TABLE 1. Selected preliminary INAA data.

| | $Na_2O$ % | FeO % | Sc ppm | Cr ppm | Co ppm | La ppm | Sm ppm | Eu ppm | Hf ppm |
|---|---|---|---|---|---|---|---|---|---|
| *77135* | | | | | | | | | |
| 3 gm | 0.71 | 8.8 | 15.7 | 1325 | 37.7 | 32.6 | 14.9 | 2.04 | 11.8 |
| Av.A–G | 0.70 | 9.2 | 16.2 | 1312 | 51.0 | 33.3 | 15.3 | 2.07 | 12.2 |
| A 2.2 mg | 0.64 | 14.0 | 15.6 | 1343 | 243.5 | 31.0 | 14.3 | 1.97 | 11.7 |
| B 5.8 mg | 0.71 | 8.2 | 16.3 | 1283 | 11.4 | 35.5 | 16.4 | 2.09 | 13.0 |
| C 5.0 mg | 0.65 | 9.2 | 16.3 | 1302 | 14.2 | 32.5 | 15.0 | 2.06 | 11.8 |
| D 3.2 mg | 0.67 | 8.5 | 16.7 | 1388 | 13.1 | 32.7 | 15.0 | 1.89 | 12.3 |
| E 3.5 mg | 0.68 | 8.6 | 17.1 | 1411 | 10.8 | 33.8 | 15.5 | 2.04 | 12.5 |
| F 5.5 mg | 0.78 | 7.6 | 15.5 | 1242 | 9.8 | 32.7 | 15.1 | 2.22 | 11.7 |
| G 4.5 mg | 0.75 | 8.6 | 15.9 | 1216 | 54.1 | 35.0 | 16.1 | 2.20 | 12.3 |
| *76015* | | | | | | | | | |
| 3 gm | 0.67 | 8.4 | 16.0 | 1295 | 17.7 | 32.2 | 14.8 | 2.00 | 11.9 |
| Av. A–E | 0.69 | 8.5 | 16.6 | 1311 | 27.0 | 30.6 | 14.2 | 1.97 | 11.4 |
| A 2.3 mg | 0.69 | 9.9 | 17.1 | 1360 | 91.8 | 30.0 | 13.9 | 1.94 | 10.7 |
| B 2.7 mg | 0.68 | 8.7 | 17.3 | 1355 | 11.0 | 31.5 | 14.7 | 1.99 | 11.5 |
| C 3.9 mg | 0.71 | 8.1 | 16.4 | 1272 | 10.3 | 31.6 | 14.8 | 1.97 | 11.9 |
| D 5.7 mg | 0.67 | 8.3 | 16.5 | 1295 | 11.8 | 31.1 | 14.4 | 1.92 | 11.3 |
| E 4.3 mg | 0.68 | 7.6 | 15.7 | 1271 | 10.0 | 28.8 | 13.2 | 2.02 | 11.6 |

the corresponding 3-g analysis is 10.5%, and most are less than 6% (Co and Fe show greater variation reflecting metal variation). The unweighted averages for the 30 mg for 77135 and the 20 mg 76015 are very little different from the 3-g sample analysis for these elements. Thus these melts were homogenized on a very fine scale, even though there are small but significant differences among the samples at the 3-g scale (larger than among the chiplets from a single rock, as determined here). This gives us great confidence that analyses of particles of impact melt rocks from regolith [e.g., 4] can reliably characterize larger melt volumes.

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## ISOTOPIC AND TRACE-ELEMENT ABUNDANCES IN MURCHISON HIBONITES.

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Two platy hibonite crystals and hibonites from two spinel-hibonite inclusions (an irregular chip and a spherule) from the Murchison meteorite were analyzed for their Ti- and Ca-isotopic compositions and trace-element abundance patterns. All the samples were hand-picked from an HF-HCl resistant residue of a large piece of the Murchison meteorite. The measurements were carried out by a Cameca ims-4f ion microprobe using standard procedures [1].

Both of the platy hibonites, B3 and B5, have group III REE patterns with negative Eu and Yb anomalies and a smooth rollover in the abundances of HREE (Fig. 1). These features are typical of platy Murchison hibonites [1,2]. Excess $^{50}Ti$ is present in both hibonites ($\delta^{50}Ti = 20.9 \pm 2.7‰$ [B3]; $12.3 \pm 5.6‰$ [B5]), and both are devoid of radiogenic $^{26}Mg$ excess (initial $^{26}Al/^{27}Al < 3 \times 10^{-6}$ [B3] and $< 8 \times 10^{-7}$ [B5]). B5 is also devoid of radiogenic excess in $^{41}K$ (initial $^{41}Ca/^{40}Ca < 2 \times 10^{-9}$[3]).

Hibonite in the spinel-hibonite inclusion A4 shows an ultrarefractory REE pattern superimposed on fractionated HREE (Fig. 1). The refractory trace elements Sc, Zr, and Hf are also enriched by factors of >500 compared to CI abundances. This hibonite shows a large enrichment in the neutron-rich isotopes $^{48}Ca$ ($\delta^{48}Ca = 51.0 \pm 5.8‰$) and $^{50}Ti$ ($\delta^{50}Ti = 65.3 \pm 8.0‰$); no detectable excess in radiogenic $^{26}Mg$ and $^{41}K$ was found in it (initial $^{26}Al/^{27}Al < 1 \times 10^{-6}$ and initial $^{41}Ca/^{40}Ca < 6 \times 10^{-9}$ [3]). Hibonite in the spinel-hibonite spherule, A3, has a group II REE pattern superimposed over a smooth fractionated hibonite REE pattern (Fig. 1). However, it does not exhibit the strong depletion in Eu and Yb often seen in CM hibonites with a group II pattern. This hibonite has a detectable excess in $^{50}Ti$ ($\delta^{50}Ti = 12.3 \pm 6.6‰$) and is characterized by a radiogenic $^{26}Mg$ excess with initial $^{26}Al/^{27}Al$ close to the canonical value of $5 \times 10^{-5}$. The very low Ca/K ratio in this hibonite ($\sim 8 \times 10^4$) makes it difficult to identify the possible presence of radiogenic $^{41}K$ excess in it.

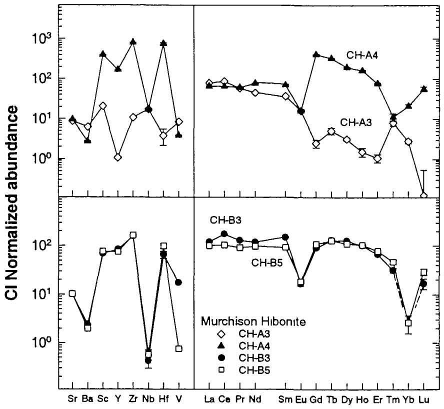

**Fig. 1.** Refractory trace elements and REE abundances in Murchison hibonites.

The isotopic and trace-element data for Murchison hibonites we obtained are in general agreement with earlier data on CM hibonites [1,2,4,5] and allow us to draw some general conclusions about the genesis of these hibonites. An ultrarefractory trace-element pattern has been reported earlier for hibonite in a corundum-hibonite inclusion (GR-1) from Murchison [5] and for a "bulk" sample of a hibonite-bearing Murchison inclusion MH-115 [6]. These hibonites perhaps represent some of the earliest solar system condensates. They are devoid of radiogenic isotopic anomalies (no $^{26}Mg$ and $^{41}K$ excesses), even though they may have a stable isotopic anomaly, particularly an enrichment in the neutron-rich isotopes, $^{48}Ca$ and $^{50}Ti$. A similar trend is also seen for most of the hibonites with a group III REE pattern. In contrast, many hibonites with a group II REE pattern, which are often superimposed by fractionated HREE, have excess radiogenic $^{26}Mg$ and $^{41}K$ but little or no enrichment in $^{50}Ti$. The presence/absence of the short-lived nuclides at the time of formation of the CM hibonites may be explained by postulating either a heterogenous distribution of $^{26}Al$ (and $^{41}Ca$) over a very small spatial scale in the early solar system [4,7] or formation of some of the ultrarefractory and refractory hibonites prior to the time when these short-lived nuclides, injected from an external source, found their way into the formation zone of the refractory phases in the nebula.

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## DEBRIS CLOUDS FROM THE EARLIEST PLANETARY IMPACTS: EVIDENCE FROM THE IRISH METEORITE BOVEDY (L3).

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A role for planetary collisions in the production of chondrules is gradually becoming accepted, even by those who still favor the melting of preformed dust balls in the nebula [1,2]. Because chondrulelike objects were definitely produced by impacts on the Moon, albeit in relatively small numbers, [3] argued that meteoritic chondrules probably resulted from impacts between young planetesimals. The reason for the abundance of chondrules in meteorites and their scarcity on the Moon was rather vaguely attributed by [3] to a different collision regime in the very young solar system compared with that on the Moon. What was different? In my view, the difference was the temperature of the colliding bodies. Planetesimals in the very young solar system were commonly so hot that they were already internally molten, leaving the kinetic energy of impact only a minor factor in producing melt.

Evidence for early molten planetesimals is compelling. The oldest dates obtained from differentiated (i.e., molten) meteorites, based on both long-lived and short-lived isotopes, are within a few million years of the start of the solar system. Mutual collisions between planetesimals are clearly documented by, for example, the mixing of silicate and metal in mesosiderites and IIE irons. Also, formation of the extremely old meteorite, Shallowater, evidently involved the collision of a molten planetesimal [4].

What was the nature and fate of the impact ejecta from these dramatic early collisions? Sanders [5] has argued that the resulting ejecta cloud of molten droplets would be a perfect candidate for the transient, local, volatile-rich, "nebular" environment deemed necessary for the development of chondrules with their retarded-cooling textures.

The Irish chondrite Bovedy (L3) contains a variety of objects that help reconstruct an image of an early planetesimal and its post-impact chondrule-rich debris cloud. Conspicuous multiple (up to five-stage) compound chondrules [6] suggest a close proximity of spray droplets in the cloud. One porphyritic olivine chondrule measures 1.4 × 1 cm; it may be a frozen blob of melt that failed to disperse into spray. In addition to its abundant chondrules, Bovedy contains a variety of interesting objects that are interpreted as fragments of the cooler solid crust or carapace that insulated the molten planetesimal interior. These fragments are of dusty regolith, of earlier generations of now-broken chondrules, and of coarse-grained peridotite. Chondrule caps attached to some angular rock fragments imply mixing of solid and molten components in the cloud. Such mixing of droplets and dust provides, incidentally, a ready explanation for the so-called relict grains in chondrules. Bovedy also contains pieces of fractionated igneous rock represented by a clast of silica pyroxenite [7], and also by a piece of igneous plagioclase (now plagioclase glass) that contains significant excess $^{26}$Mg [mentioned by 1]. One amazing chondrule with immiscibility textures between silica and pyroxenitic glasses may have been derived from the silica pyroxenite rock by impact melting [8]. Some chondrules and rock fragments are coated with progressively finer-grained rims, and they bear a remarkable similarity to terrestrial accretionary lapilli. The coating was presumably added while the objects were lofted in the debris cloud.

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## THE TRAPPING OF NOBLE GASES BY THE IRRADIATION AND WARMING OF INTERSTELLAR ICE ANALOGS.

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**Introduction:** At the low temperatures typical of interstellar dense molecular clouds (T = 10–25 K), most molecules will freeze out onto grains where they will be exposed to ionizing radiation in the form of UV photons and cosmic rays. The energetic processing of these ices leads to the production of more complex and refractory organic residues [1].

We report recent results in our ongoing effort to determine whether these processes could account for the enigmatic chondritic noble gases found in meteorites [2]. Samples are made at NASA Ames by simultaneously depositing, at about 10 K, a series of layers consisting of a combination of polycyclic aromatic hydrocarbons (PAHs), $H_2O$, $CH_3OH$, CO, $NH_3$, and the noble gases He, Ne, Ar, Kr, and Xe. After deposition, each layer is irradiated with UV photons. After the deposition of many layers, the sample is warmed to room temperature and the resulting organic residues are sealed in dry N and shipped to the University of Arizona for noble-gas measurements in a VG5400 mass spectrometer. Originally, the samples were heated overnight to 130°–150°C prior to measurement.

Using these techniques, we previously reported very promising results [3], namely concentrations of trapped Kr and Xe that approached within a factor of about 20 that seen in chondritic meteorites. In this abstract we report work in which we have used isotopically spiked noble gases to further constrain the gas-trapping process.

**Results:** In the latest set of experiments, we have made several changes to our experimental protocol. First, there was some evidence from our earlier work that much of the trapped gas might be escaping during our 130°–150°C heatings prior to measurement. In the latest of experiments, we have limited the preheating to temperatures below 100°C. In order to better track our sample gases, and to assist with the possible increased contribution from absorbed contamination that might result from our lower-temperature prebakes, we replaced our noble-gas samples with mixtures that contained isotopically normal noble gases spiked with additional $^{22}$Ne, $^{40}$Ar, $^{86}$Kr, and $^{136}$Xe.

**Conclusions:** The results continue to be extremely promising, although there are currently some discrepancies between the trapped gases we produce and those seen in meteorites. On the positive side, we are currently trapping the noble gases Kr and Xe at and above the concentrations seen in meteorites. In addition, the measured Kr and Xe elemental and isotopic fractionations in our samples are very similar to those of meteorites. However, our current samples continue to release the majority of their gases at temperatures below 250°C, a lower temperature than is seen in the meteoritic release. We will be carrying out additional experiments to see whether added photolysis of our samples *after* warm up makes them more refractory. In addition, the trapping efficiency of Ar relative to Kr and Xe has yet to be established and we are now carrying out several experiments that are optimized to address this issue.

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## NOBLE-GAS STUDY OF THE NEW LUNAR HIGHLAND METEORITE DAR AL GANI 400.

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**Introduction:** Dar al Gani 400, the third lunar meteorite found outside of Antarctica, was recovered in the Libyan part of the Sahara early this year. Preliminary studies indicate that it is a lunar highland breccia [1].

The abundance and isotopic composition of all noble gases were studied in five different bulk samples. Stepwise heating experiments are in progress to characterize the trapped components. In addition, we apply this technique to gain more information about a contamination with terrestrial noble gases probably caused by abundant carbonate veins that cut through the entire sample.

**Noble Gas Composition:** Dar al Gani 400 is one of few lunar meteorites containing only small amounts of trapped solar gases. The elemental abundance pattern of trapped gases (Fig. 1) reveals that it is depleted in $^{4}$He and $^{20}$Ne relative to the composition of solar particle radiation. It has an abundance pattern comparable to MAC 88104 [2]. Only two lunar meteorites contain less trapped gases, Y 82192 [3] and A 881757 [4], which has virtually no trapped gases. A fit through all available data yields a value for trapped $^{20}$Ne/$^{22}$Ne = 11.2 ± 0.2, which is in agreement with the SEP value of 11.3 ± 0.3.

It is difficult to separate Ar into trapped, radiogenic, and atmospheric components and to determine a retention age because Ar and K are influenced by terrestrial alteration products.