

perfect separation of color and luminance [a problem exacerbated by mixed surrounds (4, 7)], in agreement with the behavior of a major class of cortical cells (10), as well as with psychophysical evidence for color/luminance interactions (11). All of this bears on Masland's dichotomy between the multiplexing (co-coding) and parallel channel approaches. Models that do not filter parvo cells do not account for the properties of cortical cells. Moreover, the use of parvo cells for achromatic form perception without filtering to separate color is inappropriate. As Marr pointed out, the zero crossings of the P cell signal are ambiguous if the color signal is not removed (12). If the color signal is extractable, it makes little sense not to use it.

Vincent A. Billock*

Center for Complex Systems,
Florida Atlantic University,
Boca Raton, FL 33431-0991, USA

*Present address: Armstrong Laboratory, AL/CFHV,
Wright-Patterson Air Force Base, OH 45433, USA. E-mail:
vbillock@falcon.af.mil

REFERENCES AND NOTES

1. D. M. Dacey, B. B. Lee, D. K. Stafford, J. Pokorny, V. C. Smith, *Science* **271**, 656 (1996).
2. R. H. Masland, *ibid.*, p. 616.
3. C. R. Ingling Jr. and E. Martinez-Uriegas, *Vision Res.* **23**, 1495 (1983); C. R. Ingling Jr. and E. Martinez-Uriegas, *ibid.* **25**, 33 (1985).
4. F. A. A. Kingdom and K. T. Mullen, *Spat. Vision* **9**, 191 (1995).
5. C. R. Ingling Jr. and E. Martinez-Uriegas, *J. Opt. Soc. Am.* **73**, 1527 (1983); C. R. Ingling Jr. and B. Tsou, *ibid.* **5**, 1374 (1987); V. A. Billock, in *Colour Vision Deficiencies XII*, B. Drum, Ed. (Kluwer, Dordrecht, 1995), pp. 259-265.
6. P. Lennie, P. W. Haake, D. Williams, in *Computational Models of Visual Perception*, M. S. Landy and J. A. Movshon, Eds. (Massachusetts Institute of Technology Press, Cambridge, 1991) pp. 71-82. They find that for extreme cone ratios and mixed surrounds, color signals from L- and M-center cells are not equivalent. This raises the possibility that a balance of ganglion cell subtypes may be required for normal color vision [V. A. Billock, A. J. Vingrys, P. E. King-Smith, *Visual Neurosci.* **11**, 99].
7. J. B. Derrico and G. Buchsbaum, *J. Visual Com. Image Rep.* **2**, 31 (1991); E. Martinez-Uriegas, in *Visual Science and Engineering*, D. H. Kelly, Ed. (Dekker, New York, 1994), pp. 117-181.
8. V. A. Billock, *Vision Res.* **31**, 33 (1991).
9. ———, *ibid.* **35**, 2359 (1995).
10. L. G. Thorell, R. L. De Valois, D. G. Albrecht, *ibid.* **24**, 751 (1984).
11. K. T. Mullen and F. A. A. Kingdom, in *The Perception of Color*, P. Gouras, Ed. (CRC Press, Boca Raton, 1991), pp. 198-217.
12. D. Marr, *Vision* (Freeman, New York, 1982).

10 April 1996; accepted 4 October 1996

Response: As Masland [in his Perspective (1)] pointed out, our finding in the report (2) that both H1 and H2 horizontal cells in macaques receive additive input from L- and M-cones has implications for understanding the retinal circuitry that underlies spectral opponency. If it is assumed that H1 cells contribute strongly to the formation of

the receptive field surrounds of red-green spectral opponent cells, such opponency cannot arise from cone type-specific connections as originally proposed by Wiesel and Hubel (3), and recently supported by the results of Reid and Shapley (4). An alternative is that H1 cells do not contribute, or contribute only weakly, to the surrounds of red-green cells and that cone type-specific opponency comes about by selective connections between bipolar cells, amacrine cells, and midget ganglion cells. However, as Masland also noted (1), there is growing evidence against such an alternative circuitry (5). Billock points out that in theory, mixed cone surrounds do not pose a serious problem for quantitative models of color opponency and cortical color and luminance coding. We agree with this conclusion, and the formalism offered by Billock is a reasonable one. However, it remains to be shown experimentally that red-green spectral opponent cells do actually have mixed receptive field surrounds. Although, a successful computational model is necessary and important, we would emphasize that the key retinal interneurons subserving red-green opponency, their physiological properties, and precise circuitry are yet to be discovered and described.

Dennis M. Dacey

Department of Biological Structure,
University of Washington,
Seattle, WA 98195, USA

REFERENCES

1. R. H. Masland, *Science* **271**, 616 (1996).
2. D. M. Dacey, B. B. Lee, D. K. Stafford, J. Pokorny, V. C. Smith, *ibid.*, p. 656.
3. T. N. Wiesel and D. H. Hubel, *J. Neurophysiol.* **29**, 1115 (1966).
4. R. C. Reid and R. M. Shapley, *Nature* **356**, 716 (1992).
5. D. J. Calkins and P. Sterling, *ibid.* **381**, 613 (1996).

2 April 1996; accepted 4 October 1996

Response: Neither Dacey [in his report (1)] nor I [in my Perspective (2)] suggested that mixed cone surrounds pose insurmountable problems for cortical color coding. A number of plausible decoding schemes may be proposed, among them the one suggested by Billock.

If the red-green system is multiplexed, though, how about the blue-yellow system, where there is evidence for a dedicated channel (3, 4)? Would the red-green and blue-yellow axes be handled centrally in different ways? From the point of view of the retina, such a dichotomy seems quite possible—the blue-yellow system appears to have evolved independently (5). But it would require somewhat different cortical mechanisms for the two color systems, because the spatial organization of the peripheral receptive fields and the anatomical path to the cortex are different.

Given the power of current techniques (1, 3, 6), the remaining issues about the cellular basis of retinal color coding may be resolved fairly soon. Perhaps the results will raise new questions for experimentation on the striate cortex.

Richard H. Masland

Howard Hughes Medical Institute,
Massachusetts General Hospital,
Boston, MA 02114, USA

REFERENCES

1. D. M. Dacey, B. B. Lee, D. K. Stafford, J. Pokorny, V. C. Smith, *Science* **271**, 656 (1996).
2. R. H. Masland, *ibid.*, p. 616.
3. D. M. Dacey and B. B. Lee, *Nature* **367**, 731 (1994).
4. S. H. Hendry and T. Yoshioka, *Science* **264**, 575 (1994).
5. J. D. Mollon, in *Evolution of the Eye and Visual System*, J. R. Cronly-Dillon and R. L. Gregory, Eds. (CRC Press, Boca Raton, FL, 1991), p. 306.
6. D. J. Calkins and P. Sterling, *Nature* **381**, 613 (1996).

2 October 1996; accepted 4 October 1996

Evaluating the Evidence for Past Life on Mars

David S. McKay *et al.* (1) deserve praise for discovering possible evidence of past Martian life. The identification of indigenous organic compounds in a martian meteorite alone is a breakthrough, reopening the possibility of life after the chill cast by Viking. The characterization of the carbonate globules sets a new standard for study of extraterrestrial materials. However, McKay *et al.* overstate their case by contending that although "[n]one of these [five] observations is in itself conclusive for the existence of past life . . . when . . . considered collectively . . . they are evidence for primitive life on early Mars." An inorganic explanation is at least equally plausible for

four of their five observations.

With regard to polycyclic aromatic hydrocarbons (PAHs), McKay *et al.* (1) note that "in situ chemical aromatization of naturally occurring biological cyclic compounds in early diagenesis can produce a restricted number of PAHs" and suggest that "diagenesis of microorganisms on ALH84001 could produce what we observed—a few specific PAHs—rather than a complex mixture involving alkylated homologs." But aromatization works equally well for abiotic organic matter, which does not even need to be cyclic. Berthelot discovered such aromatization in 1862, producing naphthalene from methane in one

step, and other scientists (2, 3) studied this process a century later, attempting to account for PAHs in carbonaceous chondrites. PAHs with up to at least seven rings formed, in yields ranging up to 8% for pyrene and coronene (2). As in ALH84001, a few specific PAHs dominated, and alkylated homologs were rare.

Such aromatization is nothing more than an approach to a (metastable) equilibrium in a carbon-rich system, under conditions where formation of graphite is kinetically inhibited so that (mainly unsubstituted) PAHs are the next most stable product (4). The equilibrium distribution depends on the proportions of C, H, and O, as well as on pressure (P), and temperature (T). The closer the reaction approaches this metastable equilibrium, the less diverse the products and the less they "remember" the structure of the starting material. High-T processes such as combustion or pyrolysis give low alkyl-PAH/PAH ratios [not high ratios, as contended by McKay *et al.* (1)], whereas low-T diagenesis initially gives ratios greater than 1 (5), which decrease on further equilibration. The "few specific PAHs" from C₁₄ to C₂₂ comprise all homologs in this carbon number range, and thus do not require a selective (biological) source. The higher proportion of alkylated PAHs above mass 300 in ALH84001 may reflect either a different source (1) or slower equilibration of larger molecules, but these alternatives cannot be distinguished without further work.

McKay *et al.* (1) suggest that the spatial association of the PAHs with the carbonate globules favors a biological origin of both. But this association may be caused entirely by inorganic factors. Carbonate, being fine-grained, has a larger specific surface area than do the (anhydrous) silicates, and may even be an inherently better adsorbent. Thus, it may preferentially scavenge dissolved or particulate carbonaceous matter during deposition from water solution, as well as condensable organic compounds from a (volcanic or impact-generated) gas phase injected into rock fractures at a later time. Moreover, some of the phases associated with carbonate (magnetite, clay minerals) are catalysts for Fischer-Tropsch-type reactions which, at $T \geq 150^\circ\text{C}$, convert simple carbon compounds such as CO to more complex ones, progressing from aliphatics to PAHs and kerogen on sustained heating (6). Such in situ synthesis would cause organic matter to become associated with carbonate. Whatever the source and initial chemical form of this organic matter, it would be aromatized during any later reheating events. Indeed, ferric oxide, a likely cause of the orange color of the carbonate globules, is an aromatization catalyst (7).

The two sections on carbonates, magnetite, and iron sulfides (1) are biased, heading straight for biological interpretations without considering inorganic alternatives.

McKay *et al.* (1) describe several features of the carbonate globules that they attribute to microbiological activity (points 1 through 5, below). All of these features can be plausibly explained by inorganic processes. A similar mineral assemblage occurs in C-chondrites, where it has been explained by abiotic aqueous alteration (8–10). By analogy with C-chondrites, suppose that hydrothermal alteration at depth leached Ca^{2+} and Mg^{2+} from rocks, producing an alkaline solution [$\text{pH} = 12.3$ or 10.7 if saturated with $\text{Ca}(\text{OH})_2$ or $\text{Mg}(\text{OH})_2$, respectively]. On contact with the Mars atmosphere, CO_2 would dissolve, gradually lowering the pH and precipitating carbonates.

1) The chemical zoning of carbonates is expected from their solubility products (K_{sp}), which parallel the zoning sequence: $\log K_{\text{sp}}$ for Mn-, Ca-, and Mg-carbonate are -10.65 , -8.30 , and -5.17 , respectively (11). Although FeCO_3 , too, has a small $\log K_{\text{sp}}$ (-10.50), it forms solid solutions with MgCO_3 that have larger K_{sp} (9) and thus precipitate late, as observed (1).

2) The magnetite in the rims requires slight oxidation to Fe(III), and the pyrrhotite requires the presence of some S^{2-} . Both conditions were met in C-chondrites, where a similar mineral assemblage formed under guaranteed abiotic conditions. They contain troilite rather than pyrrhotite, but the stability fields of these two minerals are similar. The magnetite occurs in MgSO_4 rather than carbonate, but the size distribution overlaps the ALH84001 distribution, peaking at 200 nm (12). It would be interesting to check C-chondrite carbonates for magnetite rims.

3) McKay *et al.* (1) state that "the observed dissolution of carbonate would normally require low pH acidic conditions," which would have corroded or dissolved Fe-sulfide and magnetite, contrary to observation. But dissolution of carbonate does not require low pH. On Earth, rainwater saturated with atmospheric CO_2 has a pH as high as 5.3, and yet does a good, if slow, job of dissolving CaCO_3 as $\text{Ca}(\text{HCO}_3)_2$: witness limestone caves and hard water in limestone-rich areas. As soon as the process starts, buffering by HCO_3^- raises the pH to 6 and higher; thus, most of the dissolving is done by water of nearly neutral pH, not "low pH acidic conditions." This would also be true on Mars because pCO_2 and T are similar.

Moreover, because pyrrhotite and magnetite are inside the carbonate, they are

physically protected against attack. And being less soluble in acid than are calcite and magnesite, they also are chemically protected, as long as any carbonate remains to buffer the acid. Manifestly, some carbonate did remain, so either acid, time, or water ran out. These arguments alone are conclusive, but, for what it is worth, natural samples of both minerals often are far more resistant to acids than expected from thermodynamic calculations (13).

4) The elongated greigite (?) particles observed by McKay *et al.* are in a morphological no-man's land, and do not strengthen the case for a biological origin, all the more so because McKay *et al.* (1) do not compare these particles with abiotic greigite from nature and the laboratory. Because the greigite has been identified only on the basis of morphology, the argument is somewhat lacking in rigor.

5) McKay *et al.* (1) note that the "textures of the carbonate globules are similar to bacterially induced carbonate crystal bundle precipitates." After recapitulating observation (i), they conclude: "On the basis of these observations, we interpret that the carbonate globules have a biogenic origin . . ." Again, they reach this conclusion without comparing these textures with those in meteorites or terrestrial rocks. It is not clear how one can prove that A matches B but not C if one has not even looked at C.

McKay *et al.* (1) conclude their paper by listing five lines of evidence and then stating:

None of these observations is in itself conclusive for the existence of past life. Although there are alternative explanations for each of these phenomena taken individually, when they are considered collectively, particularly in view of their spatial association, we conclude that they are evidence for primitive life on early Mars.

For all these observations, an inorganic explanation is at least equally plausible, and, by Occam's Razor, preferable. Consistency arguments alone—weak consistency arguments especially—cannot strengthen, let alone prove, an extraordinary conclusion.

Edward Anders

Enrico Fermi Institute and
Department of Chemistry,
University of Chicago,
5640 South Ellis Avenue,
Chicago, IL 60637, USA

Present address: Hint, Engehaldenstrasse 12, 3004 Bern, Switzerland. E-mail: anders@phim.unibe.ch

REFERENCES AND NOTES

1. D. S. McKay *et al.*, *Science* **273**, 924 (1996).
2. M. H. Studier, R. Hayatsu, E. Anders, paper presented at autumn meeting of the National Academy of

- Science, Seattle, WA, 12 October 1965; *Geochim. Cosmochim. Acta* **32**, 151 (1968).
3. R. V. Eck, E. R. Lippincott, M. O. Dayhoff, Y. T. Pratt, *Science* **153**, 628 (1966); J. Oró and J. Han, *ibid.*, p. 1393; N. Friedman, H. H. Bovee, S. L. Miller, *J. Org. Chem.* **35**, 3230 (1970).
 4. M. O. Dayhoff, E. R. Lippincott, R. V. Eck, *Science* **146**, 1461 (1964); ———, G. Nagarajan, *NASA Rep. SP-3040* (National Aeronautics and Space Administration, Washington, DC, 1967).
 5. M. I. Venkatesan and J. Dahl, *Nature* **338**, 57 (1989).
 6. R. Hayatsu S. Matsuoka, R. G. Scott, M. H. Studier, E. Anders, *Geochim. Cosmochim. Acta* **41**, 1325 (1977); R. Hayatsu and E. Anders, *Topics Curr. Chem.* **99**, 1 (1981), and references cited therein.
 7. A. Galwey, *Geochim. Cosmochim. Acta* **36**, 1115 (1972).
 8. E. R. DuFresne and E. Anders, *ibid.* **26**, 1085 (1962); B. Nagy, W. G. Meinschein, D. J. Hennessy, *Ann. N.Y. Acad. Sci.* **108**, 534 (1963).
 9. K. Boström and K. Fredriksson, *Smithson. Misc. Coll.* **151** (3), 1 (1966).
 10. R. N. Clayton and T. K. Mayeda, *Earth Planet. Sci. Lett.* **67**, 151 (1984); *Meteoritics* **31**, A30 (1996).
 11. Solubility products at 298 K were calculated from data compiled by I. Barin [*Thermochemical Data of Pure Substances* (VCH Verlagsgesellschaft, Weinheim, Germany, 1989)] and other standard sources.
 12. J. F. Kerridge, *Earth Planet. Sci. Lett.* **9**, 299 (1970).
 13. L. Gmelin, *Handbuch der anorganischen Chemie*, vol. **59**, Eisen, Part B (Verlag Chemie, Weinheim, Germany, ed. 8, 1932).

29 August 1996; accepted 15 October 1996

One potential fingerprint of a dynamic organic system would be the fractionation of S isotopes by bacteria. On Earth, the reduction of sulfate ions by anaerobic bacteria such as *Desulfovibrio desulfuricans* results in biogenic fractionation of S isotopes ^{32}S and ^{34}S . These bacteria split O from sulfate ions and excrete H_2S that is enriched in ^{32}S relative to sulfate (1). With the use of this terrestrial analog, we evaluated the possibility of life on Mars (2) by measuring S isotopic systematics in pyrite from ALH84001 (3). Particularly relevant is the temporal and spatial relationship between the sulfides that we analyzed and the carbonate forms described by McKay *et al.* (2). Textural observations (3, 4, 5, 6) indicate that the pyrite in ALH84001 precipitated during the formation of the carbonate that preserved the forms identified by McKay *et al.* (1) as bacteria.

Because of the small size of the pyrite grains (1 μm to 60 μm), ion microprobe techniques were used to measure the $^{34}\text{S}/^{32}\text{S}$ ratio of individual grains. $\delta^{34}\text{S}$ values = $\{[(^{34}\text{S}/^{32}\text{S})_{\text{sample}}/(^{34}\text{S}/^{32}\text{S})_{\text{CDT}}] - 1\} \times 1000$ where CDT is Canyon Diablo troilite, were obtained for three of the largest pyrite grains in ALH84001 subsample 87 (3). We also attempted to analyze smaller pyrite grains in subsample 53. $\delta^{34}\text{S}$ values for the pyrite range from +4.8 to +7.8 ‰ (Fig. 1). Within the stated uncertainties, the pyrites from ALH84001 are demonstrably enriched in ^{34}S relative to Canyon Diablo troilite. In addition, on the basis of the

uncertainties of the ion microprobe analyses, there are real differences among the pyrite grains analyzed in this study.

Compared to the $\delta^{34}\text{S}$ values in sulfides that range from -70 to +70 ‰ on Earth, those for sulfides (primarily troilite) in meteorites show limited mass fractionation (-6.0 ‰ to +6.05 ‰) (Fig. 1). The fractionation seen in meteorites has been attributed to processes that include nebular heterogeneity, low temperature (100°C) reactions between water and elemental sulfur, and oxidation of FeS in an aqueous environment (7). Sulfur isotopic characteristics of hydrothermal sulfides are constrained by many variables, such as the sulfur isotopic characteristics of the hydrothermal fluid T, pH, and f_{O_2} ; consequently, we could not generate unique solutions.

Our results did allow us to propose several conclusions concerning the origin of the sulfides in ALH84001. If the pyrite precipitated at low T (100 to 150°C), reducing conditions and high pH (>9), a $\delta^{34}\text{S}_{\text{fluid}}$ equal to 0 ‰ would precipitate pyrite with $\delta^{34}\text{S}_{\text{pyrite}}$ between +5 and +8 ‰. These conditions are consistent with textural (6) and O-C isotopic systematics for the carbonates (8), but are inconsistent with proposed higher T derived from carbonate geothermometry (4, 5). Alternatively, under more acidic conditions, the $\delta^{34}\text{S}_{\text{fluid}}$ will be equal to that of the pyrite (+5 to +8 ‰). This requires the positive $\delta^{34}\text{S}_{\text{fluid}}$ signature to be produced before or during pyrite deposition. The positive $\delta^{34}\text{S}$ in the fluid may be attributed to a number of different processes.

On the basis of the available data, we favor the idea that the positive $\delta^{34}\text{S}$ in the fluid was produced by either impact processes or low T weathering reactions that stabilized sulfate phases, enriching surface components in ^{34}S . Fractionation of S isotopes through large impact processes has not yet been documented, although S fractionation on the lunar surface has been attributed to micro-meteorite "gardening." Regardless of the mechanism, these components may then be leached and their $\delta^{34}\text{S}$ signature transported to the location of pre-

cipitation. We concluded that either situation is not consistent with the S isotopes being fractionated by biogenic activity.

Although our results do not support the observations of McKay *et al.* (1), our basic assumption could be incorrect, in that either a modern terrestrial analog for biogenic fractionation of S may not be appropriate for ancient (greater than 1 Ga) martian organisms or that the pyrite is not temporally related to the "organic" forms. Microbial sulfate respirators may not have existed on Mars, or the identified bacteria may not have been microbial sulfate respirators. The degree of isotopic fractionation by metabolism of inorganic sulfur by microorganisms on Earth is variable and is controlled by the physiology of the cell and the transfer of sulfur among reservoirs in the cell. The impact of microbial sulfate respirators on the sulfur cycle in a planetary crust is dependent on the existence of a suitable environment. For example, on Earth, dissimilatory sulfate reduction and the resulting S isotopic fractionation is restricted to a limited group of bacteria that live in reducing environments. These few genera of microbial sulfate respirators dominate the terrestrial crustal sulfur cycle. Therefore, although sulfate reducing bacteria existed on Earth by at least 2.7 Ga (2), it is possible that "martian organisms" did not process sulfur at the same rate as the former.

C. K. Shearer

J. J. Papike

Institute of Meteoritics,
Department of Earth and Planetary Sciences,
University of New Mexico,
Albuquerque, NM 87131, USA

REFERENCES

1. M. Schidlowski *et al.*, in *Earth's Earliest Biosphere*, J. W. Schopf, Ed. (Princeton Univ. Press, Princeton, NJ, 1983).
2. D. S. McKay *et al.*, *Science* **273**, 924 (1996).
3. C. K. Shearer *et al.*, *Geochim. Cosmochim. Acta* **60**, 2921 (1996).
4. D. W. Mittlefehldt, *Meteoritics* **29**, 214 (1994).
5. R. P. Harvey and H. Y. McSweeney, *Lunar Planet. Sci.* **26**, 555 (1995).
6. A. H. Treiman, *Meteoritics* **30**, 294 (1995).
7. X. Gao and M. H. Thieme, *Geochim. Cosmochim. Acta* **57**, 3159 (1993); B. A. Paterson *et al.*, *Lunar Planet. Sci.* **25**, 1057 (1994); J. Monster *et al.*, *Geochim. Cosmochim. Acta* **29**, 773 (1965).
8. C. S. Romanek *et al.*, *Nature* **372**, 655 (1994).

27 August 1996; accepted 9 October 1996

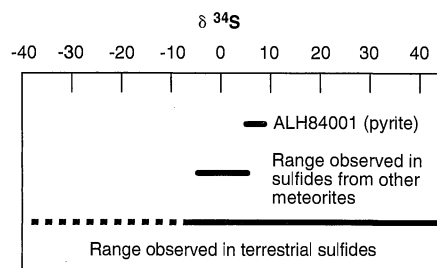


Fig. 1. Comparison of $\delta^{34}\text{S}$ among the pyrite in ALH84001, terrestrial sulfides, and sulfides in other meteorites.

McKay *et al.* (1) suggest that the PAHs they detected in martian meteorite ALH84001 may be related to past life on Mars. Although they make a convincing case that these PAHs are not a result of terrestrial contamination, they do not discuss the possible nonbiological mechanisms by which PAHs may have been introduced into this meteorite on Mars. PAHs of a clearly non-

biological origin have been found in a variety of meteoritic materials (2). One of the first discoveries from the Apollo lunar samples was that 1 to 2% of the lunar soil is meteoritic material, predominantly of the carbonaceous chondrite type (3). Consequently, the Viking GCMS instrument was expected to detect a significant percentage of meteoritic organics in the martian soil. The fact that this instrument did not detect any organic molecules was widely discussed in the aftermath of Viking, and it was generally agreed that the meteoritic organics were destroyed by the highly active photochemistry in the martian soil, driven by the intense solar ultraviolet flux that reaches the surface of Mars in the absence of an ozone layer (4). However, this mechanism would not have operated 3.6 billion years ago in ALH84001 because this rock was well below the surface (even if the thicker martian atmosphere at the time did not absorb much of the ultraviolet flux). PAHs falling onto Mars at this more clement time would have been incorporated into the active groundwater system that deposited the carbonates.

In addition to infall of material on heliocentric orbits, there is evidence for an additional local source of PAHs unique to Mars. Viking Orbiter photographs revealed that Mars has an extraordinary abundance of oblique or "butterfly" impact craters, which are thought to be produced by tidal decay of fragments from a disrupted former satellite of Mars, of which the martian moons Phobos and Deimos are surviving remnants (5). Both moons have dark red reflection spectra characteristic of Class D asteroids (6), thought to be rich in primordial organic polymers formed in the solar nebula (7). Grazing impacts at low orbital velocity would allow for greater survival of projectile material than in conventional heliocentric impacts. Thus, the unique oblique craters of Mars may have contributed a large amount of PAHs, even though they comprise only about 5% of total impacts.

McKay *et al.* state that "Comparison of the mass distribution of PAHs observed in ALH84001 with that of PAHs in other extraterrestrial materials indicates that the closest match is with the CM2 carbonaceous chondrites." This observation suggests a similar history for PAHs in these two classes of meteorites. The CM2 chondrites are composed of primordial nebular condensates that later experienced aqueous alteration in a groundwater system on their parent asteroid. I suggest that the PAHs observed in ALH84001 are primordial organics (partly from the Phobos/Deimos parent body with additional components from primitive asteroids and comets) that underwent groundwater alteration on Mars. This hypothesis could be tested by searching for

PAHs in other carbonate-bearing martian meteorites such as EETA 79001 (7) that apparently were formed much later than the putative ancient life-bearing clement period on Mars.

Jeffrey F. Bell

Hawai'i Institute of Geophysics
and Planetology,
University of Hawai'i at Mānoa,
Honolulu, HI 96822, USA

REFERENCES

1. D. S. McKay *et al.*, *Science* **273**, 924 (1996).
2. S. J. Clemett *et al.*, *Lunar Planet. Sci.* **XXIII**, 233 (1992); S. J. Clemett *et al.*, *Science* **262**, 721 (1993).
3. R. R. Keays *et al.*, *Science* **167**, 490 (1970); J. T. Wasson, W. V. Boynton, C.-L. Chen, *Moon* **13**, 121 (1975).
4. S. F. Chun *et al.*, *Nature* **274**, 875 (1978).
5. P. H. Schultz and A. B. Lutz-Garhan, *Proc. 13th Lunar Planet. Sci. Conf. [J. Geophys. Res.* **87** (suppl.), A84 (1982)].
6. J. F. Bell, F. P. Fanale, D. P. Cruikshank, in *Resources of Near-Earth Space*, J. S. Lewis *et al.*, Eds. (Univ. of Arizona Press, Tucson, AZ, 1993), pp. 887–901; B. Zellner and E. N. Wells, *Lunar Planet. Sci.* **25**, 1541 (1994).
7. J. C. Grady and J. Veverka, *Nature* **283**, 840 (1980).
8. J. C. Wright, M. M. Grady, C. T. Pillinger, *ibid.* **340**, 220 (1989).

1 September 1996; accepted 9 October 1996

Response: Several authors have raised comments concerning the observation of PAHs in ALH84001. We welcome the opportunity to respond.

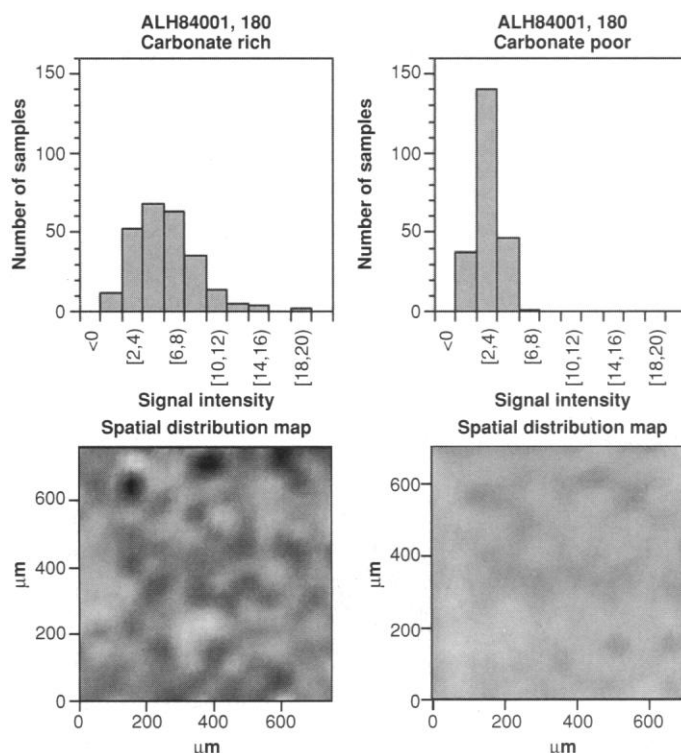
The search for indigenous PAHs on ALH84001 was undertaken to investigate two primary objectives: first, to determine whether an indigenous reduced carbon-bearing phase (organic C) was present within ALH84001; and second, to investigate whether this organic matter showed any spatial correlation with the secondary mineral phases in the meteorite and, in particular, the carbonate globules.

We observed a distribution of PAHs within interior fragments of ALH84001 (1) that appears to be strongly correlated within and about regions rich in carbonate globules (Fig. 1). Extensive contamination and control studies argue for an indigenous origin to these PAHs (2). The total PAH concentration is conservatively estimated by us to average above 1 part per million (ppm) and appears to be made up from two distinct PAH mass groupings: a strong lower-mass envelope, 178 to about 300 atomic mass units (amu) composed of simple three- to six-member fused ring compounds showing little or no alkylation; and a weaker higher-mass envelope, of more than 300 amu to beyond 450 amu, that reveals a complex assemblage of PAHs with extensive alkyl side chains showing varying degrees of dehydrogenation. Before our study, Grady *et al.* (3) determined that ALH84001 contains apparently 600.6 ppm of C of

which 257.1 ppm was loosely identified as organic on the basis of its volatilization and combustion below 450°C. The PAHs represent only a small fraction of this reduced C reservoir.

Although the observation of PAHs in themselves is no proof for either a biological or abiotic origin, failure to observe such a PAH distribution would certainly have argued against any hypothesis of biological activity. Anders correctly asserts that PAH structures are formed whenever organic substances are exposed to high T, and, given sufficient time, aromatization can proceed even at low T (4). The composition of the products of such thermal aromatization depends on the nature of the starting material, the transformation temperatures, and the duration. In general, abiotic synthesis of PAHs from simple organic precursors—for example, CH₄ or HC≡CH—is achieved at medium to high pyrolytic T and results in a PAH distribution that is characterized by a relatively simple mixture of unsubstituted hydrocarbons (4, 5, 6). Such a distribution is observed in the lower PAH mass envelope of ALH84001. A quite different product distribution is observed from the low-temperature aromatization of complex organic (or biological) precursors (or both). In this case the PAH product distribution is generally characterized by a complicated mixture of highly alkylated and partially dehydrogenated PAHs (4, 7). Such a distribution is observed in the higher-mass envelope of ALH84001. The PAH distribution we observed in ALH84001 is remarkable in that it shows features in its distribution consistent with both low-temperature and high-temperature thermal aromatization. Such a distribution could conceivably arise through a sequence of secondary heating events that perturbs or adds to an initial simple abiotic distribution, as suggested by Anders. The PAH distribution we have observed is also equally consistent with the decomposition of biological matter. In the terrestrial environment the geological diagenesis of certain terpenoids and pigments can lead to the observation of a few specific PAHs overlying a weaker, more complex, mass distribution. For example, perylene (C₂₀H₁₂) is found in a variety of marine sediments arising from the decomposition of erythroaphin pigment (7–13); retene (C₁₈H₁₈) is found in forest soils arising from the decomposition of abietic acid (12); fluoranthene (C₁₆H₁₀) and picene (C₂₂H₁₄) are found in mercury ores (14); and various phenanthrene (C₁₄H₁₀) and chrysene (C₁₈H₁₂) derivatives have been associated with French pond mud (15, 16). We regard the observed PAH distribution to be consistent with a few specific PAHs superimposed on a weaker background of more

Fig. 1. Concentrations of PAHs found on martian meteorite ALH84001, calculated by the co-addition of the four principal peaks at 178 amu (phenanthrene), 202 amu (pyrene), 228 amu (chrysene), and 252 amu (perylene). Upper panels show histograms of signal intensities for 40 $\mu\text{m} \times 40 \mu\text{m}$ sample spots on carbonate-rich (left column) and -poor (right column) portions of the meteorite. Lower panels show the corresponding two-dimensional spatial distribution maps of PAHs (co-addition of the four dominant PAH species). The total integrated PAH signal intensity in the carbonate-rich region is about 2.1 times that measured in the carbonate-poor region. The strongest spot intensity, which occurs in the carbonate-rich surface, is about 3.0 times that of the strongest spot intensity in the carbonate-poor surface.



complex aromatic molecules.

The above arguments have been expressed in terms of T of formation, but as Anders asserts, time and T are largely interchangeable for the aromatization process. Thus, it is also possible to discuss the different types of PAHs in terms of differing degrees of equilibration. At the present time, we do not know the formation and transformation of PAHs over geological time scales and under various geological environments.

Because ALH84001 is an igneous rock, the PAHs we observed must have been formed or introduced subsequent to formation of the rock during some later epoch. Because carbonate is not a special material for concentrating PAHs from an aqueous stream, it seems reasonable to assume that the PAHs formed at a time coincident with or subsequent to the time of the carbonate formation. A key question is whether the carbonates formed at low (17) or high (18) T . If the former, then abiotic pathways seem unlikely, whereas if the latter, then conversely an abiotic origin seems the most likely scenario. In this latter case the source of the PAHs may be related to that responsible for PAHs in carbonaceous meteorites, as suggested by Anders as well as by Bell. Anders has also raised the possibility that the carbonates have acquired coatings of carbonaceous matter from heterogeneous catalysis of hydrocarbon gas streams, but again this scenario seems to require high T .

Simoneit and Hites suggest that the PAHs originate from the "thermal degradation of (extraterrestrial) biopolymers," and Requejo and Sassen also point out that the higher mass PAH envelope may arise from "partial thermal decomposition of carbonaceous polymers." Indeed, our PAH distribution resembles closely what is found in studies we have made of terrestrial and meteoritic kerogens. Terrestrial kerogens have a biological origin, whereas meteoritic ones do not. Because of the igneous nature of the primary minerals in ALH84001, it is a puzzle for us to understand the origin of an abiotic kerogen-like precursor in and about the carbonate globules but not associated with the primary mineral phase.

Simon J. Clemett

Richard N. Zare

Department of Chemistry,
Stanford University,
Stanford, CA 94305-5080, USA

REFERENCES AND NOTES

1. D. S. McKay *et al.*, *Science* **273**, 924 (1996).
2. In early contamination studies reported by K. L. Thomas *et al.* [*Lunar Planet. Sci.* **26**, 1409 (1995)], we found low-mass PAHs in the Antarctic diogenite ALH77256, but subsequent studies showed that these PAHs were artifacts of the curation and sampling-handling procedures. Later contamination studies on petrologic types 5 and 6 ordinary chondrites (H, L, LL), which are known to contain little to no reduced carbon [S. J. Clemett, thesis, Stanford University (1996)], were free of similar PAH distributions.
3. M. M. Grady, I. P. Wright, C. Douglas, C. T. Pillinger,

Meteoritics **29**, 469 (1994).

4. M. Blumer, *Sci. Am.* **234**, 35 (1976).
5. J. Oro and J. Han, *Science* **153**, 1393 (1966).
6. N. Friedmann, H. Bovee, S. L. Miller, *J. Org. Chem.* **35**, 3230 (1970).
7. S. G. Wakeham, C. Schaffner, W. Giger, *Geochim. Cosmochim. Acta* **44**, 415 (1980).
8. W. L. Grady and J. R. Grady, *ibid.* **31** (1967).
9. F. S. Brown, M. J. Baedeker, A. Nissenbaum, I. R. Kaplan, *ibid.* **36**, 1185 (1972).
10. A. Aizenshtat, *ibid.* **37**, 559 (1973).
11. S. G. Wakeham, *Environ. Sci. Technol.* **11**, 272 (1977).
12. R. E. LaFlamme and R. A. Hites, *Geochim. Cosmochim. Acta* **42**, 289 (1978).
13. S. G. Wakeham, C. Schaffner, W. Giger, J. J. Boon, J. W. D. Leeuw, *ibid.* **43**, 1141 (1979).
14. T. A. Geissman, K. Y. Sim, J. Murdoch, *Experientia* **23**, 793 (1967).
15. C. Spyckerelle, A. C. Greiner, P. Albrecht, G. Ourisson, *J. Chem. Res. (M)* (suppl.), 330 (1977).
16. ———, *ibid.*, p. 332.
17. C. S. Romanek *et al.*, *Nature* **372**, 655 (1995).
18. R. P. Harvey and J. H. Y. McSween, *ibid.* **382**, 49 (1996).

9 October 1996; revised 4 November 1996

Response: We are pleased to respond to Anders's comments. ALH84001 is an igneous rock, a differentiate from the early martian crust. The carbonates in it are secondary minerals shown to have been added to the rock in cracks and pore space while it was on Mars (1–3). Most of the features we discuss (3) were found in small areas about 100 micrometers in length. We know of no inorganic example from either meteorites or terrestrial rocks that display a similar set of features as secondary alteration of an igneous rock.

Most of the questions raised by Anders concerning the PAHs are discussed by Clemett and Zare in the response above. We know of no experimental data showing that carbonates preferentially absorb PAHs compared to silicates. Some of the other minerals, particularly in the brecciated zone, are as fine-grained as the carbonates (1, 2), so it would be difficult to attribute a preferential absorption to the grain size alone. Also, we did not describe any clay minerals associated with the carbonates in our article (3). However, we have observed two nanometer-sized regions of phyllosilicates in ALH84001 adjacent to orthopyroxene (opx) and in one carbonate-opx interface. The abundance of clay minerals in ALH84001 appears to be much less than 1% by volume.

Anders (points 1 through 5) correctly points out that the same minerals that occur within carbonate globules in ALH84001 are present within certain carbonaceous chondrites that have undergone low T aqueous alteration. However, this comparison does not fit well in detail, although basic processes may apply to both. In ALH84001 Mg-Fe carbonate, magnetite, and pyrrhotite coexist at the micrometer-

scale and are clearly intergrown and formed in the same event (that is, they constitute a co-genetic assemblage). The same cannot be said for these minerals in CI chondrites, which are regolith breccias (4–7) that have undergone multiple episodes of aqueous alteration. The various lithologies in CI chondrites are a collection of materials that record different *P*, *T*, fluid compositions over time. These lithologies do not contain the same micrometer-scale intergrowth of carbonate-magnetite-pyrrhotite that we observed in ALH84001. In addition, hydrated minerals comprise the bulk of the matrix of most CIs (8). Carbonate-sulfate-magnetite, carbonate-sulfide intergrowths, and elemental S have been observed in CI chondrites (5). Carbonate-sulfate-magnetite intergrowths are intimate mixtures of irregularly shaped carbonates, rounded magnetites, and tiny sulfate laths predominantly occurring as void filling (5). The magnetites are larger than any observed in ALH84001. We did not observe any sulfate in the ALH84001 carbonate rims. CI carbonate-sulfide intergrowths are sulfide laths coherently intergrown with irregularly shaped carbonates. These sulfides are pyrrhotites and are at least an order of magnitude larger than those in ALH84001 (5, 9).

The modal abundance of carbonates varies within and among CIs (5, 10). CI carbonates have three compositions: dolomite, breunnerite, and calcite (or aragonite) (5, 11). The complex, oscillatory chemical zoning and rim development observed in ALH84001 has not been reported in any carbonates within any carbonaceous chondrites. Dolomite is the dominant phase in CI chondrites, whereas dolomite is not the dominant phase in carbonate globules in ALH84001 (1, 2, 12). While a few zoned grains have been described (4–6), none show the zonation and complex rim formation shown by the ALH84100 carbonates.

The solubility of solid carbonate in any fluid (Anders's point 1) is a function of many factors, including solution ionic strength, temperature, cation concentration, carbonate alkalinity (or $p\text{CO}_2$), pH, grain size, and solid-solution chemistry (13). We agree that the trend in cation chemistry observed for carbonate globules can be produced by inorganic processes. The lack of information about the precipitating fluid precludes us from attempting to predict trends in carbonate chemistry using conventional aqueous models that consider ion interactions. This problem is exacerbated by oscillatory chemical zoning in some globules (14) and the distinctions in chemistry observed between carbonate globules and other carbonate-rich regions of the meteorite (12). We have previously pointed out that ALH84001 (15) and other mar-

tian meteorites contain secondary carbonates (16). Euhedral magnetites found in CI chondrites can be intimately intergrown with carbonates (Anders's point 2); magnetite often occurs as a boundary between two distinct carbonate generations (5). However, magnetites in CI chondrites are much larger than (and have a different morphology from) those in ALH84001 (17). Furthermore, the magnetite in CIs is nearly all in the size range for multidomain magnetite; most of the magnetite that we describe in the carbonate rims of ALH84001 is in the single domain size range for cuboidal- and arrowhead-shaped magnetite, the predominant size range for biogenic magnetosomes.

We agree with Anders (point 3) that carbonate can dissolve at nearly neutral pH under the right conditions. However, regardless of the numerical value of the pH, the observation is that carbonates formed and then subsequently were partially dissolved. This partial dissolution was not accompanied by any detectable corrosion of the magnetites or sulfides, but would require a change in eH—pH conditions or cation concentrations. It may be that the nanophase sulfides and magnetite were physically protected by the carbonate, but the carbonate is very fine-grained in much of the area containing these minerals allowing possible fluid access along grain boundaries [figure 4 in our report (3)]. Alternatively, as suggested by Anders, these minerals may dissolve slowly under conditions that dissolve carbonate rapidly, or they may be more resistant to dissolution than predicted. Again, because the composition of the fluid is not well known, we were unable to propose specific pH conditions. We are dealing with nonequilibrium assemblages and conditions which change over time—as indicated, for example, by the strong zoning of the carbonate chemistry over short distances, as well as the oscillatory nature of some of the zoning. The main difference between CI chondrite and ALH84001 carbonates is that the latter show pronounced zoning at the micrometer scale and contain nanophase magnetites and sulfides. Although a formation mechanism has yet to be determined for the carbonate globules in ALH84001, the circular to ovoid outlines of the carbonate are consistent with biologically mediated reactions (18, 19). Terrestrial carbonates also commonly display circular and concentric morphologies (18). Such features have not been found in carbonaceous chondrites, but are common in terrestrial carbonate concretions and early diagenetic cements (18–20). It is possible to propose any number of inorganic models for the production of the carbonate, the magnetite, and the iron sulfides, but the models cannot be simple equilibrium models and

must include changing conditions and kinetic effects. Whether such models are more plausible than biogenic models is a matter of judgement.

Although we were not able to positively identify greigite based on electron diffraction (Anders's point 4), the chemistry and morphology of these sulfides particles suggests that they are greigite (3). Greigite is commonly formed in bacteria in the same manner as magnetite (21). In terrestrial sediments, much fine-grained natural greigite is interpreted as either biogenically formed or as a diagenetic reaction product between H_2S and Fe ions in solution (22). The source of H_2S is usually believed to be the product of bacterial decomposition of organic matter coupled with reduction of sulfate in solution, but inorganic sources of H_2S can also be envisioned. No greigite or other similar monosulfides have been reported in carbonaceous chondrites or in other meteorites or in lunar samples.

The morphology of the possible fossil forms in ALH84001 (Anders's point 5) is certainly not definitive, and more data are needed. We have not seen similar structures in any lunar, meteoric, or cosmic dust sample. However, ancient microfossils in terrestrial rocks are also sometimes identified primarily on the basis of morphology and size distribution, although the indigenous nature of the microfossils must be established (23). As Anders states, the presence of fossilized indigenous microorganisms would be the strongest evidence of all and would render the discussion of some of the other lines of evidence somewhat irrelevant.

Finally, Anders states that all of our data are compatible with a biogenic origin, but his main objection seems to be that we have not given equal time to inorganic mechanisms and that we have come down too hard on the biogenic side. We hope that the discussion above will clarify our view of some of the proposed inorganic mechanisms and explain why we did not favor them.

David S. McKay

Mail Code SN,

NASA Johnson Space Center,

Houston, TX 77058, USA

Kathie L. Thomas-Keprta

C23, Lockheed Martin,

2400 NASA Road,

Houston, TX 77058, USA

Christopher S. Romanek

Drawer E,

Savannah River

Ecology Laboratory,

University of Georgia,

Aiken, SC 29802, USA

Everett K. Gibson Jr.

Hhjatollah Vali

Mail Code SN4,

NASA Johnson Space Center

REFERENCES AND NOTES

1. A. H. Treiman, *Meteoritics* **30**, 294 (1995).
2. D. W. Mittlefehldt, *ibid.* **29**, 214 (1994).
3. D. S. McKay et al., *Science* **273**, 924 (1996).
4. M. Endress et al., *Nature* **379**, 701 (1996).
5. M. Endress and A. Bischoff, *Geochim. Cosmochim. Acta* **60**, 489 (1996).
6. K. Boström and K. Fredriksson, *Smithson. Misc. Collect.* **151**, No. 3 (1966).
7. L. R. Riciputi et al., *Geochim. Cosmochim. Acta* **58**, 1343 (1994).
8. K. Tomeoka and P. R. Buseck, *ibid.* **52**, 1627 (1988).
9. J. F. Kerridge et al., *Earth Planet. Sci. Lett.* **43**, 359 (1979).
10. M. Farina et al., *Nature* **343**, 256 (1990).
11. K. Fredriksson and J. F. Kerridge, *Meteoritics* **23**, 35 (1988).
12. R. Harvey and H. Y. McSween Jr., *Nature* **382**, 49 (1996).
13. R. M. Garrels and C. L. Christ, *Solutions, Minerals and Equilibria* (Jones and Bartlett, Boston, 1965); C. L. Christ and P. D. Hostetler, *Am. J. Sci.* **286**, 439 (1970); J. W. Morse and F. T. Mackenzie, *Geochemistry of Sedimentary Carbonate* (Elsevier, New York, 1990); T. L. Wood and R. M. Garrels, *Geochim. Cosmochim. Acta* **56**, 3031 (1992).
14. C. S. Romanek et al., *Nature* **372**, 655 (1994).
15. C. S. Romanek et al., *Meteoritics* **30**, 567 (1995).
16. C. A. Johnson and M. Prinz, *Geochim. Cosmochim. Acta* **57**, 2843 (1993).
17. J. F. Kerridge, *Earth Planet. Sci. Lett.* **9**, 299 (1970); J. Jedwab, *Icarus* **15**, 319 (1971).
18. C. Buczynski and H. S. Chafetz, *J. Sed. Petrol.* **61**, 226 (1991).
19. M. L. Coleman and R. Raiswell, *Am. J. Sci.* **295**, 282 (1995).
20. C. D. Curtis, M. L. Coleman, L. G. Love, *Geochim. Cosmochim. Acta* **50**, 2321 (1986); P. S. Mozley and W. W. Carothers, *J. Sed. Petrol.* **62**, 681 (1992).
21. D. A. Bazylinski, *Am. Soc. Meteor. News* **61**, 337 (1995).
22. A. P. Roberts and G. M. Turner, *Earth Planet. Sci. Lett.* **115**, 257 (1993).
23. J. W. Schopf and C. Klein, Eds., *The Proterozoic Biosphere* (Cambridge Univ. Press, New York, 1992); M. D. Muir, P. R. Grant, G. M. Bliss, W. L. Diver, D. O. Hall, *Chemical Evolution of the Early Precambrian* (Academic Press, New York, in press), pp. 155-170.

27 November 1996; accepted 2 December 1996

Response: Shearer and Papike have contributed important new data regarding the S isotopic systematics in the magmatic phase pyrite within the martian meteorite ALH84001 (1). However, it appears that Shearer and Papike have not analyzed the S phases that may be directly related to the proposed biogenic activity we proposed (2). We noted the presence of intergrowths of magnetite and pyrrhotite phases within carbonate rims. The pyrrhotite phases within the "Oreo" rims of the carbonate globules are between 20 and 100 nm in size (2). These sulfide phases are two to three orders of magnitude smaller than the pyrite phases described (stated to be between 15 and 40 μm) and analyzed by Shearer and Papike (1). We did not report results on the late-stage pyrites that had been observed earlier by Mittlefehldt (3), Treiman (4), and Harvey and McSween (5). Although these late-stage pyrites may have formed during a hydrothermal episode, they are not necessarily contemporaneous with the carbonate globules. It is our understanding that the Cs^+ beam used in the ion microprobe analysis (1) was at least 8 μm or larger in size. Ion microprobe analysis of carbonate phases for S isotopes will be a challenging analytical problem. During such analysis, masses of 32 ($^{16}\text{O}^{16}\text{O}^+$), 34 ($^{16}\text{O}^{18}\text{O}^+$) and 36 ($^{18}\text{O}^{18}\text{O}^+$) amu will be produced from the O within the carbonates. Unless efficient energy filtering can be used to separate them, these ions may interfere with masses 32 ($^{32}\text{S}^+$) and 34 ($^{34}\text{S}^+$) amu normally used when sulfide phases are analyzed for their S isoto-

pic compositions. The nanophase sulfide phases found (2) within the carbonates were smaller in size and intimately associated with the host carbonates in contrast to the late-stage pyrites associated with the silicate host phases analyzed by Shearer and Papike (1). Sulfur isotope measurements on nanophase sulfides would add important constraints to the interpretations of the genesis of the nanophase sulfides.

Everett K. Gibson Jr.

SN4, Planetary Sciences Branch,
NASA Johnson Space Center,
Houston, TX 77058, USA

David S. McKay

SN, Earth Sciences and Solar System
Exploration Division,
NASA Johnson Space Center

Kathie Thomas-Keprta

Mail Code C23,

Lockheed Martin,

2400 NASA Road 1,

Houston, TX 77058, USA

Christopher S. Romanek

Drawer E,

Savannah River Ecology Laboratory,

University of Georgia,

Aiken, SC 29802, USA

REFERENCES

1. C. K. Shearer and J. J. Papike, *Geochim. Cosmochim. Acta* **60**, 2921 (1996).
2. D. S. McKay et al., *Science* **273**, 924 (1996).
3. D. Mittlefehldt, *Meteoritics* **29**, 214 (1994).
4. A. Treiman, *ibid.* **30**, 294 (1995).
5. R. Harvey and H. McSween, *Lunar Planet. Sci.* **26**, 555 (1995).

6 September 1996; accepted 8 November 1996

Location. Location. New Location...

Discover SCIENCE On-line at our new location and take advantage of these great features...

- Fully searchable database of abstracts and news summaries for current & past SCIENCE issues
- Interactive projects, special features and additional data found in the Beyond the Printed Page section
- Classified Advertising & Electronic Marketplace

NEW URL

Tap into the sequence below and see SCIENCE On-line for yourself.

<http://www.sciencemag.org>

SCIENCE