CAN IMPACTS INDUCE VOLCANIC ERUPTIONS? by H. J. Melosh, Lunar and Planetary Lab, University of Arizona, Tucson, AZ 85721. jmelosh@lpl.arizona.edu

Introduction: A theme that runs through hundreds of papers on meteorite impacts is the idea that large impacts can induce volcanic eruptions [one of the first was 1]. 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 lava—filling the nearside impact basins. 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 probe Zond 3 made good photos of the lunar farside that showed 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 1Gyr) than the basins in which they lie [2]. Any presumption of a genetic association of impacts and 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.

Sources of Magma: To examine this problem further, consider the sources of melt on the Earth. Terrestrial magmas originate in one of two settings: Arc magmas are produced by volatile (mainly water) fluxing of the mantle overlying subducting slabs. Pressure-relief melting induced by mantle convection produces the most voluminous lavas on Earth. The first setting has little relevance for meteorite impacts: Impacts do not inject material deep into the target. During an impact the projectile material forms a liner to the growing crater cavity. Even as the crater collapses, these materials are always near the surface and do not get injected deep into the underlying rocks. At large craters such as Sudbury, Manicouagan and Chicxulub the melt sheet lies on top of the brecciated unmelted target rocks. Even a slow impact of a large comet would thus not introduce substantial quantities of water into the crust deep beneath the impact site.

Pressure relief melting by impacts is a more plausible alternative. Terrestrial mid-ocean ridge basalts originate over upwellings where the peridotitic mantle has risen nearly 1000 km from mid- or even lower-mantle sources. Similarly, mantle plumes represent places where especially hot mantle material rises large distances (perhaps even from the core-mantle boundary). Since the slope of the peridotite melting curve is 15-17 K/kbar near the surface, it is clear that uplifts of hundreds of km can easily result in the production of basaltic melt.

Uplift in Craters: The crucial question is, how much uplift do impacts induce in the target rocks beneath them? A number of investigators [e.g. 3] have noted that depth of the transient crater formed by an impact is 1/3 to 1/4 its diameter. Assuming that material is uplifted by a distance equal to the depth of the crater (25 to 30 km for a 100 km diameter crater), it is possible that the ca. 8 kbar pressure change could bring mantle material previously below the solidus more than 100 K closer to melting. Whether melting occurs or not depends on the local geotherm, but if the impacted region is already hot it is possible that melting might occur. However, this scenario overestimates the actual uplift by a factor of 3 or more. Detailed studies of impact craters show that during the impact event material is first pushed downward as the transient cavity opens, then rebounds upward. For most target material this excursion is adiabatic, so the amount of uplift effective for pressure melting is only that remaining after the collapse of the transient crater. Numerous field and numerical studies of crater collapse show that the maximum uplift is only about 1/10 of the transient crater diameter [4]. Furthermore, this maximum uplift occurs only beneath the center of the crater and it dies out rapidly beneath the crater. Even in the 100-km (transient) diameter Chicxulub crater the Moho beneath it is barely disturbed, with less than a few km uplift beneath the center [5]. Under these circumstances pressure relief melting seems very unlikely, even in the largest known terrestrial craters.

Impact Melt: Melt is, of course, 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, then release it adiabatically to low pressure. If the shock is strong enough (typically 50 GPa or more) the released material may be in a molten, or even vapor state. However, since the shock waves are strong only very close to the impact site, the melt remains on the floor of the crater as it opens and later collapses. Shock heating dies off very quickly with increasing distance away from the impact site [6]. The melt thus forms a near-surface sheet throughout the crater formation process. In large craters this melt volume may be considerable. The differential scaling of melt volume (proportional mainly to the projectile volume) and crater volume shows that as crater size increases the ratio of melt volume to crater volume increases [7] until a point is reached when the melt volume equals the crater volume. This probably does not happen for terrestrial craters until the transient crater diameter exceeds about 1000 km. Even for the ca. 200 km diameter Sudbury impact crater, where the entire Sudbury Igneous Complex is now interpreted as a differentiated impact melt sheet [8], the melt layer is only a few km thick although more than 100 km wide. Shortly after the Sudbury impact there may have been many igneous phenomena associated with the melt sheet, but nothing like the extended period of magma extrusion that is characteristic of normal terrestrial volcanism.

Cleopatra: The best possible example of volcanism associated with an impact is the crater Cleopatra on the regional slope of Maxwell Montes, Venus. This enigmatic crater [9] is clearly of impact origin at Magellan resolution. However, it is deeper than most Venusian craters and possesses a channel leading from the crater center to lower-lying plains, where it is associated with a large lava flow. Although an attempt was made to explain its form as impact-induced volcanism (B. Ivanov, unpublished 1992 lecture notes), these models required such high thermal gradients (up to 50 K/km) in the pre-impact surface that it seems more likely that the melt near Cleopatra is similar to the observed outflow sheets [10], modified by the strong regional slope (B. Ivanov, 2000 personal communication).

Volcanic Triggering at a Distance: Although there is no evidence that impacts can induce volcanism near the site of the impact (I know of no single example of such an association on any body in the solar system), some authors have opted for the idea that impacts can induce volcanism at very distant locations. The most widely discussed association is that between the K/T impact and the Deccan Traps in India [3], which are nearly antipodal to the impact site. Although the K/T impact does nearly coincide with the beginning of the Deccan volcanic episode, recent evidence suggests that the volcanism actually pre-dated the impact by a few million years [11]. Another problem with such an association is the sheer amount of energy involved. The volume of the Deccan traps is at least 500,000 km^3 (which does not take into account material now eroded). Using a latent heat of fusion of 330 kJ/kg [12], production of this volume of basalt from a source assumed to be at its melting point would require an energy of 5×10^{23} J. This is about two times larger than the entire kinetic energy of the K/T impactor (assumed to be a 10 km diameter asteroid striking at 20 km/sec)! Clearly, the impact cannot be the direct cause of the basalt but must act as a “trigger” of some kind. What kind of trigger this could be has never been explained satisfactorily. However, one might suppose that a short pulse of intense heat at the antipode might be sufficient to start an eruptive episode of some material nearly ready to erupt anyway. Boslough et al. [13] have suggested that antipodal focusing of seismic waves from the impact might cause large material motions and heat dissipation sufficient to begin an eruption. However, studies of impacts using the Apollo Lunar Seismic Network indicate that the total amount of energy radiated from an impact in the form of seismic waves is only about 10^{-4} of the total impact energy [7]. The simulations of [13] show the seismic energy is concentrated in the asthenosphere at the antipode in a volume several hundred km deep and at least this much in radius. Thus, presuming that the entire seismic energy of 3×10^{19} J is concentrated in this volume, the energy deposited comes out to be about 2500 J/m^3, or about 1 J/kg of the mantle. Using a typical heat capacity of 1 kJ/kg, we get a thermal “pulse” from this mechanism of about 1 milliKelvin! Even using a much higher seismic efficiency of 5% derived from Russian explosion tests (B. Ivanov, 2000 personal communication), the temperature rise is only 0.5 K. This seems utterly inadequate to trigger anything.

Conclusions: The bottom line of this discussion is that there is not a single clear instance of volcanism induced by impacts, either in the near vicinity of an impact or at the antipodes of the planet. This accords well with theoretical expectation from our current understanding of the impact cratering process. The possibility of impact-induced volcanism must thus be regarded with extreme skepticism.

References:

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