LARGE SCALE IMPACTS AND TRIGGERED VOLCANISM. B. A. Ivanov[1,2] and H. J. Melosh[2], [1]Institute for Dynamics of Geospheres, Russian Academy of Science, Leninsky Prospect, 38-1, Moscow, 119334, Russia (baivanov@online.ru, ivanov@lpl.arizona.edu), [2]Lunar and Planetary Laboratory, University of Arizona, Tucson AZ 85721 (jmelosh@lpl.arizona.edu).

Introduction: The idea of impact induced volcanism continues to blossom ([1-3] and other references). However, this appealing idea is seldom supported with an appropriate physical mechanism. The aim of this publication is to critically examine some frequently cited mechanisms of impact energy transformation into a trigger for terrestrial volcanism and magmatism.

Close and Remote Action: In general, we can distinguish two groups of ideas about how volcanism might be provoked by impacts. The first one assumes that the uplift of geotherms accompanying the collapse of a giant impact crater leads to pressure release melting of deep mantle/asthenosphere layers. The second group includes volcanism triggered by seismic waves focused on the antipodal point of Earth.

We have recently reviewed the first ("close") group of "impact trigger volcanism" ideas [4, 5] and found that impacts small enough to be sufficiently frequent (final crater diameter of 250 km) do not uplift rocks from great enough depths to create melt by pressure release melting. Although larger impacts really can do the job, the number of possible impacts creating craters 300 km in diameter and larger is too small to be a common cause of most of the Earth's young hot spots.

Antipodal Seismic Shaking: The remote action of giant impacts is most frequently discussed in terms of seismic wave focusing near the antipodal (to the impact site) point of the Earth. A full analysis is not easy because it requires lengthy computations of the initial phase of impact, shock wave transformation into seismic waves, and the complex pattern of seismic wave propagation through Earth's inhomogeneous mantle and core. In addition, the mechanism(s) by which seismic shaking is converted into magmatic activity through partial melting in the asthenosphere must be treated. Here we present a simple estimate of the maximum ("optimistic") value of seismic energy dissipation in a viscous asthenosphere.

The approximate treatment of the impact energy-shock wave-seismic wave transformation was performed by Boslough et al. (1996). The results are very instructive: a Chicxulub scale impact (10 km diameter asteroid at an impact velocity of 20 km/s) is modeled with a hydrocode until the stress wave amplitude drops below the elastic limit. The elastic waveforms are used to construct the elastic solution, giving synthetic seismograms for the whole Earth, based on a layered 1D PREM model.

The main result is an estimate of the seismic wave amplitudes focused at the antipodal point (at an angular distance of 180 degrees). We use these results to estimate the energy dissipation in the asthenosphere.

We begin with a first-order approximate estimate. Despite relatively large displacements (± 10 m) strains in the antipodal point are small (~50 μstrain).

The maximum stress oscillation, of order 10 MPa (100 bar), is very small in comparison with the crust/mantle strength at a depth of 10 km and deeper. Consequently, an exact elastic solution gives no energy dissipation at all: elastic strains, by definition, are completely reversible.

We use the waveforms published by Boslough et al (their Fig. 3) in a model that estimates the largest possible heating effect. The largest amplitudes (~10 m amplitude at the surface) are concentrated in ~10 oscillations that last ~1000 second. The approximate period of oscillation T is ~100 second, corresponding to a wavelength of 700 km .

For such a long wave we can assume similar parameters in the asthenosphere (~200 km deep). With an average amplitude A of 10 m such oscillations have a typical (average) velocity <v> of material displacement of the order of

$$A/T = \langle v \rangle = 0.1 \text{ m s}^{-1}$$

Strain rate may be estimated as

$$e' \sim du/dx = \langle v \rangle / (c_L \times T) = 1.4\ 10^{-7}$$

where c_L is the longitudinal seismic wave velocity. The maximum strain during one period is

$$e' \times T = 1.4\ 10^{-5} = 14.3\ \mu\text{strain}$$

close to the 50 μstrain estimated by Boslough et al.

For simplicity, we assume that this strain rate operates throughout all 1000 seconds of oscillations.

The widely cited estimate of mantle viscosity of 10^{20} Pa s is valid for slow deformation. The Maxwell time to use this viscosity is

$$\mu/G = 10^{20}/50\ 10^{9} = 2\ 10^{9} \text{ seconds} = 63.5 \text{ years}$$

Assume that the upper asthenosphere is partially molten and therefore has a much smaller viscosity (and the same shear modulus). The maximum heating occurs when the Maxwell time is equal to the oscillation period of 100 seconds;

$$\mu = 5\ 10^{13} \text{ Pa sec}$$

The dissipative rate is $\mu \times e' \times e' = 1.0$ W/m^3. Acting for 1000 seconds this gives an energy dissipation of 1000 J/m^3, corresponding to a heating of about 1 K in rock. However, in a partially melted material this heating is manifested as increased melting, not only higher temperature. Using a melt entropy of 330 J/kg-K, the methods and thermodynamic data of Jull and McKenzie[13] predict a temperature increase of 0.4 K and a partial melt fraction increase of 0.1% for peridotite at 8 GPa (250 km depth) in response to this "optimistic case" heating.

Less "optimistic" estimates use the "quality factor", Q, of the Earth's (e.g. [7]). The value of Q^{-1} measures the amount of strain energy dissipated due to anelasticity per cycle of oscillations. For the PREM model the lowest Q value of 80 is listed for depths of 80 to 220 km. Thus, we can say that during 80 cycles the strain wave energy decreases e (= 2.71828) times. Ten main oscillations in the Boslough et al. model dissipate approximately $1 - e^{-10/80} \sim 13$ % of the energy. Taking wave specific energy as $\sigma \times e \times \rho^{-1}$ (stress×strain/density) one gets $1.5\ 10^{-1}$ J/kg for $\sigma \sim 10$ MPa and $e \sim 50\ 10^{-6}$. For a heat capacity of 1 kJ kg^{-1} this gives only ~0.15 mK assuming that all the wave energy is dissipated as heat (in partially melting rock the temperature and melt fraction are computed by scaling the result cited above to the lower energy input). More exact estimates should take into account that the asthenosphere thickness is less than the characteristic wavelength.

We conclude that the direct thermal action of a Chicxulub scale impact at the antipodal point is negligible. More elaborate mechanisms such as gas bubble growth, triggering of local earthquakes, etc. might be invoked as volcanic "triggers". However, the possible presence of "natural amplifiers" poses a problem for the "triggered volcanism" idea: Much more frequent natural earthquakes (perhaps occurring only close to the "pregnant" hot spot) may forestall the seismic action of rare giant impacts. Several triggering events due to remote earthquake have been analyzed recently [8, 9].

Cratering Rate on Earth: The above concerns with the strength of the thermal triggering effect for Chicxulub size impacts may be countered with the possibility that larger impacts would cause stronger shacking. In response we want to again emphasize just how rare giant impacts are.

Comparative studies of the impact crater records of terrestrial planetary bodies resulted in a more or less reliable understanding of the cratering rate from asteroids of a given diameter (see the recent review [10] and reference list therein). The lunar cratering curve, recalculated to Earth, gives a good estimate of how often large craters form (Fig. 1). The totally independent estimates [11] of the terrestrial crater size frequency distribution made with a "nearest neighbor" technique are in good agreement with the lunar-based model. Fig.1, constructed for the whole Earth's surface, shows the statistically largest crater for a given time period.

Large surface magmatic events occur, on average, every 20 Myr [12]. Hence, even if each (!) D~100 km impact crater provoked an antipodal magmatic event, large scale impacts are not frequent enough to be suspected in volcano ignition.

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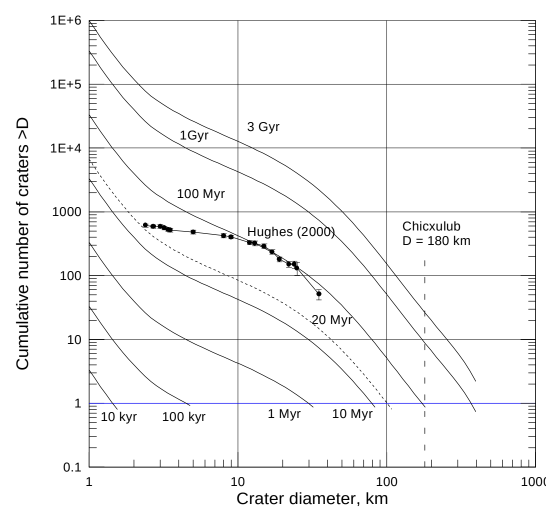

Fig. 1. The cumulative number of impact craters for the whole Earth for a given time period. A Chicxulub scale impact repeats on average every 100 Myr. During 20 Myr (the average time between LIP formation [12]) one 100-km crater, on average, is formed (dashed curve for 20 Myr). Data from [11] are shown for comparison.