preshock location. Volcanic simulations use impact projectiles on the back surface of preheated targets, producing stress waves that release at the front, unloading rapidly in much the same manner as a decompressing magma chamber.

Our ejecta recovery experiments produced a useful separation of impactites. Material originally below the projectile remained trapped there, embedded in the soft metal of the flyer plate. In contrast, material directly adjacent to the projectile was jetted away from the impact, producing an ejecta cone that was trapped in the foam recovery fixture. The high-speed ejecta showed no signs of shock metamorphism, only intense fracturing, including close intergranular fractures and some planar fracturing in feldspars. These effects are consistent with shock pressures of 5–10 GPa in the most damaged fragments while other fragments showed no significant internal damage. Material trapped in the flyer plate, in contrast, was highly shocked (10–40 GPa), with abundant planar deformation features (PDFs), amorphization, and micrometer-scale fracturing. Thus, we find that a significant component of crater ejecta shows no signs of strong shock; this material comes from the near-surface "interference zone" surrounding the impact site. This phenomenon explains the existence of unshocked meteorites on the Earth of lunar and martian origin. Impact of a large bolide on neighboring planets will produce high-speed, weakly shocked ejecta, which may be trapped by the Earth's gravitational field.

"Frozen crater" experiments show that the interference zone is highly localized; indeed, disaggregation does not extend beyond ~1.5 crater radii. A cone-shaped region extending downward from the impact site is completely disaggregated, including powdered rock (grain size <5 μm) that escaped into the projectile tube. Petrographic analysis of crater ejecta and wall material will be presented.

Finally, study of ejecta from 0.9- and 1.3-GPa simulations of volcanic explosions reveal a complete lack of shock metamorphism. The ejecta shows no evidence of PDFs, amorphization, high-pressure phases, or mosaicism. Instead, all deformation was brittle, with fractures irregular (not planar) and mostly intergranular. The extent of fracturing was remarkable, with the entire sample reduced to fragments of gravel size and smaller. Because the experimental shock stresses match those of the most energetic volcanic explosions, we conclude that explosive volcanism cannot produce shock features such as those seen at the K/T boundary. Instead, these features—similar to those seen in many meteor craters—must be the result of a large meteorite impact.

MELT PRODUCTION IN LARGE-SCALE IMPACT EVENTS: IMPLICATIONS AND OBSERVATIONS AT TERRESTRIAL CRATERS.

Richard A. F. Grieve[1] and Mark J. Cintala[2], [1]Geophysics Division, Geological Survey of Canada, Ottawa, Ontario K1A 0Y3, [2]Code SN4, NASA Johnson Space Center, Houston TX 77058, USA.

The volume of impact melt relative to the volume of the transient cavity increases with the size of the impact event [1–3]. Here, we use the impact of chondrite into granite at 15, 25, and 50 km s^{-1} to model impact-melt volumes at terrestrial craters in crystalline targets and explore the implications for terrestrial craters; details of the model are given elsewhere [4,5].

Figure 1 illustrates the relationships between melt volume and final crater diameter D_R (i.e., after transient-cavity adjustments [5,6]) for observed terrestrial craters in crystalline targets; also included are model curves for the three different impact velocities.

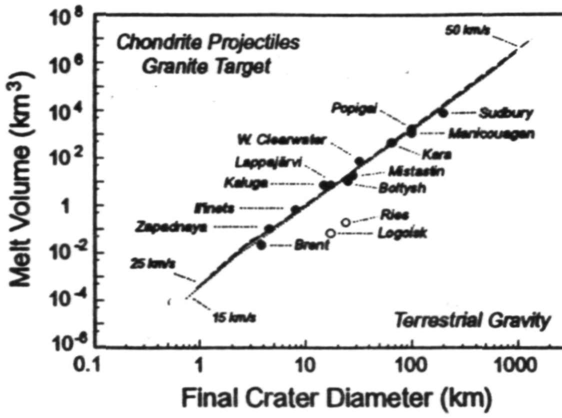

Fig. 1. Impact melt volumes as a function of final crater diameter for observed terrestrial craters and model curves. The slight breaks in slope at 3.1 km are due to application of the modification model of Croft [15].

One implication of the increase in melt volumes with increasing crater size (Fig. 1) is that the depth of melting will also increase (Fig. 2 and Fig. 5 in [6]). This requires that shock effects occurring at the base of the cavity in simple craters and in the uplifted peaks of central structures at complex craters record progressively higher pressures with increasing crater size, up to a maximum of partial melting (~45 GPa). Higher pressures cannot be recorded in the parautochthonous rocks of the cavity floor as they will be represented by impact melt, which will not remain in place. We have estimated maximum recorded pressures from a review of the literature [e.g., 7,8], using such observations as planar features in quartz and feldspar, diaplectic glasses of feldspar and quartz, and partial fusion and vesiculation, as calibrated with estimates of the pressures required for their formation (Table 1). Erosion complicates the picture by removing the surficial (most highly shocked) rocks in uplifted structures, thereby reducing the maximum shock pressures observed. In addition, the range of pressures that can be recorded is limited. Nevertheless, the data define a trend to higher recorded pressures with crater diameter (Table 1), which is consistent with the implications of the model.

TABLE 1. Estimates of maximum recorded shock pressures in the center of craters formed in crystalline targets.

Crater	D_R (km)	Pressure (GPa)	Notes
Rotmistrovka*	2.5	~25	
Brent*	3.8	<25	
Logoisk	20	~30	
Boltysh	25	35-40	
Mistastin	28	>30-35	Eroded
Slate Islands	30	>20	Eroded
W. Clearwater	32	>30-35	Eroded
Araguainha	40	>32	Eroded
Charlevoix	54	>25	Eroded
Kara	60	>35-40	Eroded
Puchezh-Katunki	80	40-45	
Manicouagan	100	40-45	Eroded
Popigai	100	40-45	

Better constrained estimates are shaded.
* Simple craters; all others are complex.

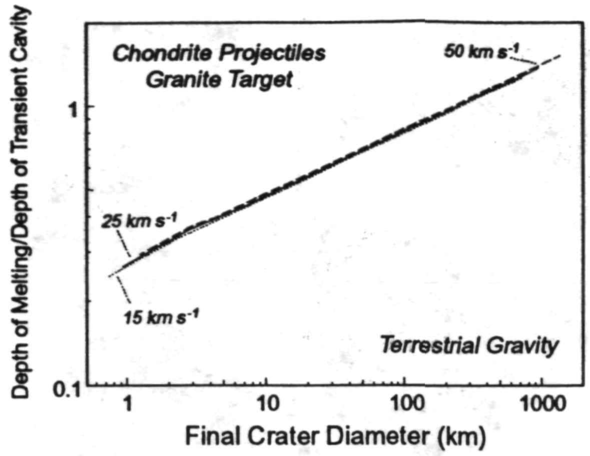

Fig. 2. Depth of melting along the axis of penetration relative to the depth of the transient cavity, plotted as a function of final crater diamater. Note that, regardless of the impact velocity, melting will approach the "base" of the cavity even at relatively small diameters.

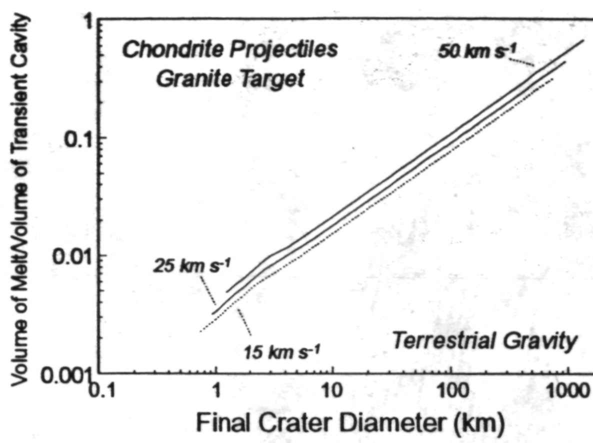

Fig. 3. Volume of melt relative to the volume of the transient cavity as a function of final crater diameter. The melt volume approaches that of the cavity at crater diameters above about 1000 km.

A second implication is that, as the limit of melting intersects the base of the cavity (Fig. 2), central topographic peaks will be modified in appearance and ultimately will not occur. That is, the peak will first develop a central depression, due to the flow of low-strength melted materials, when the melt volume begins to intersect the transient-cavity base. As the melt volume intersects an increasing portion of the transient cavity base, the peak will be replaced upon uplift by a ring. Some of the implications of this mechanism for ring formation and observations on other terrestrial planets is given elsewhere in this volume [9]. The morphology of central structures at complex terrestrial craters was also compiled from the literature [6]; again, erosion is a complicating factor as it can both destroy and create topography. Nevertheless, the general trend is what would be expected with central depressions at values of $D_R \geq$ 40 km, and finally rings appearing at $D_R \geq 100$ km. The latter is equivalent to d_m/d_{tc} values of 0.8–0.9 (Fig. 2), and the diameter at which rings consistently appear in the terrestrial record is also where shock pressures in central-uplift structures record partial melting at $D_R \geq 80$ km (Table 1).

As crater size increases, the volume of impact melt occupies a greater percentage of the volume of the transient cavity (Fig. 3). This implies that less clastic debris is available for incorporation into impact-melt sheets at larger craters. This argument has been used to explain, in part, the general lack of clasts in the bulk of the impact-melt sheet (the Igneous Complex) at Sudbury [10]. There are few detailed studies of clast-content variation in impact-melt rocks. The preserved melt sheets at Mistastin ($D_R = 28$ km) and W. Clearwater ($D_R = 32$ km) are ~100 m thick and have clasts throughout [11,12]. At Manicouagan (D_R = 100 km), however, the melt sheet is essentially free of clasts ~30 m above its base [13]. While this is consistent with the implications of the model, it could result from complete resorption of clasts in the thicker (~200 m preserved thickness) melt sheet at Manicouagan. Ultimately, the volume of melt could equal or exceed the volume of the transient cavity (Fig. 3). In this case (D_R ~ 1000 km) and at larger diameters, the resulting final landform would not resemble a classic crater. We venture that terrestrial basins in the 1000-km size range might have resembled palimpsests, a suggestion made for very large basins on the Moon and Mercury by Croft [1]. Thus, even if preserved, very large and ancient impact structures, such as those suggested to explain meter-thick, areally large spherule beds in the Archean [14] may be unrecognizable in the context of a classic crater form and its impact deposits. At these sizes, terrestrial impact structures might have appeared as low-relief pools of impact melt rocks (10^7 km^3; Fig. 1) with little clastic debris and no obvious associated crater structure. Accompanying subsolidus shock effects would be buried beneath a massive melt sheet and would also tend to anneal out. It would seem, therefore, that such ancient, large structures might not be recognizable as impact features according to common criteria.

References: [1] Croft S. K. (1983) *Proc. LPSC 14th*, in *JGR, 88*, B71. [2] Cintala M. J. and Grieve R. A. F. (1984) *LPSC XV*, 156. [3] Melosh H. J. (1989) *Impact Cratering: A Geologic Process*, Oxford, 245 pp. [4] Cintala M. J. (1992) *JGR, 97*, 947. [5] Cintala M. J. and Grieve R. A. F., this volume. [6] Grieve R. A. F. and Cintala M. J. (1992) *Meteoritics*, submitted. [7] Robertson P. B. (1975) *Bull. GSA, 86*, 1630. [8] Dressler B. (1990) *Tectonophysics, 171*, 229. [9] Cintala M. J. and Grieve R. A. F., this volume. [10] Grieve R. A. F. et al. (1991) *JGR, 96*, 753. [11] Grieve R. A. F. (1975) *Bull. GSA*, 86, 1617. [12] Phinney W. C. et al. (1978) *Proc. LPSC 9th*, 2659. [13] Floran R. J. et al. (1978) *JGR, 83*, 2737. [14] Lowe D. R. et al. (1990) *Science, 245*, 959. [15] Croft S. K. (1985) *Proc. LPSC 15th*, in *JGR, 88*, 828.

VENUSIAN IMPACT BASINS AND CRATERED TERRAINS.

Warren B. Hamilton, Mail Stop 964, U.S. Geological Survey, Denver CO 80225, USA.

The consensus regarding interpretation of Magellan radar imagery assigns Venus a young volcanic surface subjected in many areas to moderate crustal shortening [1–3]. I infer that, on the contrary, ancient densely cratered terrain and large impact basins may be preserved over more than half the planet and that crustal shortening has been much overestimated. I see wind erosion and deposition as far more effective than do others in modifying old structures. Integration with lunar chronology suggests that most of the surface of Venus may be older than 3.0 Ga and much may be older than 3.8 Ga.