Papers presented at the

Seventh Annual
Summer Intern Conference

August 15, 1991
Houston, Texas

1991 Summer Intern Program for Undergraduates
Lunar and Planetary Institute

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             NASA Johnson Space Center

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The Summer Undergraduate Intern Program is in its fifteenth year at the Lunar and Planetary Institute. Each year students are chosen from a large field of applicants to participate in a ten-week research project under the guidance of Lunar and Planetary Institute or NASA Johnson Space Center staff scientists. The purpose of the program is to provide bright undergraduates with interests in lunar and planetary science with opportunities to experience full-time research in the field of their choice and to stimulate research activities in the local scientific community.
# SEVENTH ANNUAL SUMMER INTERN CONFERENCE PROGRAM

**Lunar and Planetary Institute**  
Houston, Texas, U.S.A.  

**August 15, 1991**

Chairman:  Stephen Clifford  
Michael Zolensky

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*Overview on the Preliminary Results Concerning the Preservation State of the Marquez Dome, Leon County, Texas* |
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*Metallic Iron in Lunar Sample 79002,2030* |
| 9:50  | Erin C. Hatch  Advisor: Deborah Domingue  
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| 10:30 | COFFEE BREAK

1. **Jose M. Aguirre**  Advisor: Virgil L. Sharpton  
*Overview on the Preliminary Results Concerning the Preservation State of the Marquez Dome, Leon County, Texas*

2. **Irene Atonenko**  Advisor: David McKay  
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Magnetic Fields in the Solar Nebula

ADJOURN

* Papers presented at a special seminar August 8, 1991
Overview on the Preliminary Results Concerning the Preservation State of the Marquez Dome, Leon County, Texas; Jose M. Aguirre, University of California, Berkeley, CA 94720, and Virgil L. Sharpton, Lunar and Planetary Institute, Houston, TX 77058

Introduction. The Marque Dome, located in the Northwest corner of Leon County, Texas, is a Paleocene impact structure (58 ± 1 Ma.), occurring in unconsolidated sediments of the Tertiary Calvert Bluff Formation. Continuous Calvert Bluff deposition buried the structure and aided in its preservation, but recent east trending erosion has unearthed the central uplift, 2-3 km wide, and possibly exposed other crater units. Seismic data collected for the region indicates the crater structure to be approximately 22 km in diameter. The exposed peak provides the opportunity to study characteristics of a large impact structure formed in unconsolidated sediments. Penetrating into the structure's western flank is the Navasota River, where a spherule layer of possible ejecta material is located (1). Initiated as a result to understanding metamorphic implications of impacts and their structures on poorly consolidated sediments, this work addresses the preservation state of the exposed central peak and other crater units of the Marquez Dome.

Approach/Techniques. Locations on the central peak and along the banks of the Navasota River were examined for indications of impact melt and aerodynamically sculptured fall back material. Size fractionation of coarse and fine grain samples were attained by rinsing the field samples with deionized water through an 841 and 125 micron sieve, respectively. Discarding of the mud and clay in the samples and with the use of a binocular microscope, a percent abundance of the mineralogy and visual screening for evidence of shock metamorphism was conducted. Features observed of most interest were translucent and opaque beads consistent with aerodynamic impact droplets that were analyzed under the scanning electron microscope (SEM) and energy dispersive x-ray spectra were acquired.

Results. Table 1 illustrates results from binocular microscope clastic observations for the central peak (L2 - L4) and the Navasota River (L6 - L6c) samples. Since the central peak is Cretaceous in age, the presence of Cretaceous bioclastic material was no surprise. However, since no evidence exists in samples L2 and L4 of any impact droplets, it was unusual to find three translucent beads less than 0.5 mm in diameter in sample L3. Samples from the Navasota River (L6 Fresh - L6c) contained coarse grain spherules, ranging up to 3 cm in diameter, and beads with diameters as fine grain as 125 microns. The river bank further yielded clayey, Fe rich material, orange to brown in tone, possibly derived from
impact melt, varying in size from coarse to fine grain. Further analysis is required on this material to assess its source. Sample MD 6B01 contained several small spherules similar in topography to melt droplets. A selected amount of beads and grains were then analyzed using the scanning electron microscope. The scanning electron microscope indicated that the spherules were organically formed. Others were well-rounded clastic grains unrelated to the impact event. However, one bead (Fig. 1), 125 microns in diameter, resulted in being extremely rounded. Energy dispersive x-ray analysis showed the bead to be compositionally homogeneous and having a very smooth surface similar to air fall glass. The spectra analysis (Fig. 2) indicates the chemical composition of the bead to be approximately 50% silica and is similar to pyroxene (personal communication, Sue Wentworth). Because of the high concentration of Si, the glass bead does not fit with restraints for a typical tektite. Although, an impact on unconsolidated material is not a typical impact and the bead is consistent with melt droplets. However, the size of the bead (~125 microns) was consistent with the size of the sieve, allowing for the possibility of laboratory contamination. The very crystalline surface appearance is also hard to explain for a spherule from the Paleocene.

**Conclusion.** The work presented was unable to isolate conclusive evidence suggesting the samples collected at the central peak and the banks of the Navasota River could be interpreted as impact ejecta. The presence of the spherule layer and possible melt fragments on the banks of the river suggest that the river possibly penetrates into the crater walls. Before an impact origin to these products is confirmed, additional chemical analysis of the samples as well as petrographic analysis for shocked minerals, are required. The occurrence of spherule MD 6B01 is intriguing, but further chemical work is required to assess what, if any, connection it has with the Marque Dome impact event. These studies are currently under way.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Bioclasts</th>
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<th>Beads</th>
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<tbody>
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Table 1. Material abundance percentage was listed where though to be relevant. Tr denotes trace amounts of the material and c and f indicate coarse or fine grain size relative to the sieves.
Acknowledgements: The author would like to thank the following for their logistic, administrative, and moral support: the entire staff of the Lunar and Planetary Institute, scientists at the petrographic and SEM laboratory at NASA/JSC, and the co-author for sharing with me his insight of a fascinating field.

METALLIC IRON IN LUNAR SAMPLE 79002,2030
I. Antonenko L.P.I. Intern at NASA Johnson Space Center Houston, Texas 77058

Introduction: The soil samples that were returned from the Apollo missions contain a number of glass balls, some of which are hollow [1]. Many of these balls, or spherules, have metallic iron present on their surfaces, in the form of mounds, trains and lacy structures [2,3]. Similar iron formations have been produced in simulated lunar glasses by processes of reduction [4], thereby implying that the lunar iron could have been formed in this manner [5].

Results: Sample 79002,2030, a hollow glass spherule from Apollo 17 soil, was broken for analysis and studied, using the SEM and X-ray diffraction. The most common feature of this ball is a lacy structure of connected iron grains (Fig. 1a), found both on the interior and the exterior surfaces of the grain, though much more common on the exterior. Also prevalent are more massive formations (Fig. 1b), where large grains of iron are surrounded and joined by interstitial iron sulfide (troilite). The upper right corner of Fig. 1a shows a region where the lacy and massive iron are merging. Quite rare, but nevertheless occurring on both interior and exterior surfaces, are mounds of iron-troilite (Fig. 1c & d).

Thin sections of this sample were also prepared and studied. In the interior, iron is present mostly in small grains, giving the glass a speckled appearance (Fig. 2a). Larger configurations are also found (Fig. 2b), but these are very rare. Generally, large concentrations of iron occur on the surface or in vesicles (Fig. 2c). Troilite is also found in the interior as small speckles, difficult to distinguish, by appearance, from metallic iron.

Discussion: Many of the features identified, at first glance appear to be splatter or splash formations. Their presence on both the interior and exterior surfaces, as well as on vesicle walls, contradicts that possibility.

Vapour condensation has also been proposed as a method of formation [3]. If iron-sulfide vapour were trapped in the molten glass, the cooling gas would naturally deposit crystals on vesicle walls. There are two problems with this hypothesis. The smallest vesicle depicted in Fig. 2c could not possibly contain enough iron in a vapour phase to deposit the amount of iron observed. Also, this method would not account for the speckles of iron and iron-troilite seen in the interior of the glass. Vapour condensation could not have played a major role in lunar iron formation.

Laboratory reduction of simulated lunar glass, however, has produced features analogous to those seen on the lunar sample. The lacy iron of Fig. 1a bears a remarkable resemblance to some simulated iron [4 fig. 15, 25]. Features similar to the massive iron of Fig. 1b are also found, where it is shown that iron sulfide forms "waists" around iron grains [4 fig. 20,24]. Grains comparable in size to Fig 1c have also formed [4 fig. 21]. Therefore, reduction processes are capable of producing all of the surface features found on lunar spherules.

It is interesting to note that the troilite only appears in the massive structures and the mounds. This could be explained by considering the Fe-FeS phase diagram (Fig. 3). Since it is known that reduction processes can occur at temperatures well below the
solidus of metallic iron [1], it can be seen that at such low temperatures, even a small amount of sulphur will result in an FeS liquid. This iron-sulfide liquid, being much more mobile than the solid iron, will migrate more easily, consolidating into larger masses. As these large liquid masses reach the eutectic, the iron and iron-sulfide phases will separate, producing the iron grains with troilite "waists" that are seen (Fig. 1b). The pure iron grains, being solid, are less mobile and thus form the lacy iron (Fig. 1a). Thus, the various features can result from a single homogenous source, namely the glass itself, if the reduction model is to be applied.
METALLIC IRON IN LUNAR SAMPLE 79002,2030: I. Antonenko

Fig. 2. (a)SEM photograph of thin section of glass showing speckles of iron and iron sulfide. Also shown is a plagioclase grain that has partially dissolved into the glass. (b)SEM photograph of a larger congregation of iron in the interior of the glass. (c)SEM view showing iron in vesicles and along the edge of the thin section, which coincides with the exterior of the spherule.

Fig. 3. Schematic temperature-composition phase diagram for Fe-S system at \( P = 1 \) atm [6 fig. 1880].

Acknowledgements: Supported by the LPI internship 1991. I would like to thank D.S. McKay and C.C. Allen for invaluable advice and direction, also R.P. Bernhard, S.J. Wentworth and K.L. Thomas for assistance with equipment, samples and general support.

Investigations of the structural and textural properties of the optically active portion of planetary regoliths begin with disk-integrated photometry. Because both earth-based and remotely sensed photometric data are available, albeit limited, for many objects in the solar system, disk-integrated photometry is an ideal starting place for studying the regolith of many planetary objects. An initial and general characterization of the regolith is an important basis for further study. The structure and texture of the Galilean satellites Europa and Callisto have been analyzed with disk-integrated photometry. It is evident from these analyses that the leading edge of Europa is brighter than the trailing edge. For Callisto, on the other hand, the leading edge is darker than the trailing edge. It may actually be the case that the relative difference in reflectance between the leading and trailing hemispheres changes with increased radial distance from Jupiter. That is, as the radial distance from Jupiter increases the relative brightness of the leading edge versus trailing edge decreases. Ganymede, the third Galilean satellite, is intermediate in distance from Jupiter compared to the aforementioned moons and also exhibits an intermediate dichotomy between leading and trailing hemisphere brightness. See Figs. 1 and 2.

Because Ganymede has been observed from earth by several different groups and by the Voyager I and Voyager II spacecraft, there is sufficient data to construct solar phase curves. The earth-based telescopic observations are reliable and accurate sources of photometric data, because the cameras and filters are easily checked and calibrated. The telescopic data also covers phase angles less than two degrees which provides data for the well known opposition effect. However, the telescopic data for Ganymede only extends to phase angles of about twelve degrees, which severely retards a comprehensive study of its regolith. The Voyager data for Ganymede, on the other hand, spans phase angles from about four to over 120 degrees. This data, however, is unreliable without a suitable standard for comparison like the telescopic data. Because the Voyager filters and cameras are unpredictably effected by radiation and dust during their trip through the solar system, they must be calibrated with an accurate standard. For these reasons the combination of earth-based telescopic and Voyager data produces a complimentary data set which is self consistent, accurate, and covers a wide range of phase angles.

Millis and Thompson provide comprehensive set of telescopic data for 0.44 \( \mu \text{m} \) and 0.55 \( \mu \text{m} \). This data set produces a rotational phase curve which is used to correct the solar phase curves. The same rotational phase curve is used to correct both the telescopic and Voyager data. More than 200 Voyager images provide the necessary range of phase angles needed to construct the phase curve. The Voyager images are from the clear, blue, green, and orange camera filters. The effective wavelength of these filters are 0.48 \( \mu \text{m} \) for blue and clear, 0.58 \( \mu \text{m} \) for green, and 0.61 \( \mu \text{m} \) for orange. As can be seen, these effective wavelengths can be conveniently binned into two groups: 0.44 \( \mu \text{m} \) and 0.55 \( \mu \text{m} \) as determined by the position of the sub-observer point. The correction for each filter when binned is given by Nelson and Hapke's albedo spectrum of Ganymede. In addition, both the telescopic and Voyager data were divided into leading versus trailing hemisphere. Therefore, there are four data sets: leading and trailing hemispheres and 0.55 \( \mu \text{m} \) and 0.44 \( \mu \text{m} \).
PHASE CURVES: G. Hanson, Advisor: Dr. D. Domingue

The leading and trailing edge telescopic data was corrected for rotational variations and must be normalized to a Lambert disk.

In order for the Voyager data to be compared and combined with the telescopic data, the same corrections for rotation that were made to the telescopic data are repeated on the Voyager data. The calibration software used to radiometrically calibrate each image was the PICS software package developed by the United States Geological Survey. This software normalizes every image to a Lambert disk. After these corrections were completed, four additional corrections were made. In each case, the procedure is repeated for the leading and the trailing hemispheres. Corrections for filter, camera, mission, and telescopic versus Voyager data were made. After these corrections were completed the telescopic and Voyager data was combined into four complete solar phase curves; one for each hemisphere and each filter. Fig. 3 is an example of a complete solar phase curve for the leading hemisphere at 0.55 µm.

We have constructed four complete phase curves. These phase curves are indicative of the structural and textural properties of the optically active regolith on Ganymede.


Figure 1: The solar phase curve of the leading hemisphere of Ganymede at 0.44 µm and Europa and Callisto at 0.47 µm.
I. INTRODUCTION

Previous studies have shown that the compositions of the Apollo 15 olivine-normative basalts (ONB) are related by crystal fractionation of olivine (Ryder & Steele, 1988) (Shervais et al, 1990). However, recent work (Schuraytz & Ryder, 1991) suggests that the petrographic characteristics of the samples are not consistent with a simple model of olivine fractionation on the lunar surface. For example the abundance of olivine in the samples appears to be less than that required to account for the range of variation in MgO for the suite. Also, several samples (generally the coarser, more MgO-rich examples) scatter off the olivine control line. In order to better understand the causes of these apparent discrepancies and the overall petrogenesis of the Apollo 15 ONB suite, a more thorough understanding of the textural relationships among the samples is needed. The objective of this project is to provide detailed petrographic descriptions of samples that represent the range of chemical and textural variation in the suite.

II. APPROACH

The first step was obtaining precise modal analyses. This was accomplished by point counting thin sections using a petrographic microscope with an automated stage. The technique is analogous to placing a grid upon the thin section and identifying the minerals at each grid intersection. The abundance of constituent minerals or the mode (percent of total area) is obtained by dividing the number of points counted for each phase by the total number of points tallied. This study collected data for phenocrystic olivine, groundmass olivine, pyroxene, plagioclase, opaques, vesicles, and "others" (minerals that do not fit into one of the above groups, e.g. cristobalite, mesostasis, etc.).

The olivines of this suite show a wide range of chemical and textural variations. Olivines that have been identified under the electron microprobe as having forsteritic composition (Schuraytz & Ryder, 1991) appear as large phenocrysts that are clear under the plane polarized light of the microscope. Olivines that have been identified as having fayalite compositions are viewed as small crystals that are green under plane polarized light. The need for making the distinction between the two varieties of olivine deals with how they control the chemistry of the magma. Forsteritic olivine forms early on and controls the composition of the melt as it crystallizes. Fayalite as well as pyroxene, plagioclase, opaques, vesicles, and "others" form later on in the crystallization sequence and are not a controlling factor for the melt composition. Because of this, all mineral phases except for olivine have been recorded as one variety regardless of their composition or texture.

In addition to the traditional point counting method that uses the petrographic microscope we also explored the use of computer-aided image processing to obtain modal analyses. This method involved collecting a back-scattered electron image of a thin section with a scanning electron microscope (SEM). The back-scattered image allows the distinction among minerals to be made on the basis of gray-level or pixel intensity, which increases with atomic number of the constituent elements. Since few minerals consist of a single composition throughout an entire grain it is necessary to define a range of pixel intensities instead of a single value. This can be accomplished by defining thresholds for the grey-levels of each mineral. The computer then calculates the area percent for each range of pixel intensities.

By using the image processing approach for determining mineral abundances, many advantages are gained over point counting. One such benefit is avoiding mistaken mineral identification. Minerals such as olivine and pyroxene are often difficult to distinguish from one another when using a microscope. By using image processing approach it is no longer necessary to determine relief or to get interference figures to identify phases any longer. This method eliminates these long time consuming techniques. Figure 1 shows a portion of sample 15016,7 as it is seen under the petrographic microscopes using plane polarized light. Figure 2 is of the same area that figure 1 pictures but this image is a back-scattered image that was collected on the SEM. The SEM image clearly shows each mineral phases present as a certain grey-level. The three large crystals that are a medium grey shade are olivine crystals. This is known because the particular shade of grey associated with the crystals is known to be that of olivine. This identification only took a few seconds and is almost guaranteed to be right. When this thin section was looked at under a petrographic microscope things were not so obvious. The three grains were identified as pyroxene (instead of olivine) because the grain appears to be of lower relief than that expected for olivine and the grains surrounding the three appearance wise look the same and they are pyroxene.

The image processing approach for obtaining modal abundances is not perfect though. Many minerals (especially olivine and pyroxene) are often complexly zoned and this cause the gray levels for each mineral phases to often run together. Regardless of this, the method has been proven to be useful and fairly easy to implement. Therefore further research in making it better must be done.
UV-Blue 0.43 μm Absorption Feature for Five C-Class Asteroids
Erin C. Hatch, Centenary College of Louisiana
Advisor: Faith Vilas, Johnson Space Center, NASA

Introduction
The process of using reflectance spectroscopy to gain information concerning the composition of asteroids has been around since Bobrovnikoff first found variances in microphotometric tracings of photographic spectra of twelve different asteroids in 1929. (1) When incident sunlight is transmitted through approximately the first 200 μm of the surface material of an asteroid, the reflected light is affected by the mineralogical content of the asteroid. Some of this light is then reflected towards earth, where a telescope equipped with a spectograph records the spectrum. These spectral data give information about the mineralogical content and, also, the possible classification of each asteroid.

The asteroid classification system has conveniently divided asteroids into different classes depending on the appearance (shape, slope) of their spectra. For instance, the C-class asteroids, the class of the asteroids in this investigation, were so classified because the spectra are very similar to those of carbonaceous chondrites. However, while the classification of asteroids may provide valid boundaries between which more specific properties can be searched for, it cannot be viewed as a complete description of mineralogy. It is the attempt of this work to contribute further to our understanding of the composition of C-class asteroids.

Original classifications were made using low resolution spectrophotometry (passband = 250 A). From these spectra, only slope, shape, and large absorption features could be seen. Information as to more specific composition and, thus, similarities to terrestrial samples can be determined using higher resolution spectra. The data presented here were collected on the 1-m telescope using a spectograph coupled with a two-dimensional UV- and blue-sensitive photon counter (2D-Frutti) detector from April 25-26, 1991, at Cerro Tololo Inter-American Observatory (CTIO) in La Serena, Chile. Additional data were taken using a 1.5-m telescope with spectograph and UV- and blue-sensitive CCD at University of Arizona Observatories at Mt. Bigelow.

The dispersion of each spectrum is 2.4 A/pixel for data taken at Cerro Tololo, and ~ 9 A/pixel for data taken at the University of Arizona. This high resolution of these spectra allows for the detection of more specific and weaker absorption bands, and thus greater clarity as to compositional content of the asteroids.

Data Reduction
Flux calibrated data from Chile arrived on magnetic tape. After dumping that information onto the VAX system, three programs were run to convert from tape to VAX format, to scale the values to a reflectance value of 1.0 at 0.56um, and to align all the spectra for the same asteroid on the same night.

After the data were in a workable format, each spectrum was median filtered against itself using a program that ordered the reflectance values for an odd number of pixels centered around one chosen pixel, picking the median value of the total number, and replacing the chosen pixel value with that median value. This process decreases the resolution from 4.8 A, to approximately 24 A; however, this helps to smooth the spectra, reducing noise for added ease in detecting subtle absorption features.

At this point, the values for asteroid spectra (66 Maja, 326 Tamara, 165 Loreley, 559 Nanon, and 130 Elektra) were ratioed against spectral values for one of two Solar Analog stars that were used for calibration sources. Solar Analogs SAO 120107 and SAO 159706 were used for this study. Both of these stars have spectra that are considered close to the sun's spectrum. By ratioing these two spectra, the majority of the absorption features due to the G2-type star and atmospheric conditions are eliminated. What is left is a semi-pure spectrum of the asteroid mineralogy.

From these spectra, any spectra for the same asteroid on the same night were coadded (averaged). It is from these final spectra that the following conclusions have been drawn.

Analysis of Spectra
At the start of this investigation of the absorption features found in C-class asteroids in the lower blue-UV spectral region, we were seeking an absorption band located at about 0.4μm indicative of porphyrins (organic material). This feature has been investigated by P.N. Holden and is named the porphyrin Soret band. (3) This absorption feature has been found in terrestrial shale deposits as well as in the meteorite Orgueil. The porphyrin concentrations in the shale samples were as low as 8.25 ppm. Holden estimates than in extra-terrestrial remote sensing, such as this, porphyrin concentrations of at least 10 ppm will be required in order to detect the Soret band. Even from first examination, it was apparent that this feature was not in the asteroid spectra that were being examined. Therefore, from this knowledge, and the previous research done by Holden, it can be estimated that if organic porphyrins are present in these five asteroids (all located within the main asteroid belt (2.2 to 3.2 AU))) they are present in concentrations less than 10 ppm.

After examining the spectra more closely, however, a prominent dip of approximately 3 to 5% was seen at about 0.43μm. As seen in the following figures, this feature is seen in all five C-class asteroids investigated. Three of the spectra presented here (Maja, Loreley and Tamara) were ratioed using SAO 159706. The other two (Nanon and Elektra) were ratioed using SAO 120107. This suggests that the feature could only be caused by the asteroid itself, and not by the features present in any one calibration star. In addition, comparison graphs were made between each ratioed spectrum, and the spectra of the star and that of asteroid before ratioing. Before ratioing, the stellar features in each individual asteroid spectrum were clearly more intense than any features in the star itself. Also, the 0.43μm feature was both more intense and covered a wider
Figure 2: The solar phase curve of the trailing hemisphere of Ganymede at 0.44 \( \mu \text{m} \) and Europa and Callisto at 0.47 \( \mu \text{m} \).

Figure 3: The combined telescopic and Voyager solar phase curve for the leading hemisphere at 0.55 \( \mu \text{m} \).
UV-Blue 0.43 um Absorption Feature for Five C-Class Asteroids

Erin C. Hatch, Centenary College of Louisiana

wavelength position of this feature indicates that the mineral most likely involved is iron. From this speculated mineralogy, there are several transition candidates that can be examined.

First, a spin-allowed Fe$^{2+}$ crystal field transition could be possible (8, 9, 11, 12). This type of transition occurs between electron orbitals within a particular electronic shell, most commonly the d-shell. For example, the 1 um features in olivine and pyroxene are of such a transition. However, this transition is often relatively weak and generally located at longer wavelengths (>0.8um). Therefore, it is very unlikely that this transition could be responsible for this feature.

On the other hand, a spin-forbidden Fe$^{3+}$ crystal field transition could also be responsible (10). This type of transition is similar to the spin-allowed Fe$^{2+}$, with an electron spin flip added. For example, the sharp narrow 0.5um features in the diogenite spectra are spin-forbidden Fe$^{2+}$ crystal field transitions. These occur in the same wavelength range as the feature we are seeing, but are generally very weak. Due to the relative strength of this feature, it is extremely unlikely that this transition could be responsible for the 0.43um band we are seeing.

By the same token, a spin-forbidden Fe$^{3+}$ crystal field transition is another possible responsible transition (11). This crystal field transition is similar to the previous two, but in this case, all five d-shells already contain one unpaired electron, so an electron spin flip is required. An example of this type of transition is the sharp narrow feature near 0.5um in certain clinoxyroxenes. There is also a related broad absorption band with a dip of ~3.2% centered at near 0.7um. This feature is present in four of the five C-class asteroids we observed (see spectral graphs on the previous page). These transitions also occur in the right general wavelength range, but are generally weak as well. Again, there is a low probability that this crystal field transition could account for the feature we are observing.

A fourth transition type could be an Fe$^{3+}$ charge transition. This type involves an electron being exchanged between different electronic shells within a cation. An example of this kind of transition is the series of features forming the blue absorption edge in hematite (Fe$^{3+}$-oxide). These features occur in the same general spectral region as the 0.43um feature we’ve observed, and are also as strong in intensity. It is probable that this type of transition could occur in the five C-class asteroids we have observed.

A final charge transition type that should be considered is an Fe$^{2+}$ - Fe$^{3+}$ charge transfer (11). This involves the exchange of an electron between adjacent cations. The series of features forming the blue absorption edge in mixed valence iron-oxides and clay minerals are examples of this charge transfer transition. These occur within the correct wavelength range to account for the 0.43um feature, and are also very strong. Therefore, the Fe$^{2+}$ - Fe$^{3+}$ charge transfer transition is another probable candidate for the origin of the 0.43um feature we have observed.

There are several examples of features similar to the 0.43um feature that can be found in various iron-related minerals. The jarosite (iron-bearing aluminum-sulfates) and iron-bearing aluminum-oxides (13) spectra contain a strong feature (Fe$^{3+}$ CT or Fe$^{2+}$ - Fe$^{3+}$ CT) at the right position. In addition, several iron oxides (ex. goethite) and iron-bearing clay minerals have a strong absorption feature at about the right wavelength (11). However, as stated before, the laboratory results are not calibrated sufficiently in order to pin down the exact mineralogy of this absorption feature. Further work to come to a more specific conclusion of the source of the 0.43um feature is under way.

A completely different possibility for the source of this absorption band is the presence of a non-porphyrin, organic material. Earlier examination has shown that the 0.4um feature for porphyrins (3) does not match this feature. However, the intensity and width of this feature is characteristic to organics (9, 11). However, further investigation would be required in order to find a relationship between this feature and any particular organic absorption band. Inquiry into this possibility is presently being conducted.

We have found this feature in all C-class asteroids with sufficient spectral resolution in the blue-UV range. Investigations are presently being made into spectra of other C-class asteroids for similarities in spectral features.

* Special thanks to F. Vilas and M.J. Gaffey for all their help and guidance.

References

UV-Blue 0.43 um Absorption Feature for Five C-Class Asteroids
Erin C. Hatch, Centenary College of Louisiana

area that any one stellar feature, giving rise to justification for the feature in the ratioed spectrum.

Combination Spectral Graphs for Five C-Class Asteroids

0.43um Absorption Feature

0.7um Absorption Feature

** Listed by increasing mean heliocentric distance: 326 Tamara (2.31766 AU), 66 Maja (2.64599 AU), 559 Nanon (2.71165 AU), 130 Elektra (3.11749 AU), and 165 Loreley (3.12925 AU)

McCord and Gaffey (4) and the Eight-Color Asteroid Survey (5) have also recorded spectra for several C-class asteroids. However, the resolution is very low and any 0.43um features present in spectra of these asteroids are difficult to see.

A similar feature at the same locations was discussed in an earlier paper by Luu and Jewitt, found in near-Earth asteroid 3908 (1980 PA). (6) However, this asteroid is a V-class asteroid, and, therefore, not of the C-class asteroids that are being investigated here, so the relationship between both of these features is unclear at this time.

In addition, Jewitt and Luu have recorded high resolution spectra in this region for asteroids 659 Nestor (XC) and 1208 Troilus (FCU). (7) It is not apparent from these spectra, however, whether the 0.43um feature is present or not. Further work on the same asteroids' spectra by further reducing background noise would be necessary in order to come to any more specific conclusions.

In several telescopic spectra of Mars, this same feature is possibly present. (8) There is some evidence of a feature at the same wavelength with approximately the same intensity in manufactured iron rich amorphous soils believed to be similar to Martian soils.

Discussion

At the present time there is too little calibration in laboratory data on mineral species possibly found in C-chondrites for this spectral region to assign a specific mineral phase responsible for the 0.43um absorption feature. We can only speculate as to the cause of this strong feature in an area where only smaller features are usually evident. This region of low reflectance is due to the low level of reflectance light caused by the dark surface of the asteroid. However, the
PETROLOGICAL VARIATIONS IN APOLLO 15 .... : JOHNSON, P.D., SCHURAYTZ, B.C. & RYDER, G.

The second phase of the project involved examining the samples for important textural information. This includes inspecting each thin section for any features such as grain size, crystal types, mineral distributions, textures (e.g. vesicular, poikilitic, etc.), and any other important attributes.

III. RESULTS & INTERPRETATIONS

The first comparison taken was to contrast the modal data obtained by point counting with data determined by image-processing. Table 1 shows that the values for any abundances of the mineral phases all, except for the plagioclase value, at least within a somewhat reasonable limit of being correct. The plagioclase abundance probably is incorrect because the range of grey shades overlapped into shades that are identified as other phases. This shows that although we are able to obtain rough modal abundances for the mineral phase the accuracy of the data is not good enough for the results to be used in comparisons.

The above data shows that modal analyses can be performed on the SEM as the values collected are at least in the right ballpark when they are compare with the values taken by point count. This method will prove especially useful once a method of eliminating the overlap of mineral phases can be found so that each phase will be represented as a single grey value instead of a range of greys. Once this has been taken care of much more precise modal analyses will be able to be done on the system and hopefully in time it will also take less time than it does to do a modal analysis on the petrographic microscope.

The next problem to deal with is why the modal abundance of olivine in the samples seems to be less than that required to account for the variation in the MgO of the suite. This problem can be explained by examining the modal data and comparing it to the chemical data. For basalts, like this suite, the modal olivine abundance should be directly related to the amount of MgO. For an increase of MgO there would also be an expected increase in the amount of olivine also. Yet in figure 5 two samples (15119 and 15545) of the seven samples do not lie on this expected trend.

One answer to this situation deals with this difficulty in identifying pyroxene and olivine. If for any reason olivine was mistaken for pyroxene this circumstance can be explained. This is because if the olivine grains were not recognized for what they were and were tallied as pyroxene the total abundance of olivine would not go up but the chemical analysts would still detect the presence of the MgO and determine in the norm that there should be more olivine.

Besides the case of mistaken identity there is another possibility that can explain why these points do not lie on the trend. Pyroxene grains have been examined in some the thin sections (such as 15555,209) that have olivine cores. Some pyroxene grains have such complex zoning that it might be very difficult to pick out these olivine cores so when the sections were point counted the presence of all of the cores of olivine was not recognized. This method would also explain why there is such a variation in MgO as all of the olivine cores will be detected by chemical analyses but not enough modal olivine can be seen in the modal data because the hidden cores were not detected.

Yet one more solution for this problem deals with the error in the modal abundance values. The two samples (15119 and 15545) that are not sitting on the trend have had very few points counted (534 and 489 points respectively) for them. Samples such as 15555 and 15556, on the other hand, have had close to 12,000 points counted for each of them. This means that although samples 15119 and 15545 might have had only 2% or 3% olivine phenocrysts tallied for each, the total is likely be up between 4% and 5% as the error has been increased by having so few points counted. A way to check to see if this case would be to count another thin section for each sample or else shorten the grid spacing for the point so that more points can be collected in a smaller area.

Among the samples there exists a great deal of textural variation. One such texture is in the great span of grain sizes for the suite of samples. Table 2 lists the maximum widths for the plagioclase found found in each thin section. Walker, 1976 used plagioclase widths as a means to measure the grain size because the other minerals such as pyroxene exist both as phenocrysts and groundmass while plagioclase since it is late forming are all of the same variety. It is quite interesting to note that this method of grain scale measurement does not agree with the rough eyeball method. This is more than likely due to the degree of the absorption of the phenocrystic grains of minerals like pyroxene and olivine so that the more that survived the coarser the thin section will appear.

One other interesting textural feature is how the phenocrystic olivine changes from sample to sample. Sample 15555 which is a coarse grained sample has anhedral embayed phenocrysts that are 0.68 mm wide while the crystals show little of there original shape. Meanwhile, 15556, which is a fine grained sample, has phenocrystic olivine that is also embayed but has a much shorter width of 0.33 mm and very unlike 15555's grains the crystals are skeletal and therefor show their former grain shape. This variation of phenocryst form seems to vary with the grain size so that the coarser samples phenocrysts are anhedral to subhedral and the finer grained samples are subhedral to euohedral or at least skelataly euohedral.

IV. CONCLUSIONS

The Apollo 15 ONBs still have many questions yet unanswered about them even after this study. The situation of why there does not seem to be enough modal olivine to account for the variation in MgO could to be
PETROGRAPHIC VARIATIONS AMONG THE APOLLO 15 OLIVINE-NORMATIVE MARE BASALTS
by Paul D. Johnson1, ADVISORS: Benjamin C. Schuraytz2, Graham Ryder2
1 University of Hawaii at Manoa, Honolulu
2 Lunar and Planetary Institute, Houston

I. INTRODUCTION
Previous studies have shown that the compositions of the Apollo 15 olivine-normative basalts (ONB) are related by crystal fractionation of olivine (Ryder & Steele, 1988) (Shervais et al, 1990). However, recent work (Schuraytz & Ryder, 1991) suggests that the petrographic characteristics of the samples are not consistent with a simple model of olivine fractionation on the lunar surface. For example the abundance of olivine in the samples appears to be less than that required to account for the range of variation in MgO for the suite. Also, several samples (generally the coarser, more MgO-rich examples) scatter off the olivine control line. In order to better understand the causes of these apparent discrepancies and the overall petrogenesis of the Apollo 15 ONB suite, a more thorough understanding of the textural relationships among the samples is needed. The objective of this project is to provide detailed petrographic descriptions of samples that represent the range of chemical and textural variation in the suite.

II. APPROACH
The first step was obtaining precise modal analyses. This was accomplished by point counting thin sections using a petrographic microscope with an automated stage. The technique is analogous to placing a grid upon the thin section and identifying the minerals at each grid intersection. The abundance of constituent minerals or the mode (percent of total area) is obtained by dividing the number of points counted for each phase by the total number of points tallied. This study collected data for phenocrystic olivine, groundmass olivine, pyroxene, plagioclase, opaques, vesicles, and "others" (minerals that do not fit into one of the above groups, e.g. cristobalite, mesostasis, etc.).

The olivines of this suite show a wide range of chemical and textural variations. Olivines that have been identified under the electron microprobe as having forsteritic composition (Schuraytz & Ryder, 1991) appear as large phenocrystals that are clear under the plane polarized light of the microscope. Olivines that have been identified as having fayalite compositions are viewed as small crystals that are green under plane polarized light. The need for making the distinction between the two varieties of olivine deals with how they control the chemistry of the magma. Forsteritic olivine forms early on and controls the composition of the melt as it crystallizes. Fayalite as well as pyroxene, plagioclase, opaques, vesicles, and "others" form later on in the crystallization sequence and are not a controlling factor for the melt composition. Because of this, all mineral phases except for olivine have been recorded as one variety regardless of their composition or texture.

In addition to the traditional point counting method that uses the petrographic microscope we also explored the use of computer-aided image processing to obtain modal analyses. This method involved collecting a back-scattered electron image of a thin section with a scanning electron microscope (SEM). The back-scattered image allows the distinction among minerals to be made on the basis of gray-level or pixel intensity, which increases with atomic number of the constituent elements. Since few minerals consist of a single composition throughout an entire grain it is necessary to define a range of pixel intensities instead of a single value. This can be accomplished by defining a thresholds for the grey-levels of each mineral. The computer then calculates the area percent for each range of pixel intensities.

By using the image processing approach for determining mineral abundances, many advantages are gained over point counting. One such benefit is avoiding mistaken mineral identification. Minerals such as olivine and pyroxene are often difficult to distinguish from one another when using a microscope. By using image processing approach it is no longer necessary to determine relief or to get interference figures to identify phases any longer. This method eliminates these long time consuming techniques. Figure 1 shows a portion of sample 15016,7 as it is seen under the petrographic microscopes using plane polarized light. Figure 2 is of the same area that figure 1 pictures but this image is a back-scattered image that was collected on the SEM. The SEM image clearly shows each mineral phases present as a certain grey-level. The three large crystals that are a medium grey shade are olivine crystals. This is known because the particular shade of grey associated with the crystals is known to be that of olivine. This identification only took a few seconds and is almost guaranteed to be right. When this thin section was looked at under a petrographic microscope things were not so obvious. The three grains were identified as pyroxene (instead of olivine) because the grain appears to be of lower relief than that expected for olivine and the grains surrounding the three appearance wise look the same and they are pyroxene.

The image processing approach for obtaining modal abundances is not perfect though. Many minerals (especially olivine and pyroxene) are often complexly zoned and this cause the gray levels for each mineral phases to often run together. Regardless of this, the method has been proven to be useful and fairly easy to implement. Therefore further research in making it better must be done.
PNEUMATIC CONVEYING OF SOLIDS IN PARTIAL GRAVITY: E. Koenig,  
Lunar and Planetary Institute

As plans are being made for the establishment of a permanent lunar outpost, the concept of lunar resource utilization gives a new meaning to the phrase "living off the land." Lunar soil may be used to produce the oxygen, water, and radiation shielding astronauts will need to survive on the moon. Lunar soil may also be used a source of propellants, metals, and carbon dioxide to support plant growth.[1] These operations all will require the mining and transport of lunar regolith.

The goals of this project are to evaluate the feasibility of a pneumatic conveying system for the transport of lunar soil. A pneumatic conveyor involves the use of a gas stream to move solids through pipelines, and has several advantages over conventional hauling methods (dump trucks, mechanical conveyors). For example, pneumatic conveying systems require low maintenance and are easily automated, thus reducing the need for expensive astronaut time. Furthermore, the disadvantages of pneumatic conveying systems found on Earth may be lessened in a partial gravity environment. Wear and abrasion of the conveyor, resulting from sand-blasting effects of the transported solid, is proportional to the particle velocity cubed (for most industrial situations). [2] In partial gravity, lower velocity is required for pneumatic conveying; equipment erosion should then be reduced exponentially.

Pneumatic conveying systems have proven successful on Earth, but their behavior in partial gravity has not been explored. The main objectives of this study are to observe the behavior of a pneumatic conveying system in reduced gravity, determine key parameters of operation for the conveyor, and compare the results to 1 g behavior and theoretical analysis. Carbotek, Inc. has provided theoretical analysis and a design for a pneumatic system in which compressed air will be used to transfer solid glass beads through acrylic tubing. Flights are scheduled to test this design in .16 g (lunar, 2 flights) and .38 g (martian, 1 flight) on the NASA KC-135, a plane which achieves partial gravity for brief periods of time by flying parabolic trajectories. Two sizes of glass beads (150 µ, 50 µ) will be tested. Ground runs will also include the testing of hollow glass bubbles with the approximately the same diameter (50 µ), but \( \frac{1}{6} \) the density of the solid beads, to determine how well \( \frac{1}{6} \) density models \( \frac{1}{6} \) gravity. A summary of the data to be collected is listed below:

<table>
<thead>
<tr>
<th>SOLID</th>
<th>1 g</th>
<th>.16 g</th>
<th>.38 g</th>
</tr>
</thead>
<tbody>
<tr>
<td>150 µ beads</td>
<td>x</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>50 µ beads</td>
<td>x</td>
<td>x</td>
<td>no flight</td>
</tr>
<tr>
<td>~ 50 µ &quot;bubbles&quot;</td>
<td>x</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Both horizontal and vertical transfer will be studied, with a set of three tests. These tests include horizontal dilute phase, vertical dilute phase, and vertical dense phase transfer.

The horizontal transfer apparatus is shown in Figure 1. A feeder will supply the glass beads to an eighteen foot long run of tubing. An eductor will be located at the point of solids entry. Flowing air supplied from K-bottles will then transport the beads along the tube length; a cyclone at the end of the run will separate the solids and gas. The beads will fall into a receiver, while the air will be taken to a 55 gallon drum, passed through a filter to remove any remaining glass beads, and vented. The purpose of this test is to determine the saltation velocity for a given particle size and feed rate. This is the critical point at which the air can no longer transfer the solids, and will be found by varying the air flow for each parabola of the flight.

Figure 2 shows the vertical transfer apparatus. The feeder described above will supply solids to the base of a six foot tall tube. This run will terminate in its own separate cyclone and receiver, but the K-bottles and 55 gallon drum from the horizontal tests will be used for these tests as well. For the vertical dilute phase experiment, with a high gas/solids ratio, the air flow rate will
due to a mistake in the identification of olivine cores found within pyroxene grains. This would account for the high MgO and the low modal abundance of olivine. Another issue that might have been answered is why there is scattering on the olivine control line for MgO rich rocks. This appears to have to do with the abundance of plagioclase in the sample. Since plagioclase has been found to vary from thin section to thin section in this sample than it might have just as easily varied in the chemical samples that were used. And finally a look at the textures shows that there is a wide variety of textures. Much more data exists about the textures that can be interpreted at another time as this summer there was not enough time do it all.

References

Acknowledgments
I would like to thank Ronald Bernhard of Lockheed Engineering & Science Co. for the great assistance in aiding me with the modal analysis by image processing approach. Without his help that part of this project would not have advance as far as it did.

Table 1

<table>
<thead>
<tr>
<th>Pt Count</th>
<th>Olivine</th>
<th>Pyroxene</th>
<th>Plag.</th>
<th>Oxides</th>
</tr>
</thead>
<tbody>
<tr>
<td>Image Proc</td>
<td>5.9701</td>
<td>56.7164</td>
<td>30.9168</td>
<td>4.0512</td>
</tr>
</tbody>
</table>

Table 2

<table>
<thead>
<tr>
<th>Thin Section</th>
<th>Plag Width (mm)</th>
<th>Eyeballed Gain Size</th>
</tr>
</thead>
<tbody>
<tr>
<td>15529</td>
<td>0.07</td>
<td>F</td>
</tr>
<tr>
<td>15016</td>
<td>0.16</td>
<td>M</td>
</tr>
<tr>
<td>15535</td>
<td>0.18</td>
<td>M</td>
</tr>
<tr>
<td>15119</td>
<td>0.20</td>
<td>M</td>
</tr>
<tr>
<td>15556</td>
<td>0.24</td>
<td>F</td>
</tr>
<tr>
<td>15545</td>
<td>0.80</td>
<td>F</td>
</tr>
<tr>
<td>15555</td>
<td>2.20</td>
<td>C</td>
</tr>
</tbody>
</table>
Figure 1. Horizontal Transfer Apparatus

Figure 2. Vertical Transfer Apparatus
PNEUMATIC CONVEYING: E. Koenig

be varied at a constant feed rate to determine the choking velocity for a given particle size. This is analogous to the horizontal saltation velocity. The vertical dense phase experiment, with a low gas/solids ratio will study air pressure and flow rate requirements at varying solid feed rates for a constant particle size.

As of the first week of August, 1991, the acrylic equipment has been fabricated and delivered. Assembly of the parts in their final configuration is near completion, and ground testing is scheduled to begin in the middle of the second week of August. The test equipment data package, including structural analysis, safety analysis, and test protocol, has been completed. The experiment is awaiting inspection and approval from safety engineers. KC-135 flights are scheduled for the final week of August (Aug. 26, 27).

It is hoped that information gathered from this pilot study will be useful in the design and construction of lunar mining systems. By knowing gas pressure and flow rate requirements, system designs for pneumatic conveying experiments will be based on facts, and factors such as compressor power requirements, pipe diameters, and pressure variations can be predicted. This will allow for more efficient and advanced systems, and lead to greater overall success in outpost objectives. Someday even the martian gravity data may be needed during operations on that planet. A project such as this pneumatic study underscores the seriousness of the commitment to human exploration of space. It begins the process of making visions and theories into reality. In pushing back the limits of our knowledge, it becomes possible to push beyond the limits of our Earth.

REFERENCES

FIGURE 1. High resolution transmission electron micrographs of standard clay minerals stabilized with alkylammonium cations: (a) Interstratified illite/smectite intercalated with nonylammonium ions (9-chain carbon compound). Illite remains unexpanded at 1 nm and the smectite component has expanded to 1.8 nm; (b) Same clay as in (a) treated with octadecyl ammonium ions (18-chain carbon compound). Smectite has expanded to 2.2 nm; (c) Saponite treated with octadecyl ammonium ions ($N_c = 18$); (d) Same as (c) magnified to show 1.5 nm spacings of saponite intercalated with octadecyl ammonium ions.

For TEM analysis, small subsamples of n-alkylammonium saturated clays were carefully removed from the glass slides prepared for XRD analysis. The samples were transferred to an epoxy mount and allowed to dry. Thin sections (70 to 100 nm) were prepared on a microtome. Thin sections were placed on holey grids, carbon coated, and then analyzed under TEM.

RESULTS
Intercalation of n-alkylammonium ions stabilized the delicate structure of the standard clay minerals under the electron beam; high resolution lattice fringes were observed in all of the standard clays. The basal plane d-spacing is clearly observed in the interstratified illite/smectite and saponite clay standards shown in
METHOD FOR THE CHARACTERIZATION OF PHYLLOSILICATES IN EXTRATERRESTRIAL MATERIALS; Katalin Krivian

INTRODUCTION

The chemical and mineralogical compositions of meteorites and interplanetary dust particles (IDPs) provide important information about their origin and the environments to which they were exposed subsequently. Carbonaceous chondrites are a class of meteorites that contain high water (1 to 20 wt. %) and high carbon contents (0.6 to 2.8 wt. %) [1]. Most carbonaceous chondrites and IDPs have undergone aqueous alteration, which may have been an important process in the early solar nebula. One of the indicator minerals of this process is hydrous phyllosilicates. Most of the previous work on the characterization of phyllosilicates in these primitive materials have been performed using electron beam instruments. The mineralogical characterization of these materials, however, has been severely hampered by the tendency of phyllosilicates to be rendered amorphous, and unrecognizable, through heating and radiation effects of the electron beam. The objective of this study was to develop a technique that stabilized phyllosilicates against electron beam damage and aided the identification of the phyllosilicate type. The technique will employ a chemical method where various lengths of carbon-chained alkylammonium ions will be intercalated into the interlayers of 2:1 expanding phyllosilicates.

BACKGROUND

Phyllosilicates are crystalline, hydrated 2:1 or 1:1 layered aluminosilicates. The 2:1 and 1:1 layered structures consist of layers of tetrahedral-octahedral and tetrahedral-octahedral-tetrahedral sheets, respectively. The space located between layers is known as the interlayer. The interlayers of 2:1 layered minerals consist of hydrated, inorganic cations, usually calcium and magnesium. These hydrated inorganic cations satisfy a net negative charge created by isomorphic substitution within tetrahedral and octahedral sheets. Classes of phyllosilicates are differentiated according to the net negative charge generated in their tetrahedral and octahedral layers. For example, smectites have a negative charge of < 0.5 per one-half unit cell; whereas, vermiculites will have a charge density of 0.6 to 0.9 for each one-half unit cell. In most cases, the interlayer cation can be replaced by other cations in solution (either inorganic or organic).

Various organic cations, e.g., n-alkylammonium ions, will readily intercalate into the interlayers of most 2:1 layered phyllosilicates (except mica and chlorite). The interlayer will expand to a distance that will accommodate the n-alkylammonium ion depending on the net negative charge density of the 2:1 layer. The chain length of the n-alkylammonium ion will also have an effect on the expansion distance within the interlayer and this relationship is known for a given class of phyllosilicates [2]. The charge density of the 2:1 layer can be determined by measuring the expansion distance by x-ray diffraction analysis (XRD) [2]. Recent studies have also shown that 2:1 phyllosilicates may be stabilized under the electron beam by intercalating with large non-hydrated organic cations [3]. However, techniques have not been developed to analyze very small sample sizes (e.g., IDPs, carbonaceous chondrites).

MATERIALS AND METHODS

Synthesis of n-alkylammonium cations. Four n-alkylammonium chloride salts (carbon chain lengths of 9, 12, 14, and 18) were synthesized by a procedure analogous to that employed by Ruehlcke and Kohler [4]. Starting materials for the synthesis of the chlorides were equal volumes of the specific amine (nonylamine, dodecylamine, tetradecylamine, and octadecylamine) and ethanol (95%). The only variation from Ruehlcke and Kohler's procedure was to freeze dry the crystalline powder after synthesis, thereby removing any excess acetone, ethanol, or water.

Intercalation of phyllosilicates with n-alkylammonium ions. Seven Na-saturated standard clay minerals (Clay Minerals Repository) were each treated separately with four different lengths of n-alkylammonium ions (Nc = 9, 12, 14, & 18). The standards included a nontronite (Washington), vermiculite (Texas), illite (Montana), saponite (California), interstratified illite/smectite (Montana), smectite (Arizona), and Fe-smectite (Washington). Approximately 15 mg of the Na-saturated clay were intercalated with the appropriate alkylammonium cation according to the method of Ruehlcke and Kohler [4].

X-ray diffraction and transmission electron microscopy analyses. Mineralogy of each standard clay was confirmed by x-ray diffraction analysis of Mg- and K-saturated samples using standard techniques [5]. Clay suspensions of alkylammonium-saturated clays were pipetted onto glass slides, then placed into a desiccator and dried under a vacuum for 12 hours. The slides were then x-rayed.
DARK STREAKS ON MARTIAN SLOPES: AN ASSESSMENT OF POSSIBLE ORIGINS

David Leverington, University of Ottawa, Ottawa, Ont., Canada.
Advisor: Dr. Steven Williams, Lunar and Planetary Institute, Houston, TX.

Introduction

Intriguing slope features that seem unlike those found on most martian slopes are revealed in a small fraction of Viking orbiter images. These features, which can be confidently identified only in high-resolution Viking images, are dark and streak-like, often contrasting considerably with the higher albedos of surrounding slope materials.

The best known examples of these dark talus streaks (DTS's) can be found on the slopes of the Olympus Mons aureole deposits. These features have a number of characteristics that set them apart from other dark slope materials that may be found upon most martian escarpments and craters [1]. They originate as point sources on the slopes, with low albedo material extending downslope from these points in a linear or wedge-like fashion (fig.1). Very sharp surface boundaries exist between the streaks and the other, lighter slope materials that surround them. The lengths of the streaks range from 200 to 2500 m, and the widths range from 20 to 250 m [1]. They appear to be constrained by local topographic variations, always following slope gradients [2] (fig.2).

Three models have recently been put forward to explain the origin and physical characteristics of the DTS's. These models, which are listed below, have potential implications regarding our understanding of martian surface and subsurface composition, bedrock exposure, and slope activities.

1) They result from the mass wasting of surface or subsurface inhomogeneities [3]. This model implies that there are masses of darker (perhaps ultra-mafic) material at or near the surfaces of some regions of Mars.

2) A process of slope darkening caused by the emergence of volatiles (such as water) from subsurface sources, perhaps involving a spring-like mechanism [1]. This model implies that near-surface volatiles exist on Mars. Furthermore, it implies that these volatiles occur as part of a mechanism that allows them to occasionally have access to the surface, and to be able to withstand the pressure and temperature constraints of Mars long enough to modify significant areas of some slopes.

3) They result from the redistribution of relatively bright slope dust by recent landslides, exposing the darker material below [2,4]. This model implies that DTS features can be explained not by surface inhomogeneities or internal geologic processes, but by a mass wasting analog to aeolian dark streaks.

This project involved the creation of a complete DTS database for the purpose of providing geological and morphological information on this martian phenomena, and for the evaluation of the streak models listed above.

Procedure

The procedure involved six stages:

1) The search of all Viking images with a resolution better than 80 meters/pixel for DTS candidates, using the Image Retrieval and Processing System (IRPS).

2) The creation of a DTS catalogue, separating high quality DTS candidates from marginal examples.

3) The correlation of all high quality candidates with regard to latitude, elevation, lithology, and thermal inertia (closely related to dust cover). Data sources included recent geological charts and the Geophysical Data Facility (GDF).

4) The close examination of selected DTS examples, utilizing image enhancement software systems.

5) The utilization of a logic chart to aid in the evaluation of the DTS models in terms of the new data. This chart, which is shown below, portrays possible optimum DTS correlations for each of the three models being tested (s=strong correlations, m=moderate, n=no / weak).

<table>
<thead>
<tr>
<th>model</th>
<th>latitude</th>
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<td>m/n</td>
</tr>
<tr>
<td>model 2</td>
<td>s</td>
<td>m/n</td>
<td>m/n</td>
<td>s</td>
</tr>
<tr>
<td>model 3</td>
<td>m/n</td>
<td>m/n</td>
<td>m/n</td>
<td>s</td>
</tr>
</tbody>
</table>

6) The combination of the logic chart results with a morphological evaluation of DTS features for final conclusions.
Fig. 1. The d-spacings observed in high resolution lattice images for the interstratified illite/smectite material intercalated with 9-carbon chain and 18-carbon chain alkylammonium ions were 1.8 and 2.2 nm, respectively, for the smectite phase. These d-spacings compared well with the d-spacings determined by XRD analysis (Fig. 2). The other important aspect of this experiment was that the d-spacing increased as the chain length of the alkylammonium ion became longer (see Fig. 2).

CONCLUSIONS

The results of this study show that the fragile structure of phyllosilicates can be successfully stabilized under the electron beam by using the n-alkylammonium method. The clay mineral type may also be identified based upon the d-spacing that results from the intercalation of various chain lengths of the alkylammonium ion. Such a technique will be useful in characterizing phyllosilicates in extraterrestrial samples (e.g., IDPs and carbonaceous chondrites) where only very small sample sizes are available for analyses.

REFERENCES


ACKNOWLEDGEMENTS

I wish to thank Doug Ming, D.C. Golden, Charlie Galindo and Mike Zolensky for this internship to become a reality for me. I also would like to thank all of them for their genuine determination in helping and supporting me throughout the course of my project.

FIGURE 2. X-ray diffractograms of Mg-saturated and alkylammonium cation stabilized illite/smectite mixed layers. The diffractograms show a slight increase in basal d-spacings with increase in chain length of alkylammonium ions.
DARK STREAKS ON MARTIAN SLOPES... Leverington, D.W.

Model 3 is supported by the correlation results. The physical characteristics of the streaks are not inconsistent with the model. Furthermore, its simplicity establishes the model as a particularly attractive explanation for DTS features. Should the observation of a high albedo talus cone upon a (presumably older) DTS feature be correct, however, this model must be reworked or discarded as a possible DTS process.

Conclusions

No model tested in this study can be shown to be strongly supported by the correlative and morphological data available. A major difficulty in the evaluation of the DTS features is the lack of coverage of Mars at resolutions better than those of the Viking orbiters. Until a much larger proportion of the planet is imaged at 30 m per pixel resolution, or a single DTS region is imaged at a very high resolution of perhaps 2 m per pixel, the origin and nature of the dark talus streaks will remain uncertain.

References


Figure 1. Shown are excellent examples of dark talus streaks (DTS's). Portion of viking orbiter image 713A57. Scale bar is 5 km long.

Figure 2. DTS's always follow the local topographic gradient. This example, a portion of Viking orbiter frame 441B07, is in the Olympus Mons aureole. The scale bar is 0.5 km long. Image processed at the Lunar and Planetary Institute.

Figure 3. A talus cone of higher albedo material appears to have been deposited on top of a DTS, suggesting perhaps, that two different processes are operating. This DTS lies very near that in figure 2, and is from the same Viking orbiter image. The scale bar is 0.5 km long. Image processed at the Lunar and Planetary Institute.
DARK STREAKS ON MARTIAN SLOPES... Leverington, D.W.

Results

The following conclusions can be drawn from the examination of the best DTS examples:

1) All best examples are found only in equatorial latitudes (30N to 20S).
2) The DTS features occur in a wide variety of geologic units.
3) The DTS features are not restricted by surface elevation, with best examples found as low as 0 km and as high as 6 km. The elevation distribution appears to mimic martian hypsometry.
4) All best DTS examples are found in areas of low thermal inertia. This suggests that the areas in which these candidates are found are mantled with a significant layer of fine dust.
5) Many DTS examples originate near the peaks of their respective slopes.
6) Most martian slopes do not appear to have DTS features.
7) Contrary to [1], DTS examples are occasionally found to narrow in a downslope direction.
8) Interestingly, in one example, a talus cone of higher albedo debris appears to have been deposited upon an excellent DTS example (fig.3). This seems to suggest that two separate slope processes are occurring in this area.

Discussion

Only the best candidates were chosen to be utilized in the final data analysis to avoid contamination of the results by misinterpreted slope features. It was discovered that no DTS candidates found on images having resolutions of 30 meters/pixel or worse could be identified with sufficient confidence to merit their inclusion in the dataset utilized for the final conclusions of this project. A dataset of several hundred images was created using this and other marginal data for possible future examination of martian dark slope streaks.

The final dataset includes only 35 images, which occur in 9 separate regions. In reviewing the conclusions based on the final dataset, it must be kept in mind that the entire catalogue of high-resolution Viking imagery covers only one percent of the martian surface [5], and is not randomly distributed (having been affected by factors such as regional atmospheric transparency and the targeting of specific martian features). An equatorial latitude bias exists for high-resolution Viking images [5]. Because of these data limitations, broad global generalizations cannot be made with absolute statistical certainty regarding all DTS features, many of which are almost certainly yet to be seen.

Nevertheless, the final dataset represents the best DTS collection currently possible. The dataset has proven to be reproduceable, given its striking similarity to the DTS best example selections made in the recent past by Ferguson [6]. The dataset, including numerous excellent examples of these important slope features, is entirely suitable for their preliminary examination.

The final correlation results and their compatibility with the proposed models are as follows:

<table>
<thead>
<tr>
<th>Results:</th>
<th>latitude</th>
<th>elevation</th>
<th>lithology</th>
<th>thermal inertia</th>
</tr>
</thead>
<tbody>
<tr>
<td>model 1)</td>
<td>s</td>
<td>n</td>
<td>n</td>
<td>s</td>
</tr>
<tr>
<td>model 2)</td>
<td>There is no correlation with lithology evident in the final dataset. Furthermore, there are strong correlations with other factors that are not of prime importance to this model.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>model 3)</td>
<td>There is a strong correlation with both latitude and thermal inertia, the two most important factors involved in this model. It is, however, important to recall that there is a bias in the Viking dataset that may be partially or entirely responsible for the strong correlation with latitude.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>There is a strong correlation with thermal inertia, this model's most important factor. Weak or questionable correlations with other factors are consistent with this model.</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Model 1 is not supported strongly by the correlation results. In addition, it has been suggested that a DTS process involving subsurface and surface inhomogeneities is unlikely, in that the amount of material necessary to create a large-scale slope feature is much greater than that which can be derived from a point source [1,2]. A possible exception to this conclusion may concern a process involving ultra-mafic igneous intrusions, or a process involving the weathering of exposed rock outcrops. These exceptions are not strongly supported by DTS morphology.

Model 2 is supported by the correlation results as well as the appearance of the DTS features in general. The model is weakened, however, by its extreme complexity as well as its hypothetical nature. The presence of near surface volatiles on Mars now, or in the recent geological past, has not yet been conclusively demonstrated.
The ALHA 77011 paired group is widely scattered when compared to the rest of the samples which is odd considering they are assumed to be from the same meteorite [7]. Some possible explanations for this are that the different samples were exposed at the surface of the ice sheet for different lengths of time, that size has a key roll in limiting weathering because smaller samples have a greater surface to volume ratio than larger samples, that degrees of fracturing might have had a major influence on how fast the sample weathered. An interesting point to note in all these graphs is that L3 chondrites display a much wider range than that seen in the L6 Antarctic chondrites.

When a comparison of the rust and $^{26}$Al [8] abundances are made no trends are apparent, (Fig 5). This would imply that terrestrial residence age does not appear to be the controlling factor in how weathered a sample is [9], although further analyses are needed.

CONCLUSIONS. In using the DSC, one concern is how reproducible the results are for replicate samples. Because of time constraints, multiple re-runs were performed only on the Holbrook samples. For Holbrook 1912, the variation in the troilite peak area after three runs is .01 J/g out of 1.63 J/g for 10-mg samples. The Holbrook 1931 was also run three times. The variation between runs in the troilite peak area for Holbrook 1931 is .6 J/g out of 1.7 J/g; the variation in the rust peak is 1.9 J/g out of 16.7 J/g. Unfortunately, in all cases 10 mg may not be enough to get a representative sampling from the homogenized powders. Sample ALHA77015 was run twice because of anomalous peaks, in the region of the rust peak, that were suspected to result from the dehydration of evaporite salts. During the second run, the anomalous peaks were not observed, possibly because the two samples varied in composition. There were also anomalies in the weathered Holbrook samples which were not reproduced in subsequent runs. For future experiments >20 mg samples may be required.

There is much more information in these DSC data than could be analyzed in the short time of this project. The subtraction curves have many smaller peaks than the troilite and rust analyzed here. Fig. 6 shows a typical first-heating/second-heating subtraction curve. Four weathering peaks are pointed out in the graph. More detailed analyses of the data collected for this project might significantly increase our understanding of how Antarctic meteorites weather.

INTRODUCTION. No quantitative scale for measuring terrestrial weathering of Antarctic meteorites exists. Knowing the degree of weathering of Antarctic meteorites is important for two main reasons [1]:

(a) Weathering products can have an undesirable effect on the results of cosmochemical studies.

(b) Accurate comparisons of weathering state versus terrestrial residence times, original meteorite mass, and other factors can be made to better understand what plays a major role in weathering.

The only classification scheme of weathering for Antarctic meteorites is the A-B-C scale which is a qualitative assessment of rustiness of by visual inspection of hand specimens. "A" represents very little rust while "C" indicates a high degree of rustiness. A subscript "e" is added to this classification if evaporite deposits are seen. Although the A-B-C system is better than ignoring weathering effects, a more quantitative method would be preferable.

The purpose of this project was to determine the usefulness of differential scanning calorimetry (DSC) in devising an objective weatherometer for Antarctic meteorites. Differential scanning calorimetry was chosen because it is quantitative and objective, reproducible, rapid (1-2 hr for an analysis), and uses small samples (10-20 mg). DSC is sensitive to known or suspected weathering products, including oxyhydroxides, salts and clays.

EXPERIMENTAL. There were two main phases in this project. First, point counts of meteorite thin sections were done with a reflected-light microscope to compile modal data for metals, rust, sulfides, and silicates. For each section, 400-1400 points were taken depending on the cross-sectional area of the slide. Where possible multiple slides were used and the weighted average of the areas were taken as representative of the meteorite. Next, a Perkin-Elmer DSC-7 differential scanning calorimeter was used to analyze powdered samples of the same meteorites prepared by Eugene Jarosewich [2] except for the Holbrook, Arizona chondrite powders which were prepared by Gooding [3]. The powdered samples were made from an average weight of about 3-20 grams of meteorite [2] which should approach representative sampling [4]. For each run 10-12 mg of homogenized powder was placed inside an aluminum oxide crucible and heated from 300-1000 K at 10 K/min under continuous purge of 20 cm$^3$ Ar/min. The samples were then cooled to room temperature, weighed, and then the heating cycle repeated. At the end of each day a zinc standard (NIST SRM-2221) was run for control of instrument calibration.

In Fig. 1 a typical plot of DSC data is illustrated. Here the Holbrook 1912, which was collected shortly after it fell, and the Holbrook 1931, which was collected 19 years after it fell, are plotted to show the differences between weathered and fresh samples. In this project I utilized two main DSC peaks. The troilite peak is a sharp peak at approximately 413 K; that represents the solid-state structure inversion of (pretropical) meteoritic FeS. The "rust" peak is the bell-shaped peak at about 500-520 K; that represents the dehydration of various ferric oxyhydroxides formed by terrestrial weathering. The integrated peak areas were used to determine quantities of troilite and rust present in each sample. For each sample, second-heating data were subtracted from first-heat data to remove heat-capacity effects and more clearly reveal peaks attributed to phase changes.

RESULTS. Fig. 2 shows a positive trend for modal rust versus Fe$_{\text{total}}$(Fe$^0$ + FeS). The modal rust is from the petrographic microscope analysis, the iron ratio was obtained from chemical analyses by Jarosewich [2] for Antarctic samples, or by Mason and Wiik [5], and Gibson and Bogard [6] for Holbrook samples. Fig. 2 confirms the expected trend; rust accumulates as metal and troilite corrodes. Labels from the A-B-C system are shown for respective Antarctic L3 meteorites for reference. Fig. 3 plots rust data compiled from DSC analysis versus iron ratio for comparison with the data from the petrographic microscope. From examining the graphs it is apparent that there is some reproducibility between the them. The data collected from the DSC is probably more accurate because it was collected from homogenized powders of representative samples. It is difficult to tell if the thin sections were representative of the whole rock. For Fig. 3a, a plot is done of weathered and fresh non-Antarctic meteorites. Fig. 3b shows the Antarctic L3 and L6 meteorites with pairing groups identified.

When weight loss is plotted against rust (Fig. 4) there is also a positive trend. One cause for this correlation is that the H$_2$O bound up in the hydrated rust is released during heating; the more rust present in the original sample the greater the observed weight loss. Fig. 4 notes the various paired groups of chondrites.
Carbon Isotopic Composition of Carbon Released from Stepped Combustion of Precambrian Sediments

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Advisor: Everett K. Gibson Jr., Planetary Sciences Branch, NASA, Houston, TX 77058
Special Assistance: Richard A. Socki, LESC, NASA Houston, TX 77058

INTRODUCTION

Various geochemical studies have indicated that the Archean atmosphere is markedly different from the current atmospheric composition (1). The earliest atmospheric constituents were probably stripped away during accretion; extensive outgassing (continuously from volcanoes or catastrophically from impact melting) created a secondary atmosphere that endured for the bulk of the Archean (2). Several studies have established it as largely anoxygenic. The rise of extensive life on the planet (ca. 2.0 Ga) encouraged the eventual change from a reducing or neutral (Archean) atmosphere to an oxidizing (modern) one. Both photodissociation and photosynthesis were involved in this change.

This work attempts to identify sources of carbon released at different temperatures by examining the carbon isotopic composition of carbon species in some Precambrian sediments. Previous work with argon isotope dating (3) has indicated that volatiles trapped in selected sediments represent the atmosphere present at time of formation; these volatiles are available only in the high temperature (>800°C) fraction. The lower temperature fractions are often fractionnated organic carbon or have been subject to metamorphic processes. Study of some Precambrian sediments over a broad span of time (1.5 Ga) may provide information about the oxidizing or reducing potential of the early atmosphere and clues to the origin of life.

EXPERIMENTAL PROCEDURE

Five Precambrian sediments with ages ranging from 1.6 to 3.0 Ga were ground and passed through a 124 micron mesh. 100 mg of each was weighed and allowed to de-gas in vacuum at 150°C overnight in a fused quartz tube. The temperature was then raised to 200°C and degassed for an additional 2 hours. Samples are then heated in steps of 200-400°C, 400-600°C, 600-800°C and 800-1000°C. The evolved gas was reacted with \( \text{O}_2 \) produced by heating CuO at 800°C. Each fraction was allowed to react over a platinum catalyst heated to 1000°C. Resultant \( \text{CO}_2 \) was collected in a liquid nitrogen cold finger and separated from water and non-condensible gases cryogenically. The \( \text{CO}_2 \) yield was measured with a capacitance manometer to ±0.01 torr.

All \( \text{CO}_2 \) was analyzed on a Finigan MAT 251 gas source mass spectrometer. All isotope data are reported relative to the standard PDB (4). The machine standard was NASA-I (\( \delta^{13}\text{C} \) (PDB) = -7.85 \( \%_o \)). Carbon standards extracted on the same line include NBS-21 (5) and NASA-2 (both graphite). Reproducibility of these standards was generally better than ±0.5\( \%_o \).

RESULTS AND DISCUSSION

Data for carbon isotope composition from stepped heating of five Precambrian sediments are presented in Table 1. Each sample resulted in 4 different temperature fractions; a weighted average was calculated on the four corresponding isotope ratios to produce a sample average. Carbon yields and \( \delta^{13}\text{C} \) data are plotted in Figure 1a-e as a function of temperature as are carbon yields as \( \mu \text{g} \) C/gram sample.

The youngest sample (McArthur Group, age 1.68 Ga, PPBP 106-1-RF) showed that the bulk of the carbon (96%) was released in the 400-600°C fraction and had a carbon isotope composition of -3.15 \( \%_o \). Carbon is most likely to be derived from carbonates (Mg,CaCO\(_3\)); its isotope ratio is characteristically heavier (\( ^{13}\text{C} \) enriched) than more reduced carbon. At higher temperatures the yield drops tremendously as the more reduced (or graphite) carbon is released; corresponding to a lighter ratio (-7.89\( \%_o \)). At the highest temperature fraction, 800-1000°C, the yield drops to 0.42% of the total and the carbon isotope ratio increases. This is uncharacteristic of the more reduced carbon we expect in this fraction. The weighted average for this fraction is -3.3\( \%_o \).
Spargur, C. S.: DSC Weathering Scale for Antarctic Meteorites

Figure 3 (a, top; b, bottom). Covariation of DSC integrated "rust" peak with bulk iron ratio.

Figure 4 (a, top; b, bottom). Covariation of DSC "rust" peak with sample weight loss during heating to 1000 K.

Figure 5. Covariation of DSC integrated "rust" peak with $^{26}$Al disintegration rate.

Figure 6. DSC data for weathered sample of Holbrook 1931 chondrite, partially corrected for heat-capacity continuum.
Table 1: C-13 and Yields From Stepped Combustions

<table>
<thead>
<tr>
<th>Sample</th>
<th>T(°C) Step</th>
<th>µg C/g Samp</th>
<th>C-13 (PDB)</th>
</tr>
</thead>
<tbody>
<tr>
<td>293-1-RF</td>
<td>200-400°C</td>
<td>1174</td>
<td>-7.74</td>
</tr>
<tr>
<td>502.49 mg</td>
<td>400-600°C</td>
<td>6998</td>
<td>-39.64</td>
</tr>
<tr>
<td></td>
<td>800-1000°C</td>
<td>4026</td>
<td>-41.67</td>
</tr>
<tr>
<td>298-1-RTS</td>
<td>200-400°C</td>
<td>883</td>
<td>-26.79</td>
</tr>
<tr>
<td>110.32 mg</td>
<td>400-600°C</td>
<td>42461</td>
<td>-8.61</td>
</tr>
<tr>
<td></td>
<td>600-800°C</td>
<td>528</td>
<td>11.48</td>
</tr>
<tr>
<td>C-13 = -9.09</td>
<td>800-1000°C</td>
<td>289</td>
<td>-21.6</td>
</tr>
<tr>
<td>106-1-RF</td>
<td>200-400°C</td>
<td>434</td>
<td>8.09</td>
</tr>
<tr>
<td>99.73 mg</td>
<td>400-600°C</td>
<td>31215</td>
<td>0.01</td>
</tr>
<tr>
<td></td>
<td>600-800°C</td>
<td>13341</td>
<td>-7.89</td>
</tr>
<tr>
<td>C-13 = -3.31</td>
<td>800-1000°C</td>
<td>424</td>
<td>-1.14</td>
</tr>
<tr>
<td>388-1-RTS</td>
<td>200-400°C</td>
<td>264.8</td>
<td>-12.85</td>
</tr>
<tr>
<td>100.23 mg</td>
<td>400-600°C</td>
<td>449.2</td>
<td>-6.6</td>
</tr>
<tr>
<td></td>
<td>600-800°C</td>
<td>215.8</td>
<td>-21.62</td>
</tr>
<tr>
<td>C-13 = -12.13</td>
<td>800-1000°C</td>
<td>22.4</td>
<td>-23.3</td>
</tr>
</tbody>
</table>

Figure 2
Bulk C-13 vs. Age (Ga)
The carbon release profile of sample PPBP 099, Earaheedy Group (age 1.6-1.7 Ga), is similar to that of the McArthur Group. A substantial portion (69%) of the total carbon is released in the 400-600°C range with an enriched carbon isotope ratio (+0.01‰). At the highest temperature fraction (800-1000°C) the yield is 0.93% of the total released; the lighter carbon isotope ratio (-1.14‰) suggests a trend toward reduced carbon.

Both the Transvaal (PPBP 298-1-RTS, age 2.67 Ga) and Ventersdorp (PPBP 293-1-RF, age 2.64 Ga) have similar carbon release and isotope profiles. At lower temperatures (<600°C) the bulk of the carbon is released. Carbon isotopes (Transvaal, -8.61‰; Ventersdorp, -10.72‰ for 400-600°C) suggest release of carbonate carbon; these relatively heavier ratios belies oxidized carbon. As with previous samples, the yield and carbon isotope ratios decrease with an increase in temperature, suggesting the release of reduced carbon.

The oldest sample, Gorge Creek, (PPBP 388-1-RTS, age 3.0 Ga) also had its highest yield at 400-600°C, and a corresponding heavier carbon isotope composition (-6.60‰), again suggests carbonate. At the higher fractions the carbon isotope composition drops to -21.62‰. Again, we see a shift from carbonate-released carbon (heavier ratios) to carbon releases from reduced carbon.

Carbon in the unaltered Precambrian sediments comes off in two fractions: as carbon from carbonates, at temperatures less than 600°C, where the heavier carbon isotope ratios are indicative of oxidized carbon. The second fraction, that of reduced (or "organic") carbon appears distinctively in the last fraction, from 800-1000°C. Carbon more extensively reduced will have a heavier carbon isotope composition. Previous studies (6) used HCl to wash away carbonate carbon which resulted in much lighter carbon than are represented here. As well, previous investigators did not use step combustion on their sediments; instead they heated the volatile carbon at a fixed temperature. All the weighted averages have been plotted against the age of the sample (Figure 1a-e). It is apparent that over time the carbon isotope ratio gets heavier (\(\delta^{13}C\) enriched). The younger samples (McArthur and Earaheedy Group) contain heavier carbon isotope ratios at high temperature suggestive of trapped atmosphere components. More reduced carbon, characterized by lighter \(\delta^{13}C\) isotopes appear in greater abundances in the older samples.

**SUMMARY AND CONCLUSIONS**

Our weighted average \(\delta^{13}C\) values used agree favorably with literature values when corrected for absence of carbonate. As well, these weighted averages, when plotted against the age of the sample, might reveal facts about the earlier atmosphere (Figure 2). The lighter carbon isotope ratios, seen in older sediments, are associated with reduced carbon, a result of a reducing atmosphere. The heavier carbon isotope ratios are indicative of oxidized carbon from an oxidizing atmosphere. Previous studies have shown that reduced carbon becomes heavier (from values as low as \(\delta^{13}C\) (PDB) = -50) and eventually stabilize around \(\delta^{13}C\) (PDB) = -15 around 1.0 Ga (6). Our data is heavier still because it was not washed with HCl to remove the heavier oxidized carbon. A clear shift is noticed at about 2.0 Ga on the plot; it should be noted that at this point in evolution marked the widespread appearance of eukaryotic organisms. A eukaryotic dominance over prokaryotes ensured greater production of oxygen and a change from a reducing atmosphere to an oxidizing one (7). Clearly more unaltered sample should be analyzed to help constrain these observations.

**REFERENCES CITED**

Chemical Compositions of IDPs. Koichi Takatori

Carbon abundance for these IDPs varies from about 3 wt. % to 20 wt. %. The C values are much higher than any of the chondrite groups, especially particles L6 and P9, which have approximately 15% carbon. Only particle P2 has a carbon value comparable to the CI chondrites (Table 1). The C abundances for the 6 particles are similar to previous estimates for hydrated IDPs [1].

Conclusions
I analysed 6 IDPs, and determined their chemical composition. They have approximately chondritic compositions and the element abundances resemble both the carbonaceous and ordinary chondrites. Carbon abundance in the 4 particles is typically greater than that for chondrites, but is similar to other smooth (=hydrated) IDPs. Two particles contain abundant carbon (about 15wt%). Studies are in progress to determine the mineralogy and petrology of these particles, and to determine the host phase(s) for the C.

Acknowledgements
Lindsay P. Keller kindly provided me with his samples and I am deeply grateful to him for his guidance and patience. Kathie L. Thomas assisted in the reduction of EDS analyses, and collected many EDS data of standards. I also thank David S. McKay and Michael E. Zolensky for their kindness.

References

TABLE 1. Carbon abundance and element/Si ratios (in wt.%) for six IDPs and literature values for chondrites and other IDPs.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Ref.</th>
<th>C wt%</th>
<th>C/Si</th>
<th>Na/Si</th>
<th>Mg/Si</th>
<th>Al/Si</th>
<th>S/Si</th>
<th>Ca/Si</th>
<th>Fe/Si</th>
<th>Ni/Si</th>
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</thead>
<tbody>
<tr>
<td>L2005R1</td>
<td></td>
<td>6.48</td>
<td>0.78</td>
<td>0.10</td>
<td>0.70</td>
<td>0.05</td>
<td>2.16</td>
<td>0.10</td>
<td>4.51</td>
<td>0.11</td>
</tr>
<tr>
<td>L2005P13</td>
<td></td>
<td>7.28</td>
<td>0.49</td>
<td>0.02</td>
<td>0.68</td>
<td>0.08</td>
<td>0.17</td>
<td>0.25</td>
<td>1.76</td>
<td>0.07</td>
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<tr>
<td>L2005P9</td>
<td></td>
<td>19.57</td>
<td>1.88</td>
<td>0.08</td>
<td>0.80</td>
<td>0.12</td>
<td>0.25</td>
<td>0.00</td>
<td>2.21</td>
<td>0.06</td>
</tr>
<tr>
<td>L2005L6</td>
<td></td>
<td>13.52</td>
<td>1.20</td>
<td>0.08</td>
<td>0.93</td>
<td>0.06</td>
<td>0.25</td>
<td>0.10</td>
<td>2.14</td>
<td>0.18</td>
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<tr>
<td>L2005P2</td>
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<td>3.13</td>
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<td>0.10</td>
<td>0.66</td>
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<td>1.21</td>
<td>0.06</td>
<td>11.90</td>
<td>0.25</td>
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<tr>
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<td>0.61</td>
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<td>0.00</td>
<td>0.97</td>
<td>0.03</td>
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<tr>
<td>CI(bulk)</td>
<td>[3]</td>
<td>3.45</td>
<td>0.32</td>
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<td>0.89</td>
<td>0.10</td>
<td>0.53</td>
<td>0.10</td>
<td>1.72</td>
<td>0.09</td>
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<td>CI(matrix)</td>
<td>[4]</td>
<td>--</td>
<td>--</td>
<td>0.02</td>
<td>0.80</td>
<td>0.09</td>
<td>0.15</td>
<td>0.02</td>
<td>1.07</td>
<td>0.10</td>
</tr>
<tr>
<td>CV(matrix)</td>
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Chemical Composition of Hydrated Interplanetary Dust Particles (IDPs) and Comparison to Chondrites
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Introduction
Hydrated Interplanetary Dust Particles (IDPs) are primitive materials that are probably related to chondrites and other chondritic materials in their origin and probably are samples of a much broader range of parent bodies than chondrites [1]. We have many chemical and mineralogical data of chondrites, but not so many of IDPs because of the difficulty in handling. I examined a total of 6 IDPs and determined their chemical composition quantitatively.

Analytical Methods
The 6 IDPs were analyzed in a JEOL JSM 35CF scanning electron microscope (SEM) using a PGT System IV thin-window energy dispersive spectrometer (EDS) for elements from C to Ni. The particles were mounted on beryllium substrates and analysed using a 15kV rastered electron beam for 500 seconds. Diamond was used for a carbon standard and 21 natural minerals were used for elements (Z>8). Standard spectra were collected at 15kV for 300 seconds. Relative errors for major elements is < 10%. We estimate C errors to be < 20% relative. The particle data were reduced using the procedure of Armstrong and Buseck [2]. The data for carbon were basically from the surfaces, and contamination from silicone oil (collecting medium) may contribute slightly to the total carbon determination.

Results
The particles in this study (L2005L6, P2, P9, P13, R1, and R7) have smooth surfaces and are approximately 10 to 20 µm in diameter (Figure 1). Carbon abundances and element/Si ratios for the 6 IDPs are given in Table 1. Table also lists data for bulk CI chondrites and values for chondrite matrices from the literature. The element/Si ratios for the 6 IDPs are within a factor of 2 of CI abundances for major elements listed above, except for two particles, L2005P2 and R1. I infer from the analyses of P2 and R1 that Fe-Sulfide is a major component of P2, while magnetite or some Fe-rich phase is dominant in R1. For elements other than Fe and S, and possibly Ni, particles P2 and R1 have approximately chondritic abundance ratios.

Discussion
The data in Table 1 show that none of our particles is an exact match for any chondrite group, but most element/Si ratios overlap with the range of values exhibited by the carbonaceous and certain ordinary chondrites. Specific element/Si ratios are considered below.

• Mg/Si

The Mg/Si ratio in the 6 IDPs varies from 0.61 to 0.93, while the ratio for carbonaceous chondrites varies from approximately 0.8 to 0.9, and that for UOC (Semarkona matrix) is about 0.4. Four of the IDPs have Mg/Si ratios that are intermediate between carbonaceous and ordinary chondrites, suggesting that these particles have similarities to both carbonaceous and ordinary chondrites.

• S/Si

For the 4 IDPs that are not dominated by Fe-Sulfides, the S/Si ratios for these IDPs are comparable to the values for carbonaceous chondrites, but are generally much lower than that reported by Schramm et al.[1]. The S/Si ratios for the 4 IDPs are most similar to values for CI matrix.
Distribution of Sodium in the Atmosphere of Mercury
and
The Search for a Lunar Dust Cloud

KEITH VANDEN HEUVEL
Lunar and Planetary Institute Summer Intern

Sodium on Mercury

INTRODUCTION

Sodium vapor was first observed in the atmosphere of Mercury in 1985 (1). Since then many observations have been made in an attempt to understand the distribution of the vapor and the processes involved in its production. Observations of Mercury were taken during the months of December 1990, and February, March, and June of 1991 using the McMath main solar telescope and stellar spectrograph of the National Solar Observatory at Kitt Peak, AZ using a Bowen image slicer (2) in order to image the distribution of sodium at the time of observation. Many observations were taken to monitor day-to-day variations in order to correlate sodium distribution to any changes in solar activity for a possible relation. It is thought that the sodium vapor may be introduced into the Hermean atmosphere by sputtering of the sodium from the surface by ions trapped in the magnetosphere and by \( \text{Na}^+ \) being transported along the magnetospheric field lines toward the higher latitudes (3).

FORMATION OF SODIUM DISTRIBUTION IMAGES

The first step in the analysis of the observations was the formation of images of the spatial distribution of sodium. Since the observations were taken with an image slicer, both spatial and spectral dimensions were recorded on the CCD. This also meant that only part of the chip contained the actual image of Mercury, thus it had to be located. This was done using IRAF by plotting and then enlarging the section of the chip that contained the image. Once these pixel values were obtained, the image was plotted for spectral analysis and the location of the \( D_1 \) and \( D_2 \) sodium emission features were recorded.

The IRAF files were then converted into ASCII files for creation of the images. Once in ASCII format, the maximum continuum values of each observation were obtained so that this value could be compared to the emission features at every location within the image slicer's field of view. New files were then created that contained arrays whose elements corresponded to intensity values of emission features compared to continuum. These arrays were then fed into a LIPS routine which converted the data arrays into monochromatic images. The images were 12 x 7 pixels in size with each pixel corresponding to an angular area of 0.43 x 0.43 arc sec.

The angular size of Mercury is such that the entire observable surface will not fit on the area covered by the image slicer. Thus, several images were taken at each observation time corresponding to northern, southern, and central regions of the planet so that the images could be combined later to create a full view of the observable disc. Using LIPS, the images of a single day were combined and the overlapping pixels, if any existed, were averaged together to yield a continuous image of the sodium distribution over the surface of the observed surface. Line drawings were made of the phase of Mercury at the times the observations were made. These drawings were digitized and superimposed on the images to get a better understanding of the true location of the sodium vapor. False color was also used to enhance the images so that changes in intensity were more noticeable.

RESULTS AND CONCLUSIONS

Observations from the dates December 4-10, 1990, February 24, March 2, and June 6, 1991 were all taken but only some of them could be made into usable images. Images from Dec. 5, 6, and 9 were made as well as from Feb. 24 and June 6. All three of the December images show a definite concentration of sodium vapor at the high northern and southern latitudes, consistent with the assumption of magnetospheric processes being the cause of the sodium vapor. The February image, however, does not show such a strong high latitude concentration and the June image, due to the fact that it is incomplete, could be interpreted either way.
FIGURE 1. SEM micrographs of the six particles from the large-area collector L2005.
Mercury Sodium Vapor and Lunar Dust
KEITH VAN DEN HEUVEL

anticipated result was not observed. Instead of a sharp increase in blue light at some distance from the terminator, there was a steady increase in blue light all along the slit with no apparent sudden increase or decrease at any point near the terminator.

Polarization was then used to look for some effect of the proposed dust cloud. As was mentioned above, the light scattered by dust should be somewhat polarized. If there is dust near the terminator, this effect may be observable. The slope plots described above were used in the search for this effect. Both parallel and perpendicular observations of the same site were plotted on the same axes to see if any differences existed. Light scattered by the dust grains would be polarized. Observations made with the polarizer set at 90° should show a larger difference in slope near the terminator than one with the polarizer set at 0°. The plots should separate only at a point somewhat close to the terminator, where polarized, dust-scattered light is present.

CONCLUSIONS TO DATE

Analysis of the polarization plots did not reveal convincing evidence for a lunar dust cloud. Although many of the plots show the light near the terminator to become polarized at a definite distance, there are aspects of the data that are not consistent with suspended dust grains. The polarization occurs most noticeably at high latitudes. Many high latitude observations were made on the last evening with 76% of the lunar disc illuminated. Thus, these observations placed a large portion of the observing slit off the surface of the moon. When these plots were made it could be seen that the polarization effect occurred in the area of the slit which was not over the lunar surface. Therefore, this polarization effect could not be caused by suspended dust grains. It is possible that the moonlight itself may be polarized, but analysis of bright lunar limb observations gave no insight into this matter, as was hoped.

The fact that there is a higher concentration of shorter wavelength light as one approaches the terminator is likely due to moonlight scattered by aerosols and dust in the Earth's own atmosphere as opposed to a strictly lunar phenomena (6). Therefore, light scattered from dust grains may have been observed, but light scattered in the Earth's atmosphere seems to be more intense, thus dominating any dust scattered light.

Other methods of analysis are currently being developed and there is still a possibility that these observations may yield some conclusive information. Simpler analysis techniques will be used in an attempt to measure the amount of polarization along the slit as well as comparisons between wavelength intensities at different distances from the terminator. More observations of the moon at phases prior to and including first quarter would undoubtedly improve our knowledge of the scattered moonlight in the Earth's atmosphere as well as the question of polarized moonlight.

REFERENCES AND NOTES

6. H. A. ZOOK, Personal Contact.
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One way to get a better understanding of the sodium distribution is to compare it to the distribution of potassium, also present in the Hermean atmosphere (4). This was anticipated and observations of the potassium D\textsubscript{2} emission line were also made on each of the December dates. The observations were made in the same fashion as were the sodium observations. Due to the fact that the potassium emission feature is much weaker than the sodium emission lines, good images have not yet been made of the potassium distribution. Work is currently being done, however, to solve this problem.

The December images give strong support to a sodium atmospheric component being driven by magnetospheric processes. Images from the February and June data, although not entirely supportive, do not refute the above claims due to the fact that there may yet be an as of yet undetected solar activity correlation. More study in this area is required before any conclusions can be drawn, however. Further work on the potassium images is continuing and analysis of these images may strengthen the magnetospheric theory of sodium distribution.

The Search for Lunar Dust

INTRODUCTION

It has been suggested (5) that a large scale lunar dust cloud may exist at or near the terminator as an explanation of particular lunar phenomena. This dust has yet to be verified conclusively, but observation of the dust may be possible from Earth-based telescopes. Observations were made on April 22-26, 1991 at McDonald Observatory of the University of Texas using the 2.1 meter telescope with a Cassegrain spectrograph, TI 1 CCD, and low-resolution grating. A polarizing filter was used so that each of the sites observed could be recorded through three different polarization angles: 0\degree, 45\degree, and 90\degree. Observations were made by placing a slit parallel to the lunar equator on the unilluminated side of the moon such that the far edge of the slit was just west of the terminator (Figure 1). The spectrograph thus recorded spectra all along the length of the slit in order for variations in spectral features or intensity to be recorded as a function of distance from the terminator.

The proposed dust cloud requires the dust grains involved to be on the order of 0.1 micron or smaller (5). Thus, any light scattered by the dust cloud should be Rayleigh scattered, scattering shorter wavelengths preferentially over longer wavelengths. Also, the light should be polarized to some degree, depending on the phase angle. This polarization effect and the effects due to the Rayleigh scattering were looked for in the analysis of the data obtained.

DATA ANALYSIS TECHNIQUES

Soon after the observations were made, several methods for analyzing these data were used. These methods attempted to highlight the effects of Rayleigh scattering and of the polarization effect. In order to see these effects these data had to be analyzed both spatially and spectrally.

Rayleigh scattering was looked for first. This was done by looking for a greater change in blue light as the terminator is approached than in other wavelengths. For each observation, the 10 spectra that represented the position furthest away from the terminator along the slit, were averaged together into one spectrum to eliminate any noise anomalies. The spectral resolution was also reduced by a factor of 20, in order to increase signal to noise ratio. The values representing the far end of the slit were then compared, for every spectral region, to each spectral value along the length of the slit until the opposite, or terminator, end was reached. Each of these comparisons consisted of a ratio of the spectral values representing the different distances along the slit, to the spectral values representing the far end of the slit. Each ratio was written to a file and these data were then plotted. Several different plots, each representing a different distance along the slit, can be seen on the same axes. In this way, it could be seen if there was an increase in shorter wavelength light, represented by a ratio greater than one, as the terminator was approached. The longer wavelengths should have a ratio close to one. If this actually did occur, as it did in several of the observations, it would be consistent with the proposed dust theory. However, a better analysis would be required to ascertain if the so called "bluing" effect was caused by dust or by some other physical process.

If dust was the cause of the "bluing" as the terminator is approached, there should be a point along the slit at which this effect becomes first noticeable. This is due to the fact that the dust cloud must have some finite scale height and therefore the scattered light would be unobservable once a certain distance from the terminator is reached. Therefore, if one takes the slope of the plots described above, it can be seen where the greatest increase in scattered blue light occurs. These slopes were computed and the resulting plots of slope vs. distance were analyzed. The
The magnetohydrodynamic equations governing the evolution of the magnetic fields can be reduced to the following form by the so-called local approximation [4]

$$\gamma a = \alpha b + \frac{\partial^2 a}{\partial z^2}$$

$$\gamma b = -D_{\text{eff}} \frac{\partial a}{\partial z} + \frac{\partial^2 b}{\partial z^2}$$

where $a$ and $b$ are the vertical distribution of the vector potential and the azimuthal component of the magnetic field, respectively, $\alpha$ describes the turbulent helicity, $\gamma$ is the local growth rate of the field and $D_{\text{eff}}$ is radially varying, dimensionless dynamo number. The dynamo number characterizes the strength of magnetic field regeneration mechanisms in comparison with diffusive dissipation and can be expressed in the form [4]

$$D_{\text{eff}} = \frac{\alpha \omega h^3}{(\eta_t + \eta_r)^2}$$

where $\omega$ is the angular velocity of the large-scale fluid motion, $h$ the half-thickness of the disk, $\eta_t$ magnetic diffusivity and $\eta_r$ magnetic resistivity, which depends on the ionization degree of the nebula. For a decaying magnetic field the local growth rate is negative. The value of the dynamo number for which the real part of $\gamma$ reaches zero is called the critical dynamo number, $D_{\text{crit}}$. Only in those parts of the nebula where $D_{\text{eff}}$ exceeds the critical value, the regeneration mechanism is effective enough to overcome the dissipative effects, and the magnetic field can exist for a relatively long period of time. If the imaginary part of the growth rate is zero, the field is stationary; a non-zero value produces an oscillating field.

The magnetic field most readily excited in nebular conditions is of quadrupole symmetry, for which $D_{\text{crit}} = -10$ and the imaginary part of the growth rate is zero. The local dynamo number as a function of radial distance from the protosun can be seen in Fig. 2. The distance at which the dynamo number exceeds the critical value is very

![Fig.2. The local dynamo number as a function of radial distance from the protosun for two accretion rates and a variety of turbulent Mach numbers ($M_t = 0.05, 0.1, 0.3, 0.5$). The top of the plot marks the present positions of the planets and the asteroid belt.](image-url)
MAGNETIC FIELDS IN THE SOLAR NEBULA
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Magnetic fields are generally thought to have played an important part in the formation of the solar system. The measurements of magnetic fields of the carbonaceous chondrites, which are supposed to be unaltered relics from the formation era, indicate that magnetic fields with intensities of 0.1-1 G were present during this period [1,2]. The Lorentz stress associated with these fields could have been a major factor in determining the structure and dynamical balance of the nebula. Magnetic fields compressed during the collapse of the parent molecular cloud could have reached field strengths of the order of 1 G, but they would have decayed rapidly due to the probable presence of turbulent motion [3]. However, the very motion that is responsible for the diffusion of the field can also regenerate it through the turbulent magnetohydrodynamic process. The objective of the present study was to determine where in the protoplanetary disk magnetic fields generated by this process were likely to exist long enough to affect the dynamics of the nebula. Also, special attention was paid to how sensitive the generation of the field is to the features of a particular solar nebula model.

A steady state model for the solar nebula was constructed using the general accretion disk theory and a specific opacity law where the opacity is dominated by silicate and Fe metal grains in the inner parts and by ices in the outer parts of the nebula. The turbulent Mach number, $M_t$, and the accretion rate, $\dot{M}$, were included as free parameters in order to enable the study of several possible disk models, which may correspond to different stages in the evolution of the nebula. Properties governing the structure of the nebula - temperature, gas density, surface density and the half-thickness of the disk - varied considerably from one choice of free parameters to another, but in all reasonable cases the disk was found to be cool except for the immediate vicinity of the protostar. The coolness of the disk implies that thermal ionization could not have been an important source of charged particles in planetary distances. In the present study ionization was therefore thought to be due to the combined effect of cosmic rays and radioactive elements $^{26}$Al and $^{40}$K. The loss of charged particles was taken to be due to recombination on grains and to electrons reacting with positive ions. The ionization degree, $x$, which is the ratio of the number density of electrons to that of neutral hydrogen atoms, was calculated at different radial distances from the central object. Near the protosun the surface density is high and cosmic rays cannot penetrate through the medium to the equatorial regions of the disk. In these areas the ionization degree is determined by radioactive isotopes, especially $^{26}$Al while at larger distances the main source is cosmic rays. The dominant source of ionization depends also on the model of the disk (Fig. 1); for models with accretion rates smaller than $1.0 \, M_\odot/10^6$yr and turbulent Mach number larger than 0.1, the ionization degree is completely determined by the cosmic rays at all planetary distances.

Fig. 1. Ionization degree for a wide variety of turbulent Mach numbers and for accretion rates $1 \, M_\odot/10^6$yr and $0.1 \, M_\odot/10^6$yr at the distance of 3 AU. The different lines correspond to ionization by cosmic rays (CR) and radioactive isotopes $^{40}$K and $^{26}$Al.