

IMPACT VAPORIZATION AND LUNAR ORIGIN, S. M. Rigden and Thomas J. Ahrens, Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, CA 91125

Theories for the origin of the moon fall into three broad categories, each of which suffers from particular difficulties. Capture hypotheses are limited by the low probability of a favorable trajectory for a capture event as well as by dynamic problems (1,2). Theories that consider the earth and moon as evolving in a binary system are appealing dynamically and are not so dependant upon special circumstances as are capture hypotheses. However, they require elaborate mechanisms to explain differences in bulk composition between the earth and the moon (e.g. 3). Fission hypotheses have been popular in one form or another since first proposed by Darwin (4). Formation of the moon subsequent to the segregation of the earth's core was attractive because of the similarity in density and major element chemistry between the moon and the earth's mantle. Recently a number of impact hypotheses of lunar origin that grew from earlier fission hypotheses have been proposed (5,6,7). High velocity impact by planetesimals during the last stages of accretion is envisioned as vaporizing substantial amounts of the earth's mantle. An orbital cloud of dust or vapor subsequently undergoes gravitational collapse to form the moon. A major problem facing such hypotheses involves the transfer of angular momentum required to place the volatilized or ejected material in orbit. Cameron and Ward (5) appeal to glancing collision by a body comparable in size to Mars. Ringwood (7), on the other hand, suggests that vaporized material will be thermally ionized during the impact process and could gather orbital angular momentum from the earth via coupling to the rotating geomagnetic field. What is envisioned is impact by cool planetesimals onto a primitive terrestrial magma ocean. This process might be expected to produce enhanced vaporization of the terrestrial silicates because they are already hot and also more compressible than the cold planetesimals.

We therefore have critically examined the effects of impact into hot and molten relative to cool and solid silicate. Although the appropriate model for the target material for the hypothetical impact generation of lunar material is that of the upper mantle, we have chosen instead a simpler system. For the present our model terrestrial silicate is SiO_2 and the effects of impact at close to earth escape velocities are examined. Using the method of Ahrens and O'Keefe (8), the amount of entropy gain resulting from shocking molten SiO_2 or cold quartz is calculated. The pressure differential between the Hugoniot and an isentrope centered at zero pressure utilized a model Grüneisen parameter (9). To provide a first-order estimate for the Hugoniot of molten SiO_2 , we assume that the same compression at a given pressure is achieved as for normal fused SiO_2 and that the Hugoniot is centered at $\rho_0 = 2.15 \text{ g/cm}^3$ at 2000°K (10). For molten SiO_2 the isentropic equation of state was obtained by assuming a linear relation between particle velocity (u_s) and shock velocity (u_p), $u_s = c + s \cdot u_p$ and the relations $K = c^2/\rho_0$

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and $K' = 4s - 1$ (11). Further, we assume that for very high pressures (on the order of 10 Mbar) the Hugoniot of the initially cold and hot SiO_2 are comparable. The zero pressure density of fictive VSiO_2 liquid was assumed to have $\rho = 3.57 \text{ g/cm}^3$ (9). Pressures attained in the target materials are calculated using the impedance match method at impact speeds between 11 and 16 km/sec. Maximum pressures attained for these impact velocities lie between 2.3 and 3.2 Mbar for quartz and between 2 and 2.9 Mbar for molten SiO_2 . The amount of entropy gain required for melting and vaporization is taken from standard thermodynamic tabulations (12,13). For complete vaporization, an entropy content of $\sim 146 \text{ J/mole-atom}^\circ\text{K}$ is required whereas for incipient vaporization it is $71 \text{ J/mole-atom}^\circ\text{K}$. At impact velocities between 11 and 16 km/sec, between 25 and 35% of the target that reaches peak pressure will be vaporized for quartz and between 30 and 42% will be vaporized in the case of molten SiO_2 . In spite of the higher compressibility of molten SiO_2 and higher initial entropy content, the amount undergoing vaporization under peak pressure is comparable with quartz.

An important constraint to this problem is the spatial distribution of pressure during impact. O'Keefe and Ahrens (4) calculate the mass of vapor and melt generated in terms of the similarity variable, S . ($S = \rho_p / \rho_t (V/C_p)^2$. ρ_p = projectile density, ρ_t = target density, V = impact velocity and C_p = bulk sound-speed). At low values of S ($S < 10$), the mass of vapor was considerably less than projectile mass. Calculation of S for both quartz and molten SiO_2 suggests that once vaporization becomes important, far more vapor is generated by shocking molten SiO_2 than results from quartz. When $S \sim 11$ the mass of vapor released is approximately the same as the mass of the projectile. For quartz this would correspond to an impact velocity of 29 km/sec (See Fig. 1). However at this velocity on the order of ten times the mass of the molten target would be vaporized.

At the velocities considered here (11 - 16 km/sec) which are reasonable estimates for planetesimal velocities upon primordial impact (6) it is clear that little vaporization of either target occurs. This severely limits arguments that seek to place in earth orbit large amounts of vaporized material irregardless of whether it is initially cold or hot and molten. However, at considerably higher velocities ($>20 \text{ km/sec}$) the amount of vapor generated from a molten target is greater than from a cold target. For an impact velocity of $\sim 26 \text{ km/sec}$ a projectile of 0.2 lunar masses could vaporize a mass equivalent to the moon.

References: (1) Goldreich, P. 1966, Rev. Geophys., 4, 411-439. (2) Gerstenkorn, H. 1969, Icarus, 11, 189-207. (3) Ganapathy, R. and E. Anders 1974, Proc. 5th Lunar Sci. Conf. 2, 1181-1206. (4) Darwin, G. H. 1880, Phil. Trans. Roy. Soc. Lond., 171, 713-891. (5) Cameron, A. G. W. and W. R. Ward 1977, Lunar Science Vii, pp. 120-122. (6) Hartmann, W. K. and D. R. Davis 1975, Icarus, 24, 504-515. (7) Ringwood, A. E. 1978, Lunar Science IX, pp. 961-963. (8) Ahrens, T. J. and J. D. O'Keefe 1972, The Moon 4,

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214-249. (9) Lyzenga, G. A. 1980, PhD Diss. Caltech. (10) Aksay, I. A., J. A. Pask and R. F. Davies 1979, *Am. Ceram. Soc.*, **62**, 332-336. (11) Ruoff, A. L. 1967, *J. App. Phys.*, **38**, 4976-4980. (12) Robie, R. A., B. S. Hemingway and J. R. Fisher 1978, U.S.G.S. Bull 1452. (13) JANAF Tables NSRDS-NBS 37, 1971. (14) O'Keefe, J. D. and T. J. Ahrens 1977, *Proc. Lunar Sci. Conf. 8th*, 3357-3374.

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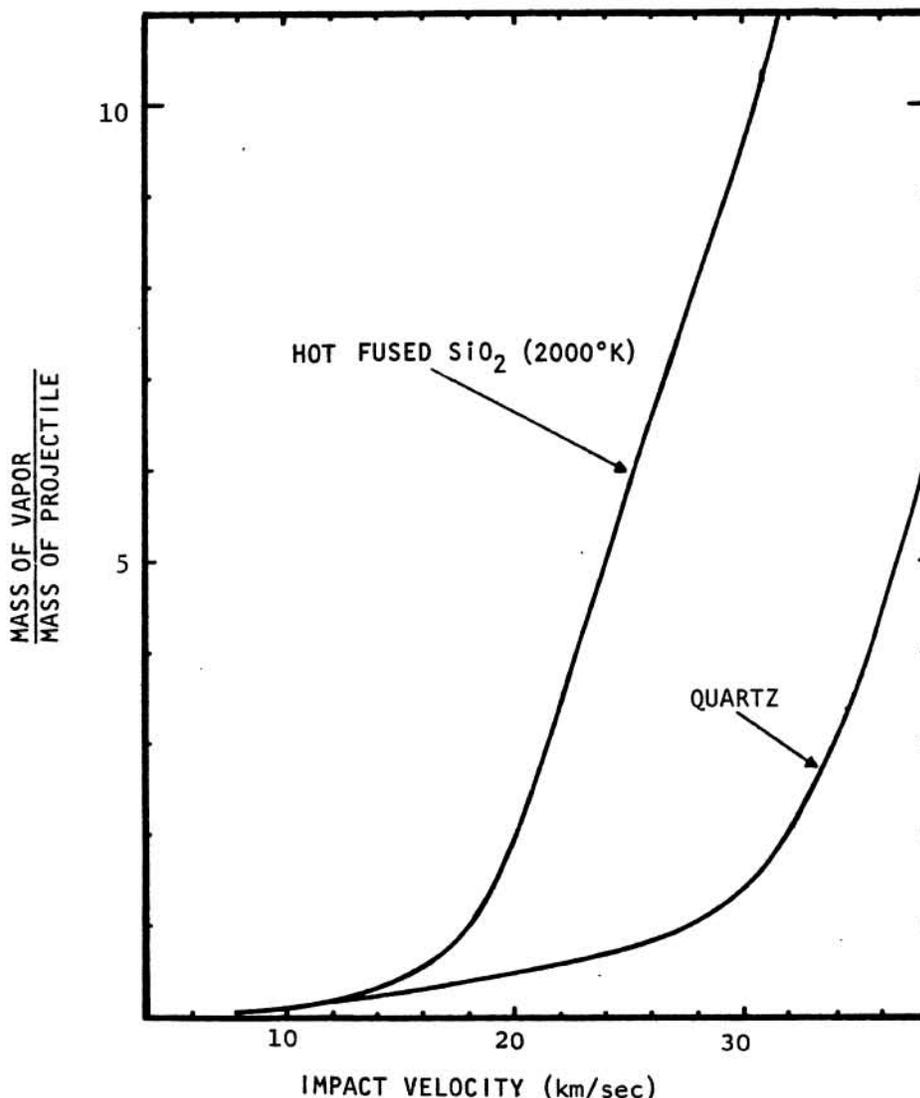


Figure 1: Curves showing varying amount of vapor generated following impact with changing projectile velocity. Substantially more than the mass of the projectile is vaporized only at impact velocities greater than ~20 km/sec for fused SiO₂.