the past 3 b.y. for the entire Earth, not just the ocean basins.

3. Glikson presents a confusing concatenation of Grieve and Pilkington's formula for structural uplift (*SU*, ~30 km for a 300-km-diameter crater) and the depth of the transient crater (~45 km for a 300-km-diameter crater). This repeats a common misconception (e.g., Jones et al., 1998) that the material beneath an impact crater is lifted the full depth of the transient crater. Instead, material directly beneath the impact point is first pushed down by the excavation flow then rebounds to near, or slightly above, its initial position (O'Keefe and Ahrens, 1993). Even the estimated *SU* (with which our numerical computation agrees well) applies only to the rocks just beneath the crater floor. Deeper-seated rocks are uplifted much less and the corresponding decompression is thus smaller. The pattern of uplift beneath a large impact crater is complex and simple rules do not apply; this is a case where numerical modeling of the detailed flow is essential.

4. Ancient spherule layers found in South Africa and Western Australia are very interesting material to search for the traces of large impacts on Earth. The decade-long controversy about the impact/endogenous nature of spherules (e.g., Reimold et al., 2000; Lowe et al., 2003) has currently shifted in favor of an impact origin (Shukolyukov et al., 2002), at least for the Australian spherules. However, at this date, Australian spherule layers are believed to represent only ~3 impact events during the past ~2.6 b.y. (Simonson et al., 2002). Many more endogenous volcanic episodes seem to have occurred in the same time period, making a causal connection less than compelling.

5. We agree with Glikson that the remote action of giant impacts should be studied in addition to effects near the impact point. We performed a preliminary study of seismically induced excitation of the asthenosphere at the antipodal point (Ivanov and Melosh, 2003b). We concluded that the effects are probably too weak to promote any additional melting. Here again, any kind of trigger effect seems possible only for an area that is already on the verge of erupting. The probability that a giant impact occurs at the antipode of a hot spot is just about as low as that of an impact occurring on the hot spot itself, so that remote triggering of volcanism (magmatism) by giant impacts is a very unusual event compared to the much more frequent nonimpact volcanic events.

Indeed, a few giant impact events certainly occurred on Earth in post-LHB history, and a few might have encountered the fortuitous circumstances necessary for them to enhance magma production. This is an important topic for future study. However the main enigmas of Earth's volcanism must be approached as purely endogenous phenomena.

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