Papers Presented at the

13th annual

Summer Intern Conference

August 14, 1997
Houston, Texas

1997 Summer Intern Program for Undergraduates
Lunar and Planetary Institute
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Thirteenth Annual
SUMMER INTERN CONFERENCE

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Sponsored by
Lunar and Planetary Institute
NASA Johnson Space Center
THIRTEENTH ANNUAL SUMMER INTERN CONFERENCE PROGRAM

AUGUST 14, 1997—LPI LECTURE HALL

Morning Session
Chair: Paul D. Spudis

9:00 a.m.  ROSS BEYER  Advisor: Deborah Domingue
Sixteen Years of Ultraviolet Observations with IUE on the Satellite Io:
Implications for Volcanic and Resurfacing History

9:20 a.m.  PAUL D. COX  Advisor: Michael B. Duke
Contributions of Impact Ejecta in the Preservation of Ice Deposits on the Moon

9:40 a.m.  TANYA DI VALENTIN  Advisor: Michael E. Zolensky
Characterization of Iron-Nickel Sulfides in the Chondrites

10:00 a.m.  Break

10:15 a.m.  RYAN C. EWING  Advisor: Paul M. Schenk
Stereotopography of Debris Aprons and Related Flow Features: Mars

10:35 a.m.  NANCY K. FORSBERG  Advisors: Robert R. Herrick and Benjamin Bussey
The Effects of Impact Angle on the Shape of Lunar Craters

10:55 a.m.  KATE GRAHAM  Advisor: David S. McKay
Scanning Electron Microscopy of Yellowstone Thermal Spring Travertines

11:15 a.m.  Lunch
Afternoon Session
Chair: Gary E. Lofgren

1:00 p.m.  KAREN JAGER  Advisor: Carlton C. Allen
JSC Mars-I: A Martian Regolith Simulant

1:20 p.m.  MUTSUMI KOMATSU  Advisor: Arch Reid
LL Chondrites and Prior's Rules

1:40 p.m.  KACPER KORNET  Advisor: Tomasz Stepinski
Global Dust to Planetesimals Transition in Viscous Solar Nebula

2:00 p.m.  Break

2:15 p.m.  SARAH NOBLE  Advisors: Gary E. Lofgren
Chondrule Precursor Aggregates in UOC's: Melting History

2:35 p.m.  MATTHEW E. PRITCHARD  Advisor: Walter S. Kiefer
The Effect of Lithospheric Rheology on Mantle Convection on Mars

2:55 p.m.  BRADLEY J. THOMSON  Advisors: Paul D. Spudis and Benjamin Bussey
Impact Craters as Probes of the Lunar Crust

3:15 p.m.  Adjourn
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Sixteen Years of Ultraviolet Observations with IUE on the Satellite Io: Implications for Volcanic and Resurfacing History

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Introduction

Since the Voyager spacecraft detected volcanic activity on Io in the late 1970's (Smith et al. 1979), many scientific instruments have been turned towards this enigmatic satellite. Io's volcanic and thermal activity has been observed with ground-based and spaceborn telescopes since that time (see Veeder et al. 1994 for a review of the literature), and spectra of Io's surface have been used to map the compositional variations on the surface (Nelson et al. 1980, Howell et al. 1984, McEwen et al. 1988). Observations over the past twenty years in different wavelength bands have determined that the surface is primarily composed of sulfur and sulfur dioxide. The International Ultraviolet Explorer (IUE) has taken spectra of Io during 1979-1981, 1984-1986, and in 1996. We analyzed these data between 2600Å and 3350Å. These three data sets, which span sixteen years, and were taken by the same instrument, have allowed us to examine large-scale compositional variations on Io's surface over a distribution of longitudes over decade time scales.

Methodology

The raw photon count data were calibrated into flux units by the IUE project data processing group and the corrected for the camera response function by using a triangular smoothing function. Flux values were converted into geometric albedo values by dividing the IUE spectra by corresponding solar spectra and correcting for distance variations (from the Sun and Earth) and solar phase angle differences. The solar spectra used were measured by the SME and SOLSTICE instruments at comparable wavelengths and on similar dates (B. Knapp, private communication). Errors were computed based on the observational errors for the satellite spectrum, the solar spectrum, the distance measurements, and the solar phase angle calculations (Bevington and Robinson 1992).

We then divided the planet into 30° wide longitude bins centered on the leading and trailing longitudes, in order to co-add nearby data together in a meaningful way (Note that even though we use the observation's sublongitude to assign longitude bins, we realize that the majority of the light detected is coming from a roughly circular patch with a radius of ~30° centered on the observation’s sublongitude and ~0° latitude). Each bin containing more than one spectrum was analyzed for variations amongst the spectra, and upon finding no significant differences, the spectra were averaged together to yield a representative spectrum for that longitude region. We have no spectra for the 345°-15° bin, the 195°-225° bin, and the 315°-345° longitude bins. This still provides adequate coverage on the leading and trailing hemispheres during each of our three observation epochs. Once the data for each bin for each epoch were averaged, error bars were calculated based on the variation between the spectra in each bin and their measurement errors. We compared the averaged spectra of each bin to look for spatial variations within an epoch and for temporal variations within each bin between epochs.

Results

Longitudinal Variations: The data from all three observational epochs show similar spectral characteristics. The leading hemisphere spectra display a steep absorption slope where the albedo rises with wavelength from 3100Å to 3350Å. In contrast, the trailing hemisphere spectra display more shallow absorption slopes in this wavelength range. The spectra between 217° and 255° have a slope less steep than that of the spectra in the leading hemispheres, while the spectra between 271° and 315° show an even more shallow slope indicating a greater absorption in this region of Io. Nelson et al (1980) find that this steep absorption feature is characteristic of the laboratory spectra of sulfur dioxide frost. The steepness of this absorption slope is a relative measure of the abundance of SO₂ frost. Nelson et al (1980) conclude from their analysis of the 1979-1986 IUE observations that the longitude regions 72° to 137° on the leading hemisphere have
the greatest \( \text{SO}_2 \) frost relative abundance. They attribute the more gradual slopes of this feature in the spectra of the trailing hemisphere regions to the larger relative abundance of sulfur allotropes mixed in with the \( \text{SO}_2 \) frost. We arrive at the same conclusions with our analysis of the extended IUE data set. \( \text{SO}_2 \) frost has a greater relative abundance from 29° to 156° longitude. Similarly we find that the absorption slope in the spectra of the trailing hemisphere longitude regions are shallower (Figure 1) indicating a greater relative abundance of sulfur allotropes mixed with the \( \text{SO}_2 \) frost. Like Nelson et al (1980) our spectra only allow us to determine relative abundance. Our data suggest that \( \text{SO}_2 \) frost is most abundant on the leading hemisphere, and appears to begin being contaminated by something with a higher UV absorptivity between 3100Å and 3350Å (most probably allotropic sulfur) between 217° and 255°. Then, between 271° and 315° we find that this absorption is very strong and that there is either a lesser degree of \( \text{SO}_2 \) frost or a high degree of allotropic sulfur at these longitudes. As mentioned above, the spectra for all three epochs display this same behavior, however, there is always an exception. There is a single spectrum centered at 156° taken in 1986 that displays a shape similar to the leading hemisphere data but with a characteristically higher albedo value of about 2.5% at all wavelengths (Figure 2). The volcanic feature Prometheus is located at about 152° longitude and this spectra may be evidence that Prometheus was thermally active at the time of this observation.

**Temporal Variations:** In our longitude bins from 45° to 165° there is no significant change in the spectra from epoch to epoch, which indicates (with the exception of the spectra at 156° possibly indicative of activity at Prometheus) that there has been no large-scale compositional changes over sixteen years on the leading hemisphere of Io. Similarly the longitude bins from 225° to 315° also show no changes in their spectra greater than one sigma over the three epochs. It is important to note that some of the historically most active features like Loki and Pele are in the trailing hemisphere and that their activity has not altered the large scale composition of the surface materials over sixteen years.

**Conclusions**

Our analysis confirms the longitudinal variation in composition seen by Nelson et al. (1980) in the ultraviolet and Howell et al. (1984) in the infrared. Namely, that \( \text{SO}_2 \) frost is more abundant on the leading hemisphere of Io than on the trailing hemisphere. Sulfur allotropes are more abundant in the trailing hemisphere which increase the absorptivity seen between 3100Å and 3350Å. More specifically, the spectra centered between 217° and 255° display a slope characteristic of an \( \text{SO}_2 \) frost abundance somewhere between that of the leading hemisphere and the longitudes farther west and the spectra centered on longitudes 271° to 315° show the highest abundance of sulfur allotropes observed in the spectra of Io. Our analysis also shows that there has been no significant change in the spectra of Io at any longitude during our three observation epochs, which implies that no large scale compositional change has occurred on Io in sixteen years. Therefore, despite the vigorously active volcanic activity on Io, and the proposed resurfacing events, whatever is being deposited has the same UV spectral characteristics as the underlying materials on a large, global scale.

**References**

Comparison of Leading and Trailing Hemisphere Spectra from the 84-86 Data

Figure 1. Examples of differences in the SO₂ frost absorption slopes of the leading versus trailing hemispheres.

Possibility of Prometheus Activity in 1986

Figure 2. Higher albedo values across the entire wavelength range are seen in the 1986 data for the region centered at 156° longitude. This brightening may be evidence for a short-lived thermal event. The thermally active volcano, Prometheus, is located at ~152° longitude.
Contributions of Impact Ejecta in the Preservation of Ice Deposits on the Moon

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A Monte-Carlo model is being developed to examine the extent to which meteoroid impacts could protect a layer of ice deposited by a comet into a lunar cold trap. The results will be helpful in determining possible configurations by which ice may be present on the moon.

INTRODUCTION

In the early sixties the idea that low temperatures of the permanently shadowed areas of the moon could have existed over geologic time scales suggested that volatiles, especially ice, might have become trapped there [1]. The value of the presence of water on the moon quickly sparked a few investigations into this subject. These studies attempted to establish the removal and deposition mechanisms present in these areas. Thirty years later a conglomerate of mechanisms exists. Still, there is no definite grasp on the magnitude of the accepted mechanisms. The division caused by this pool of theories has heightened since the recent controversial experimental data acquired by the Clementine spacecraft [2] and the seemingly conflicting data from the Arecibo Observatory [3].

Cold traps on the moon are subject to water deposition, water removal, and mechanical processes. Figure 1 shows the values for these mechanisms and their respective uncertainties. Water sources on the moon include meteoritic water, cometary water, and the products of Fe++ reduction in the regolith. The flux of meteoritic and cometary water deposited on the moon was calculated by [4] and [5]. Values for the meteoritic flux differ because they assume different mass fluxes and water composition in the metoroid population. Removal of water occurs by dissociation caused by H Ly-α radiation [5] and sputtering due to high-energy solar wind particles [6]. Sublimation also removes ice from the surface. This process is dictated by the vapor pressure of that ice, which is a function of temperature [7]. Vaporization by meteoroid and micrometeoroid impact is another removal process. There are no experimental determinations of the amount of ice vaporized by meteoroid impact due to the complexity of experiments in low pressure environments [8]. In order to determine the total flux of incoming water to the cold traps, migration of particles from the non-shadowed portions of the moon must be modeled. This is a statistical process in which water molecules deposited at random locations of the moon travel along ballistic trajectories with mean lifetimes proportional to absorbed radiation [4]. Additional mechanisms in the cold traps are considered negligible for the scope of this paper [9,10,11].

The repeated ejection of regolith from surrounding meteoroid impacts could shield a layer of ice formed by a comet impact. The theory of buffering due to regolith coverage has been shown by [12] in the area of megaregolith development. Also, [5] and [13] have both mentioned that if water were found in the cold traps it would be “sandwiched” between layers of regolith. The goal of this Monte Carlo model is to quantify the amount of ice that is protected by impact gardening as a function of time.

Monte Carlo Model

The model assumes a comet deposits a layer of ice on a regolith surface inside a cold trap. Because the time between cometary impacts suggested in [4] and [5] are different, an input parameter is used to set the desired time. Initial calculations assume that the comet layer was deposited 250 my ago. The model considers two elements: (1) the deflation of exposed ice and (2) erosion and covering of ice by impact. The deflation rate may be varied and is applied to any uncovered ice surface. The erosion and deposition rates are determined from data describing the flux of meteoroids on the moon, crater shapes and ejecta distributions.

The simulation consists of an array consisting of a layer of ice — whose thickness is an input parameter — on top of a layer of regolith. The size of the array is 1000 x 1000 x 100 units in the x, y, and z directions, respectively. The resolution is 10 cm along the x and y coordinate axes and 1 cm in the z giving a total size of 100 x 100 x 1 m. A larger area would be desirable but is not feasible computationally at the resolution needed to model the craters used in this simulation. The array is bombarded with meteoroids which form craters from 10 cm to 100 m in diameter at random locations. The crater diameters are acquired by the using the probability distribution of Equation 1, where
Contributions of Impact Ejecta in the Preservation of Ice Deposits on the Moon: P.D. Cox

Rates in Lunar Cold Traps

\[ p(x) = 1 \times (x + 0.001)^{-3/4} \]  
\[ t(r) = C r^{-3} D^{2.74} \]

\( x \) is a random uniform deviate anywhere from 0 to 1 produced by a random number generator. This equation is calculated by inverting the function found by approximating the flux curve of [14]. It is assumed that craters smaller than 10 cm contribute little to ice coverage, therefore, the ice they vaporize can be approximated by a rate that is added to the removal rate of dissociation and sublimation. This input parameter is around 200 microns per million years according to the current models [4,5]. Craters larger than 2 meters in diameter cannot be accurately modeled with the volume of this model. Therefore, the model has been run for craters up to 2 m diameter to obtain the basic distribution. The small number of large craters formed in the period of time considered allows these effects to be calculated separately and superimposed on the distribution.

Once a crater size is acquired, a crater is excavated at a random point in the array. The volume of ejecta is calculated and is distributed according to Equation 2, where \( D \) is the crater diameter; \( C \) is a constant; \( t \) is the ejecta thickness as a function of radial distance from the crater rim [14]. The constant is proportional ratio of the volume of regolith within a crater and the craters total volume. This ejecta produces the regolith that will protect the covered ice by the removal rate. If the thickness is great enough, the ice will also be protected from subsequent impacts. The crater itself is assumed to be a spherical bowl. The impact position is considered the center of a sphere, where any material within the radius is removed, whether ice or regolith. It is not unreasonable to assume that, on the average, as much regolith is transported out of the modeled area as is transported into it. Therefore, edge effects are treated by mirroring \( x \) and \( y \) coordinates.

After the creation of a calculated number of craters, the removal rate is applied to any exposed ice and the total number of ice cells are counted. Time is tracked by
counting the number of craters between 10 and 20 cm. Because of the logarithmic form of the crater distribution curve, the number of craters formed in this interval is an approximate measure of time in a continuous distribution of meteoroids. According to the flux given in [15], three craters between these limits will be created each million years in an area of the dimensions modeled. For a removal rate of 200 microns/my, one exposed cell from each column is removed after every 50 of such craters. For a total model time of 250 my, 750 of these craters would be allowed to form.

Conclusions

Results of models with different assumptions will be presented. The minimum amount of preserved ice is determined by the effect of the largest craters. Preliminary indications are that for a 10 cm ice layer, after 250 million years, between 1-10% will be covered by sufficient regolith to preserve the ice indefinitely.

REFERENCES AND NOTES

[9] Although the vapor pressure of volatiles such as crystalline and amorphous water ice have not been verified experimentally to temperatures at or below 100 K, the extrapolations show that they are small enough to cause the sublimation to be in the same range as the water production.
[10] Diffusion has not been included, but it is assumed that at the temperatures present in these cold trap areas the migration of water molecules to the surface is negligible due to the rates associated with small thermal gradient, pore size, and vapor pressure of ice [Clifford S.M. (1986) Soil Science 141, 289-297].
[11] The filling of small craters as proposed by [Langevin et al. (1982) JGR 87, 6681-6691] has not been included in the model to keep the level of complexity to a minimum.
CHARACTERIZATION OF IRON-NICKEL SULFIDES IN THE CHONDrites

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INTRODUCTION

Sulfur is an element found in all galactic environments, including the most primitive extraterrestrial material which is comprised of cometary and asteroidal particles collected in the Earth's stratosphere; this material is known as chondritic interplanetary dust particles (IDPs). The most common sulfur-bearing minerals are Fe and Fe-Ni sulfides such as troilite (FeS), pyrrhotite \((\text{Fe}_{1-x}\text{S})\) and pentlandite \((\text{Fe,Ni})_9\text{S}_8\). Although there is a vast number of sulfide minerals, each has its own well determined stability conditions. The objective of this research is to characterize sulfides in chondritic meteorites in order to establish the conditions (temperature and pressure) under which they were created. Subsequently, it will be possible to infer whether the sulfides were formed in the solar nebula or on the asteroids, and if formed on the asteroids, deduce how much alteration has occurred there.

Troilite is thought to be the first sulfur-bearing mineral to originate in the early solar system. The model for the genesis of this primary mineral indicates that troilite results from sulfidation of a metal (Fe-Ni) grain in an \(\text{H}_2\text{S}\)-containing environment. Friable troilite grains, which are exfoliated from the metal nucleus, are submitted to continued sulfidation, and pyrrhotite is thereby produced since the troilite is no longer in direct contact with the metal [1]. Thus, pyrrhotite is classified as a secondary origin sulfide.

Some asteroids are known to have contained ice. When this ice melted, various reactions must have occurred on the hydrous body, forming products such as phyllosilicates and new generations of sulfides (pyrrhotite and pentlandite). Pentlandite is known to form during alteration; the observation of this phase in hydrous IDPs [1] concurs with formation during aqueous alteration.

EXPERIMENTAL PROCEDURE

Sulfides from three chondritic meteorites were analyzed during the course of this study. They include Vigarano, a reduced CV3 chondrite, and Mighei and Nogoya, both CM2 chondrites. A significant number of sulfide grains (~20 from each site) from chondrule rims and the meteorite matrixes were analyzed using a Transmission Electron Microscope (TEM) coupled with an Energy Dispersive X-ray Analyzer (EDX). These instruments were used since they are capable of resolving and chemically characterizing very fine scale features (from 10 nm) such as the sulfide grains.

RESULTS

The sulfides analyzed were randomly selected from different areas of microtomed sections. Their morphologies ranged from rounded, oval, spherical, rectangular to anhedral, and their sizes extended from 0.1 to approximately 1 micron. Results from the morphology comparisons of the rim and matrix sulfide grains will be reported at a later date.

The Fe, S and Ni atom percentages from the various analyses were plotted in triangular diagrams (Figure 1). The data reveals that sulfides from the chondrule rims and the meteorite matrix have similar compositions for each chondrite.

Observation of the triangular plots of the Mighei and Nogoya analyses reveals the dominating
presence of an unknown sulfide with an Fe-Ni content similar to pentlandite, but with an increased S content. It is possible that a solid solution not observed in terrestrial sulfides exists within this area of the diagram. A few troilite and pyrrhotite grains were also observed in the Mighei meteorite. Sulfide analyses from the Vigarano matrix have yet to be recorded. However, most of the sulfides from the Vigarano rim, displayed a composition between troilite-pyrrhotite and pentlandite.

Because pentlandite is a known alteration product, the presence of Fe-Ni sulfides indicates an increased degree of alteration in the Nogoya meteorite, and lesser alteration in Mighei. These results concur with the CM alteration model established by Browning et al. [2]. The predominant presence of Ni-poor sulfides in Vigarano, a reduced CV3 chondrite, seems to indicate that this meteorite is the least altered. These findings agree with studies effected by Zolensky and McSween [3], which state that the effects of aqueous alteration on CV chondrites are not as pervasive as in the CM meteorites.

CONCLUSIONS AND FUTURE WORK

Had the chondrule sulfides been distinct from those of the matrix, it would have been plausible to infer a different origin for each. If such were the case, it is possible that the rim sulfides would be specimens of primary materials formed by sulfidation in the early solar nebula, and later accreted onto the parent body. Because the chondrule rim sulfides had nickel-rich compositions resembling the matrix sulfides, it is hypothesized that these materials were formed in similar conditions, probably on the asteroid. Due to the occurrence of alteration processes, inferred from the Ni-rich composition of the sulfides, we suggest that any primary sulfides that may have been present were modified during hydrous reactions.

If an altered hydrous asteroid is submitted to subsequent heating, the mineral assemblages indicative of its alteration history may be obliterated. However, some Fe and Fe-Ni sulfides could be resistant to such degradation. These materials would then be fossils recounting the past alteration of the parent body. Consequently, future studies of apparently unaltered chondrites should take note of such sulfides as possible indicators of a more complex geochemical history.

The origin of sulfides discussed in this paper is simply the beginning of a greater goal. As previously stated, since the stability fields of the different sulfides have been determined, identifying which sulfides are present in a certain sample will provide information on the physical-chemical constraints of the early Solar System, thereby giving us a better understanding of processes such as dust condensation, asteroid formation and subsequent evolution.

ACKNOWLEDGMENTS

I would like to thank Dr. Michael E. Zolensky for his patience and support. Without him I would never have been introduced to the fascinating world of sulfides. I would also like to thank Kathie L. Thomas-Keptera for all her help with the TEM.

REFERENCES

Figure 1-Triangular Diagrams of Sulfide Analyses (Atomic %)

- **Vigarano Rim**
- **Mighei Matrix**
- **Mighei Rim**
- **Nogoya Matrix**
- **Nogoya Rim**
- **Pyrrhotite**
- **Troilite**
- **Pentlandite**
Stereotopography of Debris Aprons and Related Flow Features: Mars

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Debris aprons are one of several features that may indicate the presence of ground ice on Mars today. The ultimate goal of this project is to determine the presence of ice in the Martian regolith and its role in these flows. The geomorphology of these features have been observed, but little has been done on the quantification of their slopes and thickness. These lobate features, convex in profile radiate out from the base of large massifs. The surface morphology suggests that they are flowing today or have flowed at one time in Martian history. Talus cones, lineated valley fill, and landslides also exhibit similar flow characteristics, and were also studied as possible flow analogs. Measuring the slopes and thickness of debris aprons and applying them to simple flow laws gives estimates of the time required for these features to form, which may relate to variations in temperature and ice quantities in the Martian past.

Introduction
Debris aprons are features on Mars that appear to have deformed viscously, like a flow. Examination of these debris aprons may reveal the presence of ice in the regolith. The northern hemisphere area of study is around the cratered highlands in Protonilus Mensae, Nilosyrtis Mensae, and Deuteronilus Mensae (40 N, 300 W). Debris aprons in this area were first identified by Squyres [1]. The southern hemisphere study area, east of the Hellas impact basin in the heavily cratered southern hemisphere highlands, has most recently been studied by Crown [2]. Debris aprons are concentrated in the mid latitudes, between 30 to 60 degrees north and south latitudes. Ice, at these latitudes could be stable up to 10 m below the surface of Mars [Clifford, [3], where the midlatitude mean annual temperatures are around 200K. Because of similar flow morphologies, terrestrial features such as debris flows, landslides, talus lobes, and rock glaciers, are potential analogs to Martian debris aprons. Rock glaciers, a common analog, are thought to flow by means of interstitial ice in the rock debris. However, little is know about the flow mechanism of terrestrial rock glaciers[4-7].

Features
Debris aprons form at the bases of large mesas or massifs(figure 1). They have a lobate and convex profile and seem to flow away from the escarpment as a viscous mass[1]. These debris aprons exhibit lineations in moderate and high resolution Viking images. Lineations tend to be either perpendicular or parallel to the scarp and the lobate snout at the terminus of the flow. These features point to extension and compression of a viscous mass that has undergone flow deformation. Other features such as lineated valley fill indicate some sort of flow. These features form between mesas and are characterized by longitudinal ridge and furrows parallel to the valley in which they lie. These have been compared to Antarctic ice streams [8]. They may also be caused by compression of material ablating from the surrounding mesas. There is little indication that suggests that debris aprons formed catastrophically, like a landslide or large rock fall, however this should not be completely ruled out. Squyres [9] and Lucchitta [8] speculate that these features flowed away from the base of the massif as a slow creep, with the erosion from the mesa supplying the material in the debris apron. Interstitial ice in the debris may act as a lubricant for the features to creep. This interstitial ice may deform by dislocation of polycrystalline ice, at the atomic level, over long period of time, causing the flow features seen today.

Data
Analysis of debris aprons was done using Viking stereo images of selected sites to produce topographic
maps of these features for the first time. Once a suitable stereo pair has been identified, the images are calibrated and registered. An automated stereogrammetry program, developed at the Lunar and Planetary Institute, maps the brightness patterns of the images and determines the amount of parallax associated with each feature, from which elevation is calculated.

Scarp heights, debris thickness, and slopes were measured for the debris aprons from the produced topographic maps. The heights of the scarps ranged from 1 to 3 km, with their slopes varying from 7 to 50 degrees. Flow thickness range from 0.4 km to 1.5 km. Slopes of the debris aprons range from 1.7 degrees in Protonilus Mensae to steeper slopes of 6.5 degrees near Hedriaca Patera, in the area east of Hellas. Terrestrial data corresponds quite well with this data. Surface slopes of rock glaciers in the Brooks Range Alaska, noted by Calkin et al, in Giardino [10] vary from 2 to 5 degrees on the upper surface, which is similar to the low slopes found on Martian debris aprons. However, the magnitude of the Martian debris aprons is much larger in length, width, and thickness when compared to terrestrial rock glaciers.

**Results**

The data from the debris aprons is compared to height and length ratios of landslides on Earth, Venus, and Mars (figure 2[11]). The larger of the Martian debris aprons measured to date, have ratios between terrestrial subaerial landslides and terrestrial submarine landslides, which are rather high. Others have ratios more similar to terrestrial landslides.

Simple flow laws were used to evaluate strain rates and flow time scales of Martian debris aprons. The basal shear stress, $\tau$, is found using the equation

$$\tau = \rho gh \sin \alpha$$

where $\rho$ is the average density of the flow material ranging from 950 Kg/m3 for pure ice to 2800 Kg/m3 for the density of rock on Mars, $g$ is the gravity at the surface of Mars taken as 3.72 m/s2, $h$ is the thickness of the flow deposit at the thickest point, $\alpha$ is the surface slope. Using the shear stress of this body, the strain rate can be calculated using Glen’s law

$$\dot{\varepsilon} = A \tau n$$

where $n$ is a constant with a value of around 3 and $A$ is a flow law constant that depends on temperature, crystal size, crystal orientation, and impurity content. $A$ is described by

$$A = A_0 e^{-Q/RT}$$

where $A_0$ is $4.3 \times 10^{-4}$ s-1 kPa-3, $Q$ is activation energy for creep taken to be 60 kJ mol-1 for $T < 263$ K [Clifford 1987], $R$ is the gas constant.

A debris apron in Protonilus Mensae was as a test case. This debris apron has an average surface slope of 2 degrees, a thickness of 0.584 km, an assumed temperature of 200K, density of 950 Kg/m3, that of 100 % ice, which allows for a minimum shear stress and strain rate. Using these flow laws, we estimate a basal shear stress of 72.0 kPa and a strain rate of 6.6 X 10^-15 s^-1. From the observed strain, these strain rates indicate emplacement values of 10 to the 7th yrs. However, this is a minimum time for the flow to occur, an ice/rock mixture would take much longer. The albedo of pure ice would be apparent in images and therefore these flows are assumed to have some rock content entrained in the flow. If the rock content is 50%, the average density will be roughly 1.8 g/cm2[12]. This ice/rock mixture can raise the strain rate by an one to two orders of magnitude[13]. This raises estimates of the time taken for these materials to flow, up to 1 billion years. This is not an unreasonable amount or time in an environment that remains somewhat stable, such as Mars.

**Interpretation**

The morphology of these features suggest that they may behave similar to terrestrial rock glaciers, but extreme amounts of time must be available for flow to occur at these low slopes and temperatures, and this may indicate that these debris aprons are still flowing. These debris aprons do not have very many overlying craters and appear to geologically young (<1 billion years).

Variations in Martian climatic conditions may explain these flows. Mars has a highly variable obliquity cycle and during periods of high obliquity the mid-latitudes would be exposed to higher temperatures, which would enhance flow rates and speed formation times. These features may also have solid ice cores, along with the interstitial ice. This solid ice core would allow the debris aprons to flow more similar to ice, and they may be behaving similar to glaciers on Earth today. Salts can act to lower the freezing point of ice, if the regolith contained a high content of salts, along with the high basal pressures of these masses, the debris would of these bodies may be more likely to flow faster. Another possibility is impacts or Mars quakes could cause mechanical fluidization or loosening of the debris making up these features and allow them to flow. Because of the low slopes and low temperatures there is evidence against these features being able to flow as dry rock material, however
when ice is added to the model the likelihood that these featured flowed increases. In summary, the debris aprons analyzed in this study exhibit similar slopes and height to length ratios as the terrestrial analog, rock glaciers. However, due to the current climatic conditions on Mars, these features can not flow as easily as Earth rock glaciers. These debris aprons would take over 1 billion years to flow their current distances at current temperatures, if a shorter time constraints are imparted, then alternative explanation for the mechanism for these flows or a variation in Martian climatic stability must be assumed.

Acknowledgments: Thanks to my advisor Paul Schenk and to Brian Fessler, for his extensive help and patients.

References

Figure 1 Debris apron in region east of Hellas, -46S, 254W. Mosaic of Viking orbiter images. 584b13, 14, 15 and 17

Figure 2 Solid dots indicate scarp height to runout length ratios for debris aprons on Mars. [11]
Introduction
Gilbert [1] and Shoemaker [2] showed that the most probable angle of impact is 45°. Twenty-five percent of all impacts occur at angles less than 30° and three percent occur at angles less than 10°. Gault and Wedekind [3] performed laboratory experiments in which craters were created in various materials utilizing a variety of impact angles. From these experiments, it was determined that the diameter of the resulting crater decreased with decreasing impact angle. Results have also shown that the crater rim of the oblique impact remains circular down to an angle of 10°, but the ejecta distribution becomes asymmetrical upon impacts of less than 45°. At low angles of impact (less than 5°), a butterfly ejecta pattern is formed, producing 'forbidden' zones (where no ejecta is present) both uprange and downrange. It was also shown that ricochet occurs for some very low-angle impacts [3].

Through various studies, such as Wilhelms [4], it has been shown that the morphologic effects of varying impact angles observed by Gault and Wedekind [3] are mimicked by lunar craters. However, the three-dimensional shapes of fresh planetary impact craters have never been compared to those experimental results. The purpose of this study is to use high-resolution topography, generated from stereo photogrammetry, to study the shapes of lunar impact craters and draw comparisons with the experimental data.

Methods
Stereo photogrammetry has been used in this study due to its ability to provide high-resolution topographical information. The laser altimetry data collected does not produce data to such a high degree of spatial resolution. Stereo photogrammetry is produced by having two pictures taken of an identical feature from two different angles. Knowing the angle of parallax created between these two pictures, topographical data can be determined. Unfortunately, stereo imagery is not available globally for the moon.

Information used for this study was extracted from topographic data collected by the Apollo missions. The Apollo 15 - 17 missions collected stereo images covering approximately 10% of the lunar surface. More recently, Clementine collected stereo data using an ultraviolet-visible (UVVIS) camera [5], covering a region from 45 N to 45 S and 15 - 26 W and 35 - 110 W. The Apollo data was used because the angle of parallax between the pictures taken was smaller than the angle of parallax used in the Clementine data. This enables a better three-dimensional image to be formed from the Apollo data than can be formed from the Clementine. Also, the data from the Apollo missions has already been processed to produce Lunar Topographic Orthophotomaps (LTOs), contour maps at approximately 100m horizontal resolution and 100m contour intervals.

Sixteen fresh craters were chosen to be more fully examined. There were numerous criteria involved in the selection of craters for study, such as they needed to have LTOs available in order to have digital elevation models (DEMs) created. To ensure that the effects of degradation due to age and size were kept to a minimum, the craters were relatively young and small, with diameters ranging from 10 - 35 km. Care was taken to choose asymmetric craters that exhibit a variety of ejecta distribution patterns and crater rim circularity. Symmetric craters of corresponding sizes were also chosen to allow comparison of symmetric and asymmetric profiles.

Once the craters were chosen, DEMs were created from the LTOs. Profiles were then taken of these digital models and compared. Profiles of the asymmetric craters were taken in the downrange and perpendicular to the downrange directions.

The crater Angstrom was studied both with Clementine and Apollo data. Results from the two methods are comparable, demonstrating that the Clementine data can be used to produce accurate topographical information.

Observations
Profiles from the DEMs were analyzed for trends with varying impact angle and crater diameter. The most noticeable feature of the asymmetric craters was the raised rim in the downrange direction. This is due to an increased amount of ejecta being thrust in the direction of the projectile's path. This was also observed by Gault and Wedekind [3]. This is exhibited very clearly by the profiles taken of the crater Menelaus (Figure 1). The difference in the downrange and uprange heights appears to be greater as the diameter of the crater increases. For
example, the largest asymmetrical crater studied was Necho with a diameter of 33.75 km. There was a 1 km difference between the height of the downrange rim and the uprange rim, yielding a rim height difference - diameter ratio of .029. The smallest asymmetrical crater studied, Messier A (diameter of 13.25 km) exhibited a 100 m difference in rim height, resulting in a ratio of only .007.

For smaller craters, i.e. diameters equaling 10 - 27 km, the depth to diameter ratio of the symmetric craters, with respect to the asymmetric craters, is greater. The average depth to diameter ratio of the symmetric craters at these diameters is .1905. The average ratio of the corresponding asymmetric craters is .160. For craters with diameters greater than approximately 27 km, the opposite is true, i.e. the relative depth to diameter ratio of the asymmetric craters is greater. The average ratio of the asymmetric craters is .112, while for the symmetric craters the ratio is .0925. The floors of the asymmetric craters appear to flatten as the diameter of the crater decreases. Conversely, the floor of symmetric craters appears to flatten as the diameter increases.

By comparing the experimental data, in which the impact angle is known, and the lunar data it is possible to deduce the unknown lunar impact angles. Two craters for which this is possible are Messier and Messier A. The butterfly ejecta pattern of Messier and its saddle-like rim lead to the conclusion that Messier was formed by an extremely low angle impact, possibly as low as 1°. This would be a ricochet feature. Messier A has distinct whisker-like ejecta rays in the downrange direction, which are also produced by very low angle impacts. It is speculated that Messier A was formed by an impact of 5°. (Figure 2)

The craters chosen to be studied were chosen because they exhibited various ejecta distribution and shape in different diameters. Equal numbers of asymmetric and symmetric craters were studied. While care was taken to attempt to choose a representative group of craters, it is possible that the craters chosen, especially the asymmetric, do not represent the norm, but are anomalies. Even in the symmetric craters of comparative diameters, differences occur with respect to relative depth and rim height. In order to get more definite results, a larger sampling of craters needs to be studied.

Conclusions

In comparing the data collected by Gault and Wedekind [3] and that created through the DEMs, it can be concluded that the experimental craters are topographically mimicked by lunar craters. This can be determined by the profiles that display raised rims, and asymmetrical ejecta coverage. Furthermore, using these profiles, the impact angles of select asymmetric craters can be deduced. On the following page, some of the craters studied are compared with their suspected laboratory counterparts. Messier and Messier A, both exhibit the characteristic profiles of low angle impacts. A comparable diameter crater, Peek, is an example of a symmetric crater.

This study has begun to compare the topographic relationships between the laboratory craters and the actual lunar craters. It is possible, knowing the basic shape of the crater and the ejecta blanket distribution, to determine the angle at which the impactor hit the lunar surface. More work needs to be done to further recognize other trends and lead to a more complete understanding of cratering mechanics.

Acknowledgments

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References
Figure 1

Menelaus

Figure 2

Geometry of craters formed in powdery dust (<100 μm) by oblique impacts of pyroxene spheres at velocity approximately 6 km/s.

Messier and Messier A

Peek
Scanning Electron Microscopy of Yellowstone Thermal Spring Travertines
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NASA/JSC

Introduction
The search for evidence to substantiate or deny the possibility that life once had, and may still, exist on Mars, and the quest for understanding the origin of life on Earth are two human undertakings which will be necessarily intertwined in the future of scientific investigation. This project links these two areas of study in that the extremophiles indigenous to Yellowstone thermal springs -- an environment thought to be a good terrestrial analog to the environment on and below the surface of early Mars -- appear to be the most primitive extant organisms according to the 16S rRNA-based phylogenetic tree [1].

The purpose of this project was to investigate, using scanning electron microscopy, the organisms and biogenic material present in travertine samples originating from several hot springs in the Yellowstone National Park. These images will be used to: 1) expand our knowledge on the morphology of these extremophiles, 2) analyze the purported biogenic forms found in the carbonates of the martian meteorite ALH84001 [2], and 3) gain further insight on subsurface microbial ecology for future studies.

Materials and Methods
Six travertine samples from three different thermal springs were collected by Dr. Carl Allen (Lockheed-Martin/NASA,JSC). The travertines consist of two forms of calcium carbonate crytals, calcite and aragonite. Two samples were taken from different locations in each of the Painted Pool, Narrow Gauge, and Soda Butte springs. The water temperatures at the collection sites were 35.0° C, 66.3° C and 72.8° C, and 13.6° C respectively, with pH in the 6.2 to 6.9 range. These samples were fractured to expose a fresh surface, etched with 0.5% HCl for thirty seconds and rinsed with double-distilled water. The travertines were mounted on carbon planchettes or discs using aluminum foil and carbon paint. All samples were coated with a gold-palladium alloy. A Philips 40XL Field Emission Gun scanning electron microscope was used for imaging. Chemical composition of both the biological features and the underlying substrate were analyzed using an Oxford Isis energy-dispersive X-ray spectrometry (EDS) system.

Results
Images were taken of several different types of biological forms in the Yellowstone samples categorized in the following way:
1) “normal-sized” bacteria (1µm or larger in size) 2) nanobacteria (200-1000 nm) 3) biofilm 4) filaments 5)colonies or chains of spherical or rod-shaped cells.

The Soda Butte sample retrieved from a biologically active area of the spring (SB2 Active) was found to contain an abundance of large (≥1µm) cells which had collapsed, probably due to exposure to the vacuum of the SEM. Several cells had ruptured, giving them a ‘deflated’ appearance. Chemical analysis showed that the cells had a high carbon content and elevated Ca and S content relative to the calcium carbonate substrate, indicating that these bacteria had not yet become mineralized (fig.1).

The Soda Butte Dormant sample contained fossilized forms revealed by etching. The forms were predominantly in the shape of segmented chains, each segment on the order of 1 µm in length, with total chain length in some cases reaching up to 15 µm. A small filament resembling these chains was found in one case, consisting of segments approximately 40 nm in length (fig.2). This filament was too small to get a sufficient chemical analysis with EDS.

A 4 µm-long ovoid body, connected to another ovoid half its size via a filament, was also imaged. This structure had a grainy surface texture, with a fluffy, colonial appearance (fig.3). Chemical analyses of all the forms in SB2 Dormant revealed Ca, C, and O, indicating that mineralization to CaCO₃ had occurred.
Evidence of biofilm was abundant in both of these samples. A mineralized example of biofilm in the dormant sample was found to be morphologically very similar to structures found in ALH84001 (fig.4). Chemical analysis of SB2 Active biofilm examples showed that they contain an elevated level of Ca and S relative to the surrounding travertine.

The Painted Pool samples were found to contain examples of normal-sized bacteria, filaments, and biofilm remnants, often in close proximity (fig.5).

The Narrow Gauge samples were found to contain the largest abundance of microbial biofilm remnants of all the samples, having many aragonite rosettes well-covered with this organic material (fig.6). However, I was unable to find forms in these two Narrow Gauge samples that can be determined to be bacteria with a high degree of certainty.

**Discussion**

This study has provided a closer look at the life in Yellowstone hot spring environments. Use of a SEM has expanded the field of view, previously gained from optical microscopy studies [3], of the biology associated with these travertines. Not only have normal-size bacteria been documented, but smaller biogenic structures such as filaments and biofilm have been recognized in these travertines. Biofilm remnants from the Yellowstone samples have been found to be similar in size range and morphology to structures imaged in ALH84001. Future research will include attempts to isolate, culture and genetically sequence bacteria from these samples. Further SEM work and chemical analyses are also planned.

Hydrothermally active sites are often cited as plausible environments for the origin of life. High temperatures promote chemical reactions, and liquid water is essential for life as we know it. Abundant heat energy due to volcanism and protection from intense UV radiation which bombarded the early Earth, suggest the possibility that life may have begun in the interior and only later invaded the surface. The common ancestor of the three main lineages, Archaea, Bacteria, and Eucarya, was very possibly a hyperthermophile according to 16S rRNA phylogenetic classification [4]. Analysis of their 16S rRNA shows that they are the most primitive organisms still existing, with chemolithoautotrophs comprising the shortest lineages on the phylogenetic tree [5]. Further study of these organisms is essential as preparation for the search for extant or extinct extraterrestrial microbes on Mars.

Large amounts of water, perhaps even oceans, may have been present on early Mars, and "high heat flows expected at that time would have resulted in high rates of volcanism". This implies that hydrothermal environments, with volcanically supplied reactive clays and minerals, were present on early Mars, making it an environment appropriate for the evolution of life [6]. In light of this, the ecology of thermally active environments is one of the best resources we have before the Martian sample return in 2008.

**References:**

Fig. 1 Normal-sized bacteria in Soda Butte Active

Fig. 2 Smal-sized filaments in Soda Butte Active

Fig. 3 Mineralized bacterial body in Soda Butte

Fig. 4 Mineralized biofilm similar to structures in martian meteorite ALH84001

Fig. 5 Bacterial filament from Painted Pool

Fig. 6 Biofilm from Narrow Gauge
JSC Mars-1: A Martian Regolith Simulant
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Advisor: Carl Allen, Ph.D., Lockheed Martin Engineering & Sciences, Houston, TX 77058

Introduction
JSC Mars-1 is a simulant of the regolith of Mars. The simulant was developed from the <1 mm size fraction of altered volcanic ash (Hawaiite) from the Pu‘u Nene cinder cone located on the Island of Hawaii. Tephra from this cone has been repeatedly cited as a close spectral analog to the bright regions of Mars [1,2,3]. The simulant will be used for support of scientific research, engineering studies, and education.

Simulant Preparation
The simulant was collected on the slopes of Pu‘u Nene from a palagonitized zone 40-60 cm thick beneath the soil overburden. The tephra was dried and sieved to collect the <1 mm size fraction. It was packaged in moisture-proof containers for shipping and storage.

Simulant Characterization
The simulant was described using a variety of analytical techniques including scanning electron microscopy (SEM), X-ray diffraction, the Franz magnetic separator, sieving, reflectance and Mössbauer spectroscopy, neutron activation analyses, X-ray fluorescence, and loss on ignition. The analyses were performed on a mixture of samples from three separate containers in order to reduce the error due to the variations over the collecting area.

Magnetic Properties
The simulant is composed of two distinct populations of particles, dark glassy unaltered grains and orange altered grains. These two phases can be roughly separated magnetically, with the dark unaltered grains being more magnetic. Approximately 22% of the smallest size fraction of the simulant can be lifted by a strong hand-held magnet while 40% is unresponsive to magnetics. The other 38% is attracted to a strong magnet, but is not lifted by it. The phase responsible for the magnetic attraction is titanomagnetite found as crystals within the grains (figure 1).

Spectra
JSC Mars-1 is yellow-brown in color. Its VIS/NIR spectrum is compared to a composite Martian bright area spectrum in figure 2. Both spectra contain similar ferric absorption features in the visible and relatively flat absorption in the near-IR. The simulant contains H₂O and OH absorption bands at 1400 and 1900 nm. The presence of ferric absorption features near 600, 750, and 860 nm in the Martian spectrum imply higher levels of red hematite on Mars than in the simulant [4].

The spectrum for the dark, unaltered material is almost completely flat with an overall reflectance lower than the bulk simulant (figure 3a). This spectrum still contains the ferric absorption edge through the visible, but there is only slight, if any, ferric absorption in the 800 to 900 nm range and almost no water absorption bands. The orange, altered material spectrum is the other extreme (figure 3b). It contains a pronounced ferric feature in the 800 to 900 nm region and deep water absorption bands.

Chemical Composition
The major and minor oxide composition of JSC Mars-1 is listed in Table 1. The data are compared to Viking and Pathfinder analyses.

The chemical compositions of altered and unaltered materials were also determined using the scanning electron microscope. The altered material contains less Si, Ca, and Mg.

Figure 1: Backscatter SEM image of a polished section of a magnetic grain. The white minerals are titanomagnetite, the medium gray mineral in the upper left is olivine, and the dark gray elongate minerals are Ca-plagioclase.
and more Al, Fe, and Ti than the unaltered glass. This microscale trend mirrors the macroscopic trend of the alteration of Hawaite to palagonite. Figure 4 shows a micrograph of a typical altered grain. Many of the seemingly unaltered surfaces of the grains were actually covered with a thin layer of altered material. Unaltered portions of grains, exposed by polishing, were analyzed by SEM.

Twenty-two weight percent of the simulant was driven off when the sample was heated to 900°C for two hours [4]. The volatile component consists mainly of H₂O but also contains SO₄ and Cl.

Mineralog y
The majority of the simulant is amorphous, either volcanic glass or altered gel. Minerals detected by X-ray diffraction and the SEM include anorthite, olivine, titanomagnitite, augite, magnetite, and quartz (figure 1). Iron Mössbauer spectroscopy detected magnetite, hematite, olivine, augite, and/or glass [3]. The majority (64%) of the iron in the simulant is present as nano-phase ferric oxide (Fe₂O₃), yielding a Fe²⁺/Fe³⁺ ration of 1/3 [4].

Grain Size
The particle size distribution for the simulant is listed in Table 2. In comparison, the blocky material which covers 78% of the area near the Viking I lander sight (VL-1) on Mars ranges in size from 0.1 to 1500 μm [4]. Edgett and Christensen [7] list the average grain size of Martian material to be around 500 μm while Martian dust has a grain radius near 10 μm [8,9].

Densities
The bulk density of JSC Mars-1 is 0.87 +/- 0.02 g/cm³. This value can be increased to 1.07 +/- 0.02 g/cm³ by vibrating the sample. The particle density is 1.91 +/- 0.02 g/cm³. The unaltered dark particles are more dense (2.15 +/- 0.09 g/cm³) than the altered lighter material (1.76 +/- 0.03 g/cm³). The density of the larger particles is also greater than that of
grained drift material near the restricted distribution from 0.48 to 0.94 g/cm³. For JSC Mars-1, the column density of 1.2 +/- 0.2, and the coarse, blocky material has a value of 1.6 +/- 0.4 [10].

Table 1: Chemical compositions of JSC Mars-1 compared with Martian surface fines analyzed by Viking and Pathfinder

<table>
<thead>
<tr>
<th>Oxide</th>
<th>JSC Mars-1(^{\dagger}) Wt. %</th>
<th>Viking(^{\dagger}) Wt. %</th>
<th>Pathfinder(^{\dagger}) Wt. %</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO(_2)</td>
<td>43.7</td>
<td>43</td>
<td>44.4</td>
</tr>
<tr>
<td>Al(_2)O(_3)</td>
<td>23.4</td>
<td>7.5</td>
<td>9.5</td>
</tr>
<tr>
<td>TiO(_2)</td>
<td>3.8</td>
<td>0.65</td>
<td>0.9</td>
</tr>
<tr>
<td>FeO</td>
<td>3.5</td>
<td>n.d.</td>
<td>n.d.</td>
</tr>
<tr>
<td>Fe(_2)O(_3)</td>
<td>11.8</td>
<td>17.6</td>
<td>19.1</td>
</tr>
<tr>
<td>MnO(^+)</td>
<td>0.3</td>
<td>n.a.</td>
<td>0.6</td>
</tr>
<tr>
<td>CaO</td>
<td>6.2</td>
<td>6</td>
<td>5.5</td>
</tr>
<tr>
<td>MgO</td>
<td>3.4</td>
<td>6</td>
<td>8.8</td>
</tr>
<tr>
<td>K(_2)O(^+)</td>
<td>0.6</td>
<td>0</td>
<td>0.7</td>
</tr>
<tr>
<td>Na(_2)O</td>
<td>2.4</td>
<td>n.a.</td>
<td>4.3</td>
</tr>
<tr>
<td>P(_2)O(_5)</td>
<td>0.9</td>
<td>n.a.</td>
<td>1.5</td>
</tr>
<tr>
<td>SO(_2)</td>
<td>n.a.</td>
<td>7</td>
<td>5.3</td>
</tr>
<tr>
<td>Cl</td>
<td>n.a.</td>
<td>0.7</td>
<td>0.4</td>
</tr>
<tr>
<td>LOI(^\ast)</td>
<td>21.8</td>
<td>n.a.</td>
<td>n.a.</td>
</tr>
</tbody>
</table>

\(^{\dagger}\) from [4]; XRF data normalized without LOI

\(^{\dagger}\) preliminary results from [6]

\(^+\) Values for K\(_2\)O and MnO are upper limits

\(^\ast\) Weight loss after heating for 2 hours in air at 900°C (includes H\(_2\)O, SO\(_2\), and Cl)

the smaller sized particles, which leads us to believe that the smaller particles are comprised of a larger proportion of altered material. The disturbed bulk density for fine-grained drift material near VL-1 is 1.1 +/- 0.15 g/cm³ and the coarse, blocky material’s value ranged from 0.48 to 0.94 g/cm³. The disturbed bulk density of JSC Mars-1 is 1.1 +/- 0.2, and the coarse, blocky material has a value of 1.6 +/- 0.4 [10].

Table 2: Particle Size distribution of JSC Mars-1

<table>
<thead>
<tr>
<th>Particle size</th>
<th>% of whole</th>
</tr>
</thead>
<tbody>
<tr>
<td>&lt; 149 µm</td>
<td>25 %</td>
</tr>
<tr>
<td>149 - 250 µm</td>
<td>24 %</td>
</tr>
<tr>
<td>250 - 450 µm</td>
<td>30 %</td>
</tr>
<tr>
<td>450 - 1000 µm</td>
<td>21 %</td>
</tr>
</tbody>
</table>

Discussion

JSC Mars-1 was determined to be the Martian regolith simulant because of its spectral similarities to the Martian bright regions [3]. Other similarities the simulant has with the Martian regolith are 'added bonuses.' The fine surface materials on Mars are suggested to be a mixture of weathered and non-weathered materials [11]. Candidates for the weathered material include palagonite, scapulite, and smectite clays, but none of the hypotheses can be proved over another [11]. JSC Mars-1 simulant is a mixture of unaltered volcanic ash and palagonite.

Acknowledgments

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References

LL chondrites and Prior's Rules  
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ABSTRACT  
A study of metamorphic sequence LL4 - LL7 shows subtle but systematic changes in the composition and relative abundances of olivine, low-Ca pyroxene, and metal with increasing petrologic type. These variations indicate Prior's Rules apply within the LL group chondrites. The metamorphic temperature for LL4 - LL7 is estimated 675-880°C (estimated precision is plus minus 50 °C.) Our data from Wo content in orthopyroxene shows the temperature increases from LL4 to LL6, then decreases from LL6 to LL7. These results indicate that the parent body experienced progressive metamorphism LL4 to LL6, and that some regions were metamorphosed for longer time at relatively lower temperatures, to produce the LL7s.

INTRODUCTION  
The three principal group of ordinary chondrites, H(high total Fe), L(low total Fe), and LL(low total Fe, low metallic Fe), constitute ~80% of all meteorites observed to fall. In going from H to LL chondrites, the abundances of siderophile elements decrease and the degree of oxidation increases. The proportion of oxidized Fe (i.e., FeO in silicates) increases at the expense of metallic Fe. This oxidation reaction can be represented as follows;

\[
\text{Fe} + \text{O} + (\text{Mg, Fe})\text{SiO}_3 \rightarrow (\text{Mg, Fe})_2\text{SiO}_4  
\]

metal  pyroxene  olivine

Thus with increasing oxidation the silicates become more Fe-rich and the olivine to pyroxene ratio increases. Because Fe is more readily oxidized than Ni or Co, the bulk metal becomes increasingly rich in Ni and Co. These relationships are known as "Prior's Rules"[1]. Ordinary chondrites exhibit various degrees of recrystallization, with accompanying changes in the degree of variability of olivine and pyroxene compositions, the structural state of low-Ca pyroxene, the degree of development of secondary feldspar, the present of igneous glass, the texture of the chondrite matrix and the bulk carbon contents. These changes, occurring as a function of degree of recrystallization, prompted Van Schmus and Wood [2] to divide ordinary chondrites into a sequence of petrologic types (types 3, 4, 5, 6, and later addition type 7). Type 3 is the least recrystallized (the olivine and pyroxene show a range in compositions and are "unequilibrated"; texture is highly chondritic.) and type 7 is the most recrystallized. Relatively little attention has been devoted to the relation between Prior's Rules and petrologic types, largely because the range of mineral composition within each group is very small. In this study, we examine the effects of metamorphism in petrologic groups within LL chondrites from the Antarctic collection.

The purpose of this study is to address the questions: (1) are there progressive variations in mineral abundances and compositions within the LL-group chondrites? (2) Do Prior's Rules apply within the LL-group? (3) Does the metamorphic sequence within the LL-group represent i) progressive annealing at successively higher temperature, ii) progressive oxidation, as was proposed by McSween and Labotka [3]. The LL-group was chosen for this study because LL-group has a broad compositional range, especially in olivine composition, making it easier to establish internal variations.

EXPERIMENTAL PROCEDURES  
Thin sections of 33 Antarctic LL chondrites were petrographically examined and classified. Petrologic types were assigned according to the criteria of Van Schmus and Wood [2]. For the most of these meteorites the silicate minerals were already analysed. Metal and sulfide grains were analysed for Fe, Ni, Co, S with a Cameca SX-100 electron microprobe. The proportions of metal, sulfide and silicate minerals were estimated from Back-scattered Electron Images, also using the Cameca SX-100 electron microprobe. In most cases, BSE images covered 9.9x5.6mm. The samples examined correspond to petrologic types 5, 6 and 7. LL4 samples are rare in the Antarctic collection and for comparative purposes the data of Heyse [4] were used.

RESULTS  
Variability of olivine and pyroxene  
Heterogeneous olivine and pyroxene crystals indicate a high degree of disequilibrium: they become more
Mutsumi Komatsu : LL chondrites and Prior's Rules

Fig.1 FeO contents of olivine and orthopyroxene vs. petrologic type.

homogeneous in composition with progressive degrees of metamorphism. The degree of homogeneity can be assessed by measuring the FeO content of multiple olivine grains within individual meteorites. Because olivine equilibrates more rapidly than pyroxene [5], olivine data are more readily linked to petrologic types. Our data show a progressive decrease in percentage standard deviation of FeO content of olivine with increasing petrologic type (LL5: 0.53, LL6: 0.43, LL7: 0.34%).

Composition of olivine and pyroxene vs. petrologic type

Fig.1 summarizes the mean olivine and low-Ca pyroxene compositions for each petrologic type. Differences in the mean compositions of olivine and pyroxene reflect progressive changes in the availability of oxidized Fe in the bulk meteorites. The data show a range of compositions within each petrologic type but average values of iron content for both phases increase with petrologic type consistent with an increase in oxidation state with increasing metamorphic intensity.

Abundances of metallic minerals vs. petrologic type

LL chondrites contain 0.15–3.0 volume% of metallic minerals, mainly taenite and kamacite. Metal grains are irregularly dispersed, 50-800nm dimension. Kamacite grains are very rare and were observed only in samples with a mean taenite composition Fe/Ni > 1. Most kamacite grains coexist with taenite grains; in some meteorites, they exist as a mixture of taenite and kamacite that has intermediate composition. The proportion of metallic minerals and sulfide minerals are plotted in Fig.2. There is a considerable scatter in the amount of metal (volume%) in each petrologic group. If Prior's Rules are valid within the LL group, the abundance of metal should decrease with progressive oxidation. No evidence of this correlation was observed, though this could readily be a consequence of the low precision of the abundance data for the small areas measured.

Prior [1] and Rubin [6] showed that kamacite becomes systematically richer in Co and poorer in Ni in going through H-L-LL sequence. The amounts of Ni and Co in metal as a function of petrologic type in LL chondrites are shown in Fig.3. Good correlation between petrologic type and kamacite Co (Co in kamacite), and taenite Co are observed. Kamacite Ni shows a negative correlation with petrologic type. The Co data within the LL group are in good agreement with Prior's Rules.

Thermometry

Heyse [4] and Scott et al. [7] showed Wollastonite content (Wo=CaSiO3) of orthopyroxene increases systematically with petrologic type and explained it as a temperature effect. Our Wo data in orthopyroxenes are shown in Fig.4. Wo content increases with petrologic type from type 4 to type 6, but decreases from type 6 to type 7. Lindsley [8] proposed a two-pyroxene thermometer which is based on Ca-Mg partitioning between coexisting pyroxenes. We estimate the metamorphic equilibration temperatures for type 4, 5, 6, 7: 675°C, 800°C, 880°C, 800°C (estimated precision is plus minus 50°C).
CONCLUSIONS

Subtle but systematic variations in the oxidation state of Fe and composition of metallic minerals with increasing metamorphism is observed. This means, there are systematic changes within the LL group, and that Prior's Rules can be applied within LL group chondrites. The metamorphic sequence is also a sequence of progressive oxidation (as was proposed by McSween and Labotka [3]), suggesting open system metamorphism on the LL parent body.

Our estimates of metamorphic temperature increase from type 4 to type 6, then decrease from type 6 to type 7. Although McSween and Labotka [3] proposed the metamorphic grade type 4 to type 6 is evidence of progressive metamorphism and annealing at higher temperature, we found this trend doesn't applied to type 7 meteorites. Our data on metamorphic temperatures indicates that the parent body experienced the progressive metamorphism type 4 to type 6, and that some areas metamorphosed for longer times at relatively lower temperature, to produce the type 7s.

REFERENCES

Global Dust to Planetesimals Transition in Viscous Solar Nebula

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1. Introduction

Astronomical observations of young low-mass stars have led to wide acceptance of the idea that these stars are surrounded by accretion disks. We believe that planets form by accumulation of solid matter in these disks, beginning with dust, into progressively more massive particles, and eventually culminating in the few large bodies. It is very important to realize that the global evolution of solid particles is not identical to the global evolution of gas in the disk. The evolution of the gas is governed by viscous stress, and thus is proportional to the differential rotation and the vigor of turbulence, the evolution of solids is governed by the friction between the gas and the solids, and thus is proportional to the relative velocity between solid particles and the gas, as well as the size of the solid particles. The gaseous disk spreads and it mass diminishes. Initially, the solid particles are very small and coupled to the gas. With the time they gain mass due to process of coagulation and decouple from the gas motion. Eventually, as the density of the gas decreases and the mass of solid particles increases, the radial distribution of solids converges to its final form. The final form is a major factor in determining the character of the nascent planetary system, inasmuch as any further processes does not much alter large-scale radial distribution of solid material. Therefore, if we want to understand planet formation, we need a model of accretion disk that not only is able to follow the evolution of the gas, but can keep track of the solid component as well. In this work we neglect all effects the presence of solids may have on the evolution of gas. The major effect of solid particles on the evolution of gas is through the opacity, which in protoplanetary disks is dominated by dust grains. We assume that the coagulation process is not perfect, so there is always enough amount of small grains to support the opacity. The second effect is the gas-solid friction, which is usually negligible compared to other forces acting on the gas, and its omission is justifiable. Therefore if we had a model of the evolving protoplanetary gaseous accretions disk, we are able first to calculate all quantities describing the gas, and then using them obtain the evolution of solid particles. Unfortunately comprehensive models of viscous solar nebula are time consuming and fragile numerical models. Therefore they are not very practical, if we are interested in obtaining many models with different initial conditions.

2. Methods

Recently the analytic model of viscous solar nebula was obtained (Stepinski 1997, submitted to Icarus). Using this model we can calculate space-time distribution of surface density of the gas, \( \Sigma(r,t) \), quickly and dependably for a broad range of initial conditions. To determine the evolution of solids we also need the relative velocities between solids and the gas, thus we need \( V_r(r,t) \) – the radial velocity of the gas (its tangential velocity is almost Keplerian). It can be determined from the gas continuity equation

\[
\frac{\partial \Sigma}{\partial t} = -\frac{1}{r} \frac{\partial}{\partial r} (r V_r \Sigma)
\]

The motion of solid particles in a gaseous environment is influenced by the drag force due to collisions with the gas molecules. The drag law has a different form depending on the size of a solid and on the local conditions of a gaseous environment. In general, the aerodynamics of solid particles in the solar nebula present a nonlinear problem to be solved on a computer. However, with some approximations we calculate the radial velocity of solids, \( V_{rd} \), from a following formula:

\[
V_{rd} = V_k \frac{2 \Delta \Lambda + V_k}{1 + \Lambda^2}
\]

Here \( \Lambda(r,t,a) = \Omega_k(r) t_s(r,t,a) \), where \( t_s \) is the characteristic friction time (so-called stopping time), \( \Omega_k = V_k/r \) is the Keplerian angular velocity, and \( 2 \Delta = (1/\rho) (\partial P/\partial \rho)(r/V_k^2) \) is the residual gravity divided by the central gravity. The particle size is denoted by \( a \). The presence of two terms in the factor \( [2 \Delta \Lambda + V_{rd}/V_k] \) reflects the existence of two mechanisms driving the motion of solid particles. The first term corresponds to driving particles by virtue of the gas tangential velocity. The second term corresponds to driving particles by means of the gas radial velocity. Thus \( V_{rd}(r,t,a) \) is a rather complicated function of all its drivers, but we can calculate it for any given gas evolution scenario and for an assumed size of solid particles. The knowledge of \( V_{rd} \) completely determines the behavior of particles providing they do not coagulate and cannot be thermally destroyed. Such an academic case cannot help us to understand the formation of the solar system, but it is very helpful in appreciating the difference between the evolutionary history of particles and the gas.

Fig.1. shows evolutionary trajectories of particles of different sizes suspended in the solar nebula that started its evolution from a disk of mass \( m_d = 0.245 M_\odot \), having angular momentum \( j = 5.6 \times 10^{52} \) (cgs units), and the initial size of about 11 AU. Disk evolution is driven by turbulence characterized by viscosity parameter \( \alpha = 10^{-2} \). In general particles can be divided into long lived, characterized by evolutionary timescales comparable to the evolutionary time scale of the gas, and short lived, characterized by timescales much shorter than the lifetime of the gaseous disk. There are two kinds of long lived particles. The first kind is composed of small particles, which are strongly coupled to the gas and approximately move together with it. On Fig.1. these particles are represented by \( a = 0.01 \) cm. The trajectories of such particles follow closely the streamlines of
the gas except very late into the disk evolution. The second kind is composed of large particles which are completely decoupled from the gas and move on Keplerian orbits. On Fig.1, such particles are represented by \( a = 1 \) km which is a typical size of a planetesimal. Their trajectories are straight lines as they have no significant radial velocities. The intermediate size particles are partially decoupled from the gas and, as a result, they have radial velocities much exceeding radial velocities of the gas, so they move very quickly toward the Sun and are therefore short lived particles. There are two examples of such particles on Fig.1, one with \( a = 1 \) cm which have a lifetime of about \( 1.0 \times 10^5 \) yrs, and second with \( a = 100 \) m which have a lifetime of about \( 6 \times 10^3 \) yrs. Even shorter lifetimes are possible for particles with sizes between 1-10 m.

Particle trajectories, however, depends not only on their sizes, but also on the gaseous environment. We can consider a very similar gas model with the only difference being the initial mass of the gas \( m_d = 0.15 M_0 \). Fig.2 shows evolutionary trajectories of particles of the same sizes as in the first model. They are quite similar for both kinds of long lived particles, but markedly different for short lived particles. Such a nebula is initially larger (even so it has less mass), cooler and less dense. The drag laws predict that in such an environment intermediate size particles have smaller radial velocities and live longer.

3. Conclusions

The work done so far is preliminary, without coagulation and thermal effects the obtained evolution of solid particles is not directly relevant to the shape of the solar system. Our next step is to include coagulation into our calculations. With coagulation we can start our computation from small particles suspended in a gaseous disk characterize by initial conditions. As the gas evolves small particles move with the gas, but as they coagulate they decouple from the gas and develop relatively large inward radial velocities. At this stage they are in danger of being lost to the sun, unless further coagulation increases their size and decreases their radial velocities. Once particles become planetesimal size they will remain at their locations. Taking account for all of these changes we can determine the surface density of planetesimal swarm and thus the character of the planetary system.

Some important conclusions can be drawn even from the present calculations. We have demonstrated
that particles of different sizes behave very differently. Moreover, particles of the same size can behave very differently in different nebula locations and/or evolutionary times. This suggests that the evolution of solids in the nebula depends very much on initial conditions. There are probably only certain initial conditions that can lead to the solar system. It is very easy for the 1-100cm particles to be lost to the Sun. This suggests that solar-like planetary system may not be common. However, particles can escape destruction by growing fast to large sizes and stop moving inward. In any case, the most general conclusion is that particles evolve differently from the gas, and since this is their distribution, rather than the gas distribution that determines the character of the terrestrial planets and the character of the cores of giant planets, the full study of their evolution must be undertaken in order to understand the formation of the solar system.
CHONDRULE PRECURSOR AGGREGATES IN UOC'S: MELTING HISTORY
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Chondrules are presumed to have formed in the solar nebula and understanding their formation is key to determining the processes at work in the nebula. A model for the crystallization of chondrules proposed by Lofgren [1] predicts a complete range of melting of crystalline aggregate materials from barely perceptible to total. If true, then one would expect to find aggregates in various stages of melting, i.e. becoming chondrules. We found such particles in thin sections of several unequilibrated ordinary chondrites (UOCs). These aggregates are chondrule precursors of the type previously described by [2,3]. They characteristically display irregular shapes, a wide range of grain size and/or compositional variation, and varied thermal histories. They are usually composed of fine grained debris and/or crystal fragments. Large isolated crystals or relic crystals are not uncommon, some particles even contain whole chondrules or fragments. Many of these aggregates are enclosed in a nebular derived rim.

Techniques:
We examined polished thin sections of five UOCs: Bishunpur (LL3.1), Semarkona (LL3.0), Krymka (LL3.0), Chainpur (LL3.4), and LEW86134 (L3.0). Nearly 100 particles were identified and studied under both optical microscope and electron microprobe. Backscatter electron images, chemical x-ray maps and mineral analyses were completed using the Cameca SX-100 electron microprobe at JSC.

We will report on a series of aggregates that show progressive degrees of partial melting. There are several clues one can use to determine the amount of heating/melting a particular particle has experienced. An unmelted aggregate will have angular metal and silicate crystals, and no visible mesostasis glass. The initial sign of melting is rounded Fe-metal, the product of melting eutectic mixtures of troilite and kamacite at approximately 990°C. With increased degree of melting, the metal begins to move outward, however, if there is rapid quench, it may not move far. If the metal does migrate, it often remains attached to the particle, forming an iron and/or sulfide-rich rim.

Evidence of silicate melting is highly dependent on the subsequent cooling rate. Slower cooling rates will result in dendritic mesostasis and crystals with euhedral overgrowths. Faster rates will produce rounded grains and a glassy mesostasis. In general, with increased melting, the more rounded the grains, and the more mesostasis is present. Also, the more melting a particle has experienced, the more round its overall shape will become.

Petrography:
The aggregate particles are described below in an attempt to demonstrate the wide range of melting textures observed. Less than 10% melting:

Krymka P-10 (fig. 1) is an irregularly shaped fine grained clastic aggregate. The grain size variation is in the range 1-100µm, with the majority of grains being in the 10-30µm range. It is predominately composed angular and broken grains of olivine, Fo72-85. There is no evidence of silicate melting and metal is largely angular. The particle is enclosed in a sulfide-rich nebular rim.

LEW86134,8 P-7 is a clastic olivine aggregate with one large (400µm) skeletal olivine. Aside from the skeletal olivine, grain size ranges from 1-250µm. The Fo content varies from 78-92. On average, the larger grains, particularly the skeletal olivine, are more Mg-rich than the finer grains.

Some Melting (10-30%):
Krymka P-6 (fig. 2) is a very fine grained clastic aggregate. Grain size ranges from 1-50µm. There is no evidence of silicate melting, but the particle was heated enough for the metal to become rounded and begin to move outward. It is enclosed in a nebular rim of metal and ultra fine-grained material.

Chainpur P-30 has a large (300µm) zoned olivine crystal surrounded by smaller olivines and pyroxenes which appear to have aggregated around it. The entire particle is somewhat depleted in Fe: Fo92-99, En94-95. Many of the grains have Fe-rich overgrowths, evidence of some silicate melting. In addition, it appears...
that much of the metal may have moved to the rim and what remains is well rounded. The nearly spherical particle is enclosed in a metal-rich nebular rim.

**Significant Melting (30-50%)**

Semarkona P-6 (fig. 3) is an irregularly shaped clastic aggregate. It is composed of very Fe-poor pyroxene (En98.99) and olivine (Fo99) in a Ca-rich mesostasis. The most notable feature is a 250µm relic olivine crystal with extensive Fe reduction. Besides the relic, grain size ranges from 10-100µm. This particle has clearly been heated. The mesostasis is dendritic and the silicates are significantly rounded. The Fe-metal is also rounded. The particle is enclosed in a discontinuous nebular rim. The olivines are very Mg-rich, Fo98.99. The entire particle has become nearly round and is enclosed in a discontinuous rim of ultra-fine grained material with very little metal or sulfide.

**Greater Than 50% Melting:**

Chainpur P-40 is an irregularly shaped, olivine-rich aggregate that has clearly been heated significantly. There are several obvious relic olivine crystals that show reduction. The rest are largely euhedral olivine crystals, though some do show rounding, set in a purplish glass mesostasis. Grain size, aside from the relics, ranges from 10-150µm. There are many large Fe blebs throughout, the silicates, on the other hand are significantly reduced in Fe (Fo98.99). The particle is enclosed in a rim of metal, sulfides, and ultra fine grained rim material.

**Chainpur P-27 (fig. 4)** is also an olivine-rich clastic aggregate which has been significantly melted and rapidly cooled. There are several large metal blebs throughout the particle. Grain size ranges from 10-150µm. Grains are largely rounded, though some still have crystal faces, set in a glassy mesostasis. The olivines are very Mg-rich, Fo98.99. The entire particle has become nearly round and is enclosed in a discontinuous rim of ultra-fine grained material with very little metal or sulfide.

**Discussion:**

The meteorites we studied were all of very low metamorphic grade (type 3.0-3.4), and the particles were randomly distributed within the thin section with melted particles juxtaposed to unmelted particles. Thus melting could not have happened in the meteorite, but must have occurred prior to meteorite aggregation. The presence of unaltered nebular rims on most particles suggests formation in the nebula.

It is difficult to draw a line where this type of particle ceases to be an aggregate and becomes a chondrule. One can imagine that with a little more melting many of these particles will begin to erase the evidence of their aggregate history and they would be considered chondrules. Even a small amount of melting, under the proper conditions, can significantly affect a particle, making its history difficult to discern. Lower degrees of melting as a means of creating chondrules with porphyritic textures is consistent with Lofgren's chondrule crystallization model, i.e. porphyritic textures evolve when a melt is crystallized with nuclei present [1]. Under slow recrystallization conditions, partially melted crystals and fragments can serve as nuclei. These relic cores may be difficult or impossible to detect if the composition of the overgrowth remains the same. Such low levels of melting are a good way of preserving volatiles.

**Conclusion:**

We have demonstrated that there is a continuum from unmelted to about 60% melting in these aggregates (greater degrees of melting are no longer recognizable as aggregates). This melting series gives new insight into the chondrule forming process.

**Acknowledgments:**

I'd like to thank Dr. Lofgren for giving me this opportunity, and for his support and guidance throughout this project. I would also like to thank Vincent Yang for his assistance with the microprobe and for his patience.

**References:**

Figure 1. Clastic aggregate with 0-5% melting.

Figure 2. Clastic aggregate with 10-15% melting.

Figure 3. Clastic aggregate with 35-40% melting.

Figure 4. Clastic aggregate with 60-65% melting.
THE EFFECT OF LITHOSPHERIC RHEOLOGY ON MANTLE CONVECTION ON MARS; Matthew E. Pritchard, University of Chicago, Chicago, IL 60637; and Walter S. Kiefer, Lunar and Planetary Institute, Houston, TX 77058

**Introduction:** The Tharsis province on Mars is a huge, roughly axisymmetric bulge about 5500 km across, 9 km high and the home to 12 volcanoes and extensive volcanic flows. The Elysium province also has several volcanoes and is about 2000 km across and 3-4 km high [1]. The formation of these provinces give important clues to the structure of the interior and to the thermal evolution of the planet. To truly interpret these clues, we need to have a model of how high topography is maintained over such broad areas. Two very different models have been proposed. One model is that the bulges are being actively supported by the uplift of mantle convection [2] while the other is that they are passively supported by lithospheric processes [3]. Our model combines these two processes because we want to trace the relative importance of the active and passive contributions over the history of the planet and obtain a better understanding of how these broad bulges of high topography are supported.

**Thermal History Model:** In order to model the relative importance of convective and lithospheric effects with time, we must determine the amplitude of convective uplift and the thickness of the elastic lithosphere. The key parameter to determining convective uplift and the thickness of the elastic lithosphere is the mantle temperature. The mantle temperature governs the viscosity of the mantle material and determines the vigor of the convective motions. At high temperatures, rocks behave viscously over long time periods, allowing the mantle to flow and convect heat from the interior to the surface and thus the mantle cools with time. Models have been developed that simulate mantle convection within the terrestrial planets by balancing the various heat sources (heat from radioactivity and the core) and sinks (heat lost to the surface from cooling of the mantle) to determine the amount of mantle cooling with time [4]. Thus, at each point in time, a temperature is computed that can be used to find other model parameters, like the viscosity and strength of convection. From this, the amount of surface topography due to convection alone can be determined [5,6]. The strength of the crust and mantle materials varies with temperature and pressure, so from the thermal model one can also calculate the thickness of the strong, elastic portion of the lithosphere as a function of time [3,7].
This project follows the general method of previous research [e.g. 4], but makes some modifications to the details of the convection model. We assume a concentration of the primary radioactive elements in accordance with values obtained from the SNC meteorites [8], although there is debate about the actual concentration of these elements on Mars. We begin our model 4 billion years before the present with the mantle convecting beneath a rigid surface --which is in agreement with geologic evidence for limited horizontal motion [9] -- and use convection parameters consistent with this scenario [5]. In analogy with [10], we assume an exponential increase in crustal thickness with time to 50-75 km at present, consistent with gravity measurements [2], and with a characteristic thickening time of 500 million years.

**Results:** So far, we have made a self-consistent model of mantle convection that includes a thickening crust as an upper boundary layer. Using this model we have calculated the evolution of the mantle and core temperatures with time (Figure 1). As previous workers have noted, our model shows a rapid decrease in temperature in the first one billion years and then a much slower decrease in temperature during later times. Our final calculated core temperature of 1790 K is hot enough to maintain a liquid metal core on Mars [11], consistent with the observation of a weak or absent Martian magnetic field [12]. Using the mantle temperature and other parameters, we have computed the predicted topography from convection as a function of time [5,6], neglecting lithospheric resistance to uplift (Figure 2). The final calculated value of the surface topography is 5.5 km for a Tharsis-scale feature, compared to a current measured value of about 9 km. Figure 3 is the thickness of the elastic lithosphere, which controls the lithosphere’s resistance to uplift. Elastic thickness increases in time to a final value of about 97 km, which is consistent with observations of surface tectonics [13].

**Future Plans:** Work is in progress to make a more realistic thermal model of the crust including concentration of radioactive elements into the crust and different values of thermal conductivity in the crust and mantle. There are further plans to couple the convective topographic uplift to the lithosphere. Elastic forces in the lithosphere will tend to resist uplift [14]. On the other hand, viscoelastic processes will partly offset this effect [15], allowing larger amounts of convective uplift.
References:

Figure 1. (left). Core Temperature (dashed line) and Mantle Temperature (solid line) as a function of time. Figure 2. (upper right) Surface topography of a Tharsis-scale feature caused by mantle convection as a function of time. Figure 3. (lower right) Elastic lithosphere thickness as a function of time, using a crustal thickness of 50 km.
Impact Craters As Probes of the Lunar Crust

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INTRODUCTION

The volumes of lunar lava flows provide important information about the thermal evolution and resurfacing history of the Moon and have been a subject of much debate. In the absence of direct surface measurements of flow thickness, indirect techniques must be employed to constrain the dimensions of mare units. Four methods have been developed for estimating the thickness and volume of mare basalts: 1) analysis of stratigraphic and topographic relationships [1]; 2) depth estimates from flooded craters [2-4]; 3) geophysical techniques such as the use of gravimeters [5,6] and seismometers to probe the lunar subsurface; and 4) crater penetration [7,8]. I have estimated lava thickness by using craters to probe the Moon at depth, using compositional data from the Clementine mission. I focused on craters with diameters in the 18-28 km range which have formed on mare or near mare/highland boundaries. Craters in this size range have depth to diameter (d/D) ratios of about 1/10, indicating that the apparent crater volume approximately equals the volume of the excavation cavity [14]. By analyzing compositional data, I determined if the crater has completely penetrated the overlying layers of basalt and excavated basement highland material. The compositional data for those craters found to puncture the mare were combined with topographic data to reconstruct the crater target and to obtain estimates of the pre-impact basaltic thicknesses.

METHOD

The Clementine spacecraft obtained images at eleven wavelengths from the ultraviolet to the near infrared [9]. I produced color co-registered mosaics using images taken from the UVVIS instrument at wavelengths of 415 nm, 750 nm, and 950 nm. The spatial resolution of the images varied between 100 to 200 m per pixel. From these mosaics, I constructed ratio images where red = 750/415 nm, green = 750/950 nm and blue = 415/750 nm. The ratio images aid analysis by emphasizing differences in composition and maturity [10,11]. The separate bands of the mosaics were used to make maps of FeO and TiO2 composition. I made iron maps using the Lucey et al. method [12] where iron concentrations are calculated using the reflectivity of the 750 and 950 nm images (R750 and R950, respectively). This equation:

\[
\Theta = \tan^{-1} \left\{ \frac{(R_{950}/R_{750}) - 1.26}{R_{750} - 0.01} \right\}
\]  

(1)

yields the spectral parameter \( \Theta \), which closely corresponds to the actual Fe abundance when applied pixel-by-pixel to the mosaics. A similar equation was employed to make the maps of titanium concentration [13]. In the resulting images, the oxides are expressed as
percent by weight. My analysis concentrated on the iron maps because iron dominates the reflectance properties of the lunar surface and the iron weight percentages have smaller relative uncertainties than those for titanium.

Typical highland iron concentrations are vary from less than 3 to more than 10 wt\%, while typical mare concentrations range from 14 to 20 wt\% [12]. For the selected craters, I determined the mean iron abundance of the ejecta blanket by averaging the values in the area from the rim crest out to a distance of about one crater diameter. For analytical purposes, I assumed that the ejecta of the crater was a linear mix of the basement and overlying material, according to the equation:

\[
E = p \cdot M + q \cdot H
\]

In this equation, E is the average weight percent of iron the ejecta, p is the fraction of mare material, M is a representative local mare iron wt\%, q is the fraction of highland material, and H is a representative local highland iron wt\%. With the additional constraint that \( p + q = 1 \), the relative amounts of mare and highland material in the ejecta can be calculated.

Without detailed topographic data, the numbers obtained above only represent relative measurements of the amounts of mare and highlands. The best currently available topographic data is in the form of Lunar Topographic Orthophotomaps (LTOs), which were constructed in the 1970’s using stereo pairs of Apollo metric and panoramic photographs [14]. Several LTOs were digitized to produce digital elevation models for each of the craters. Once digitized, these DTM’s were processed to be of similar resolution to the compositional images. I took a series of 36 profiles radiating from the center of the crater outward and combined them to produce an averaged profile for each crater. From these profiles, the apparent crater depth (d; the difference in elevation from the rim crest to the crater floor) could be determined to a high degree of precision. The thickness, t, of the pre-impact mare basalt is reconstructed by taking the observed fraction of basalt in the crater ejecta, and restoring an excavation cavity of depth d based on the measured profile, in which basalt overlies highland material.

RESULTS

The results for four impact craters are summarized in the following table:

<table>
<thead>
<tr>
<th>Maria</th>
<th>Crater</th>
<th>lat/lon</th>
<th>D, diameter (km)</th>
<th>d, Apparent. Crater Depth (m)</th>
<th>Ejecta Ave. FeO wt%</th>
<th>t, Mare Thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Imbrium</td>
<td>Euler</td>
<td>23°N 29°W</td>
<td>28</td>
<td>2261</td>
<td>15.0</td>
<td>1350</td>
</tr>
<tr>
<td>Imbrium</td>
<td>Timocharis</td>
<td>27°N 13°W</td>
<td>34</td>
<td>2870</td>
<td>13.6</td>
<td>1430</td>
</tr>
<tr>
<td>Crisum</td>
<td>Picard</td>
<td>15°N 55°E</td>
<td>23</td>
<td>2140</td>
<td>15.4</td>
<td>&gt;2140</td>
</tr>
<tr>
<td>Tranquillitatis</td>
<td>Dionysius</td>
<td>3°N 17°E</td>
<td>18</td>
<td>2481</td>
<td>9.7</td>
<td>400</td>
</tr>
</tbody>
</table>
CONCLUSIONS

Looking at these results, this technique seems to provide reasonable values of mare thicknesses. Dionysius, for example, lies on the southwestern boundary of Mare Tranquillitatis. From the iron map, it appears to be almost perfectly bisected by the mare/highland boundary. The low iron signature of highland material on the left half on the interior of the crater contrasts the high iron signature of the right. The large ejecta blanket of Dionysius also follows the same pattern. The value for the mare thickness at Dionysius is quite thin (~400), which is what one would expect near the edge of the mare. All of the thickness estimation techniques mentioned in the introduction yield lesser thicknesses near the edge of basins. Conversely, thicknesses increase as one moves towards the center of the mare. This observation is supported in my data as evidenced by the much greater basalt thicknesses in the targets of craters Euler and Timocharis, which are located closer to the center of their respective maria. The iron map of Picard indicates a paucity of highland material in the ejecta blanket, suggesting that this crater did not excavate to a depth greater than the basaltic thickness. Andre et al. [7] also concluded that the basaltic thickness at Picard exceeds the apparent crater depth.

These results are encouraging, and this method needs to be applied to a larger population of craters (N> 100) to map how the thickness varies laterally. A larger sample of craters might also lead to a better understanding of the effects of such complicating factors such as slumping in the crater, discontinuous ejecta blankets, and inhomogeneous ejecta mixing process.

ACKNOWLEDGMENTS: Cari Corrigan for DTM’s of Picard and Timocharis

REFERENCES

[13] Lucey et al., 1997, JGR, in press