for the other observed siderophile element abundances in the boundary clay layer, which must be derived from the impacting projectile.

Because the volcanogenic aerosols are dispersed only by the atmosphere, a wind-modified "halo" of Re and possibly Au should surround the impact site. The other siderophile elements were probably dispersed globally and relatively uniformly on ballistic trajectories.

Impact-induced volcanism may have contributed to extinction mechanisms based on acidification or atmospheric dust. If the volcanism were of sufficient duration and magnitude, an atmospheric dust load could have been maintained for years instead of the months predicted from an impact (Pollack *et al.*, 1983). Lewis *et al.* (1982) suggested that nitrogen oxide acids produced by the shock wave of the projectile caused extinction of calcareous-shelled marine organisms. Acids produced from volcanogenic sulfur emissions could have dominated acidification of ocean surface waters.

The impact of a projectile with a high specific kinetic energy and/or a low siderophile element concentration may produce a mass extinction, but leave a small geochemical signature relative to the K/T event. If the projectile impacts ocean crust, the volcanogenic Re will be the salient siderophile element signature of the extinction-causing impact.

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MAGNESIUM AND CALCIUM ISOTOPES IN HIBONITE-BEARING CAIs

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GR-1, a second corundum-, hibonite-bearing inclusion from the Murchison C2 chondrite (MacPherson *et al.*, 1984), differs from BB-5 (Bar-Matthews *et al.*, 1982) in that it appears to contain two generations of hibonite. Ion microprobe magnesium isotope analyses have been made on the core (C1, C2, Table 1) and rim (R1, R2) hibonites, as well as the corundum. The hibonite gave consistent ^{26}Mg depletions (mean -2.6 ± 2.5) as well as being isotopically light ($F^{Mg} = -5.1 \pm 3.5$). A late formation age, which may be inferred from the lack of ^{26}Mg excess, is difficult to explain given the high formation temperature of this inclusion (MacPherson *et al.*, 1984; Bar-Matthews *et al.*, 1982). An inhomogeneous distribution of ^{26}Al within the solar nebula would appear to be one solution, however, this is less likely if ^{26}Al is continuously introduced into the ISM by novae (Clayton, 1984).

Table 1
Murchison GR-1 Mg isotopes

Hibonite Run	F^{Mg}	δ^{26}Mg	Al/Mg	^{26}Al/^{27}Al*
C1	−3.3	-2.6 ± 5.2	545	$< 7 \times 10^{-7}$
C2	−8.5	-3.8 ± 8.0	1025	$< 6 \times 10^{-7}$
R1	−4.8	-2.3 ± 3.2	306	$< 7 \times 10^{-7}$
R2	−3.8	-3.1 ± 8.3	389	$< 2 \times 10^{-6}$
Corundum				
H1	—	-3 ± 35	21870	$< 2 \times 10^{-7}$
H2	—	$+3.6 \pm 6.6$	36674	$< 4 \times 10^{-8}$
H3	—	$+3 \pm 15$	45560	$< 5 \times 10^{-8}$

F^{Mg} = ‰ mass fractionation relative to terrestrial hibonite standard ± 5‰.
^{26}Al/^{27}Al* = calc. initial ratio.

Calcium isotope mass fractionation was measured for a number of hibonite-bearing inclusions, at low mass resolution, using Ca^{2+} ions at masses 20, 21 and 22. Molecular interfaces appear to be negligible. Calcium in the chemically pure HAL hibonite is isotopically heavy [Table 2, conventional analyses given similar values (Lee *et al.*, 1979)]. Calcium in the Dhajala hibonite is also heavy, as is magnesium (Hinton and Bischoff, 1984). In contrast BB-5 and GR-1 have isotopically light calcium, as well as magnesium (GR-1) and probably formed under different conditions.

Table 2
Ca isotopes in hibonite

Sample	$F^{Ca}(44)$	$F^{Ca}(42)$
Allende		
TS28	0.1 ± 3.0	-0.5 ± 6.9
HAL	7.4 ± 1.6	7.7 ± 3.6
Murchison		
GR-1	-4.9 ± 1.9	-4.4 ± 4.7
BB-5	-4.6 ± 2.2	-4.4 ± 5.0
Dhajala		
HIB1	12.5 ± 1.2	10.5 4.3

$F^{Ca}(42)$ and $F^{Ca}(44)$ are mass fractionation of $^{42}Ca/^{40}Ca$ and $^{44}Ca/^{42}Ca$ ratios relative to terrestrial hibonite standard, in ‰.

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GRAINSIZE EVOLUTION AND DIFFERENTIAL COMMINUTION IN AN EXPERIMENTAL REGOLITH

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The comminution of planetary surfaces by exposure to continuous meteorite bombardment was simulated in the laboratory by impacting the same fragmental gabbro target 200 times. The target weighed 4000 gr and consisted of gabbro-fragments 2-32 mm in size. Stainless steel projectiles ($D = 6.3$ mm; 1.02 g; $Vi = 1.35 \mp .05$ km/s) delivered 9.7×10^9 ergs per shot, or a specific energy of 2.5×10^6 ergs/g target mass. Peak pressures were < 15 GPa and thus insufficient to cause significant melting. As a consequence, these experimental conditions were well suited to address the role of comminution and *in situ* gardening of planetary regoliths.

Mean grainsize continuously decreases with increasing shot number. Initially it decreases linearly with accumulated energy, but at some stage comminution efficiency starts to decrease gradually. Two factors contribute to this decrease: 1) a change in dominating fracture mechanism from intergranular to intragranular as grainsize decreases, and 2) a change in the partitioning of kinetic energy from one initially characterized by collisional fragmentation of coarse boulders and blocks to one dominated by genuine cratering in a fine-grained, homogeneous medium.

The gabbro was coarse-grained (1-5 mm) and had an initial mode of 54% plagioclase, 22% opx, 13% cpx, 5% orthoclase, 5% quartz, and the remainder (1%) being made up of biotite, ilmenite and magnetite. Point counting techniques, aided by the electron microprobe for mineral identification, were performed on a number of comminution products. The modal composition of the 125-250 μm fraction is enriched in plagioclase (65-70%) and depleted in opx + cpx (20-25%). These depletions/enrichments are present in the very first sample removed (after shot 5)