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August 11, 2005 — Houston, Texas
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August 11, 2005
Houston, Texas

2005 Summer Intern Program for Undergraduates
Lunar and Planetary Institute

Sponsored by
Lunar and Planetary Institute
NASA Johnson Space Center
# AGENDA

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<tr>
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<td>Breakfast in LPI Great Room</td>
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| 8:40 | **RYAN ANDERSON, University of Michigan** (Advisor: W. Kiefer)  
Determination of Fill Thickness in Ancient Martian Impact Basins |
| 9:00 | **KATHERINE BURGESS, Earlham College** (Advisor: D. Musselwhite)  
Experimental Petrology of the Martian Meteorite North West Africa 1068 |
| 9:20 | **RUTH CARLEY, University of Cambridge** (Advisors: Drs. E. Heggy, S. Clifford, and R. Morris)  
| 9:40 | **MELANIE CARRIERE, Queens University** (Advisor: T. Stepinski)  
Automated Extraction of Martian Valley Networks |
| 10:00 | **DEBRA HURWITZ, Pomona College** (Advisor: D. Nunes)  
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| 10:20 | **KAORI JOGO, Kyushu University** (Advisor: L. Nyquist)  
$^{53}$Mn-$^{51}$Cr Dating of Chondrules and Chondrites |
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| 11:00 | **JUNICHI MAKISHIMA, University of Tokyo** (Advisor: G. McKay)  
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| 11:40 | **RENEE NAPHAS, Embry-Riddle Aeronautical University** (Advisors: Drs. J. Moses and T. Greathouse)  
Latitudinal/Seasonal Variation of Hydrocarbon Abundances on Uranus and Neptune |
| 12:00 | **KATHERINE NEFF, Indiana University** (Advisor: K. Righter)  
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| 12:20 | **AKIKO TAKAYAMA, Kobe University** (Advisor: L. Keller)  
Mineralogy and Mineral Chemistry of Primitive Interplanetary Dust |
| 12:40 | **ELIZABETH VENECHUK, Scripps College** (Advisors: C. Allen and D. Oehler)  
Layered Rock, Bright-Rim Craters and Faulting in the Arabia Terra, Mars |
| 1:00 | Closing Remarks |
| 1:15 | Adjourn — Group Photos |
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Introduction: The Mars Orbital Laser Altimeter (MOLA), on the Mars Global Surveyor (MGS) spacecraft, has measured the global topography of Mars with a vertical accuracy of ~1 m and a resolution of approximately 1 km at the equator [1]. This data reveals hundreds of roughly circular, low relief topographic basins called Quasi-Circular Depressions (QCDs). Most QCDs do not have an obvious morphologic expression in visible imagery. Figure 1a shows an example Viking image mosaic of a QCD. The circle indicates the topographic rim location. The dashed line indicates the topographic profile shown in Figure 1b. The depression is clearly apparent in topography but is not visible in the image.

The similarity of QCD topography with that of known impact basins suggests that QCDs are also impact structures. The subdued relief of QCDs is interpreted as the result of post-impact deposition of material on the crater floor. The cumulative number versus diameter curve for QCDs follows a power law similar to that for impact structures strengthening the impact interpretation [2,3]. The observed density of QCDs indicates that they formed very early in Martian history [4].

The objective of our study was to constrain post-impact fill thickness in large QCDs. We present measurements of 36 QCDs with diameters from ~120 km to ~675 km. In addition, 5 visible basins with diameters ~350 km to ~880 km are included. Comparison of measured depths with the depths of similar sized, pristine impact structures [5] allows us to estimate the thickness of post-impact fill in these basins. Spatial patterns in the distribution of fill thicknesses contribute to our overall understanding of the depositional history of Mars. In addition, fill thickness of large basins can be used to constrain future gravity models.

Methods: QCDs were chosen from a list of ~560 basins >200km in diameter [3]. Each QCD was studied for evidence of concentric topographic contours and any arcuate ridges, cliffs, or other features

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**Figure 1**: a) Viking image of a QCD. A circle is superimposed on the rim. b) A topographic profile through the QCD, indicated by the dashed line, shows a clear depression.
that could define a rim. Only once a rim was independently defined was it compared with the diameter from the list. In cases where they were very close, the list diameter was used. In cases with a large discrepancy, the QCD was kept only if it was well defined.

Due to time limitations, it was not possible to examine all ~560 QCDs. Our study resulted in measurements of 30 highland and 6 lowland QCDs. The lowland QCDs were less frequent in the list and generally not as well defined, resulting in the difference in sample size between the two regions. We also measured 5 basins previously mapped in visible imagery [6,7]. The same process was used to measure the QCDs and visible basins.

An average rim elevation was found by taking profiles through prominent rim features. By using only the most prominent features, the effect of degradation [8] on the averaged rim height is minimized. A minimum point was chosen for each QCD within a concentric ring half the diameter of the depression. Obvious craters within the QCD were avoided when choosing this minimum point. These precautions assured that the minimum was representative of the QCD floor and not subsequent modification. All measurements were made using the program Gridview [9].

Results: It is well established for both the Moon [10,11,12] and Mars [5,13,14] that fresh impact craters follow a power law relationship between depth and diameter. Recent work gives the power law relation for Martian craters > ~100 km as: \(d = 0.44D^{0.38}\) where \(d\) is depth and \(D\) is diameter [5]. The depth-diameter plot for the QCDs and basins measured is shown in Figure 2.

The depth of the QCDs measured is given by subtracting the minimum elevation from the average rim elevation. All the QCDs measured fell well below the power law for fresh craters, as expected. The lowland QCDs are generally shallower than highland QCDs. The visible basins show a wide variation in depth, with some shallower and others deeper than comparable QCDs.

Subtracting the measured depth from the predicted depth for a pristine crater of the same diameter gives an estimate of fill thickness for the QCDs [11]. This fill thickness appears to have some regional dependence. Of the six lowland QCDs measured, four had >3 km of fill. In contrast, only seven of the 30 highland QCDs had >3 km of fill. An ocean covering the northern lowlands during early Martian history has been proposed [15] but is not universally accepted. Sediment deposition within such an ocean could account for the large fill thickness observed in the lowland QCDs.

Six of the seven highland QCDs with >3 km of fill are near the Argyre impact basin, suggesting that ejecta from that impact may be a significant source of the fill material in that region. The region around Hellas may be similarly buried, but more measurements around the basin must be made before anything conclusive can be said.

The three QCDs with the smallest depth anomaly are clustered in the central southern highlands, about midway between Hellas and Argyre. Two of these three stand out clearly from the rest of the QCDs measured on the depth-diameter plot. They are comparable in depth to visible craters of the same size.
Conclusions: Measurements of depth and diameter show that QCDs fall well below the predicted depth for fresh craters, as expected. The fill thickness in QCDs can be estimated by subtracting measured depths from predicted depths. This fill thickness appears to be greatest in the northern lowlands and in the highlands surrounding Argyre. Both volcanic and sedimentary sources of fill in the northern lowlands have been proposed. Fill surrounding Argyre is likely ejecta from that impact.

This work is based on a small sample of basins from the original ~560. Future measurements will give more complete coverage and therefore a better understanding of the global fill thickness variation. Gravity models of QCDs and large basins will help to constrain fill thickness densities. This will give a better understanding of what depositional processes were important in the early history of Mars.


Figure 2: Depth-diameter plot of QCDs and basins measured. Most lowland QCDs are noticeably shallower than highland QCDs of the same diameter. All but two of the highland QCDs are shallower than 2 km. Visible basin depths are scattered, but are not significantly deeper than QCDs even for very large basins. The solid line is the expected depth for large, pristine craters on Mars. [5]
EXPERIMENTAL PETROLOGY OF THE MARTIAN METEORITE NORTH WEST AFRICA 1068. Kate Burgess1 and Donald Musselwhite2; 1Geosciences Department, Earlham College, Richmond, IN 47374; 2Lunar and Planetary Institute, 3600 Bay Area Blvd., Houston, TX 77058.

Introduction
Martian meteorite North West Africa 1068 (NWA 1068), found in the Moroccan desert in April 2001, is an olivine-phyric shergottite [1]. Present in the meteorite is a fine-grained groundmass and porphyritic olivine grains. Mineral proportions are: 52 vol % pyroxene, 22% maskelynite, 21% olivine, and 5% oxides, sulfides, and phosphates. It has been suggested that the large olivine grains are cumulates. Its major element composition is similar to other olivine phic shergottites, but unlike those, NWA 1068 is undepleted in LREE and has a pattern more like that of Shergotty, Zagami and Los Angeles [2]. Little is known for certain at this time about processes of the Martian mantle. Meteorites that are shown to come from primitive basaltic melts would give clues as to where reservoirs are located and what processes have occurred over time. On the other hand, knowing that a melt is not primitive also would help us to more clearly understand these questions and their answers.

Methods
We used a synthetic glass as our starting material, using the bulk composition of the meteorite [1] (Table 1). The oxides and carbonates were carefully weighed and mixed, then melted and ground again in order to insure a homogenous material. The sample was then put in a gas mixing furnace to set the oxygen fugacity. We used the modeling programs MELTS and PMELTS to estimate where the liquidus of our starting composition was likely to fall and if we could find a multiple saturation point.

Table 1. Bulk composition of NWA 1068. [1] Barrat et al. SiO2 percentage calculated by subtraction. Desired composition calculated by removing 1% CaCO3 due to terrestrial weathering. *Weight percents were recalculated for use of Fe2O3, MnO, CaCO3, Na2CO3, K2CO3, and Ca5(PO4)OH.

<table>
<thead>
<tr>
<th>Oxide</th>
<th>[1] Desired</th>
<th>Weighed Comp.</th>
<th>Actual Oxide Wt.</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO2</td>
<td>46.20</td>
<td>45.77</td>
<td>45.78</td>
</tr>
<tr>
<td>TiO2</td>
<td>0.77</td>
<td>0.78</td>
<td>0.81</td>
</tr>
<tr>
<td>Al2O3</td>
<td>5.75</td>
<td>5.86</td>
<td>5.85</td>
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<tr>
<td>Cr2O3</td>
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<td>0.64</td>
<td>0.64</td>
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<tr>
<td>FeO*</td>
<td>20.48</td>
<td>20.87</td>
<td>23.23</td>
</tr>
<tr>
<td>MnO*</td>
<td>0.46</td>
<td>0.47</td>
<td>0.52</td>
</tr>
<tr>
<td>MgO</td>
<td>16.50</td>
<td>16.81</td>
<td>16.78</td>
</tr>
<tr>
<td>NiO</td>
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<td>0.03</td>
<td>0.04</td>
</tr>
<tr>
<td>CaO*</td>
<td>7.91</td>
<td>7.04</td>
<td>11.53</td>
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<tr>
<td>Na2O*</td>
<td>1.14</td>
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<td>1.93</td>
</tr>
<tr>
<td>K2O*</td>
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<td>0.16</td>
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</tr>
<tr>
<td>P2O5*</td>
<td>-</td>
<td>0.41</td>
<td>1.01</td>
</tr>
<tr>
<td>CO2</td>
<td>100</td>
<td>100</td>
<td>108.41</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>100.00</td>
</tr>
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</table>

We have conducted high pressure and temperature experiments on the bulk composition of this rock. The purpose of this project was to map the liquidus trajectory of the bulk composition and determine which solid phases are stable at Martian upper mantle conditions. We also want to determine if our first formed olivines are similar in composition to those in the meteorite itself.
Results

We analyzed a total of seventeen samples, fourteen from the piston cylinder and three at 1 atm. Table 2 lists each of the experimental runs and their conditions and products. Fourteen of these are considered completely successful runs. The glass portion of sample N12-20 has more MgO than was originally used in the bulk composition, which is not typical of the other experiments and indicates contamination by the bushing. Sample N14-5 appears to have been contaminated by alumina also used in the assembly. Low weight percent totals in the microprobe analysis made us unable to determine a reasonable crystal percentage for sample N14-4 so it was also excluded from further data analysis.

Figure 2a shows a plot of $Mg/D$ vs. $Fe/D$. All but one of our samples falls within 1σ of the correlation line of Jones [5], indicating that the olivine is in equilibrium with the melt. Figure 2b shows a plot of Mg# of olivine versus the percent crystallization. Our data shows a consistent trend of increasing Fo content with decreasing % crystallization with the first formed olivines at Fo81. The dotted line represents the composition of the olivines in the original meteorite [1]. The Fo content of our first formed olivines is close to 10 Fo units above those found in the original rock. Figure 3 is a plot of run products and conditions with inferred phase relationships. At

<table>
<thead>
<tr>
<th>Run Label</th>
<th>T (°C)</th>
<th>P (kbar)</th>
<th>Duration (hours)</th>
<th>Products</th>
<th>% xtl</th>
<th>Mg#</th>
</tr>
</thead>
<tbody>
<tr>
<td>N14-4</td>
<td>1440</td>
<td>14</td>
<td>23.5</td>
<td>ol + px + gl</td>
<td>3</td>
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<td>0.77</td>
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<tr>
<td>N10-13</td>
<td>1420</td>
<td>10</td>
<td>22.5</td>
<td>ol + gl</td>
<td>9.5</td>
<td>0.79</td>
</tr>
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<td>N10-14</td>
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<td>22</td>
<td>ol + gl</td>
<td>11</td>
<td>0.78</td>
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<tr>
<td>N10-15</td>
<td>1480</td>
<td>10</td>
<td>19</td>
<td>ol + gl</td>
<td>7</td>
<td>0.80</td>
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<tr>
<td>N12-16</td>
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<td>12</td>
<td>21.5</td>
<td>ol + gl</td>
<td>2</td>
<td>0.81</td>
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<td>10</td>
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<td>N12-19</td>
<td>1500</td>
<td>12</td>
<td>23</td>
<td>gl</td>
<td></td>
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<tr>
<td>N12-20</td>
<td>1440</td>
<td>12</td>
<td>23</td>
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<td>17.25</td>
<td>ol + px + sp + gl</td>
<td>31</td>
<td>0.75</td>
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</table>

Table 2. Run conditions and products. Italicized samples gave questionable data, as discussed in text. Ol = olivine; px = pyroxene; sp = spinel; gl = glass.

Figure 3. Plot of calculated and experimental results for liquidus of NWA 1068 bulk composition. Multiple saturation is present in both PMELTS and experimental data. Experiments above the liquidus produce only glass, while those between the liquidus and pyroxene-in line produce olivine and glass. Pressure error is within 0.5 kbar. No experiments were run in the pyroxene only field. Gl = glass; ol = olivine; sp = spinel; px = pyroxene.
lower pressures, olivine is the liquidus phase, but the pyroxene-in line converges with the liquidus and crosses at about 16 kbar and temperatures slightly higher than 1520°C. The dashed line indicates that we expect pyroxene to be the liquidus phase above 16 kbar. Three samples also contained spinel, allowing us to enter an experimental spinel-in line. Also shown in the plot are the calculated olivine and pyroxene-in lines for PMELTS [6]. We see the same pattern as in our experimental data, including a multiple saturation point for olivine and pyroxene, although the point is offset by 3 kbar and 100°C.

Conclusions
The results of this project show that the olivine megacrysts present in NWA 1068 are cumulate in origin. As seen in Figure 2b, the Mg/Fe ratio of the bulk has been increased relative to the parent melt by the accumulation of early formed olivine. The settling of olivine could have occurred at any time during or after eruption, before solidification. If the bulk composition was modified by the simple addition of olivine, then the parental melt composition can be derived by subtracting enough olivine to achieve the Mg/Fe ratio necessary to produce olivines with Fo72 on the liquidus. Although this may change the exact point of the convergence, the multiple saturation point should still be at the same general conditions. Future experiments need to be completed at higher pressures and temperatures to verify the pyroxene only field.

Acknowledgements
This project would not have been possible without the help of Don Musselwhite. I would also like to thank Kevin Righter and Loan Le for putting up with me in their labs at JSC.

References

Table 3. Electron microprobe results for abundance of oxide phases in our samples.

<table>
<thead>
<tr>
<th>Sample</th>
<th>SiO₂</th>
<th>Al₂O₃</th>
<th>Cr₂O₃</th>
<th>FeO</th>
<th>MnO</th>
<th>MgO</th>
<th>CaO</th>
<th>Total</th>
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<td>11.98</td>
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</tr>
<tr>
<td>N15-25</td>
<td>45.80</td>
<td>6.47</td>
<td>0.52</td>
<td>19.78</td>
<td>0.46</td>
<td>11.98</td>
<td>8.64</td>
<td>96.95</td>
</tr>
</tbody>
</table>

Figure 2. (a) Partitioning of Fe and Mg in olivine. Trend line: y=0.277x+0.126 [5]. (b) Percent crystal vs. Magnesium number of olivines. Mg# of olivine megacrysts in NWA 1068 is 72 [1].
1. Introduction

Mars is the subject of several radar investigations with the objective to understand the physical properties of the Martian surface and subsurface. Ground penetrating and synthetic aperture radar will potentially be added to ongoing investigations from Earth-based and orbital sounding radars. A primary objective of the ESA-orbital sounding radars, MARSIS (0.5-5Mz) and SHARAD (15-25MHz) is to map the Martian subsurface looking for evidence of subsurface water.

The scientific return from these radar instruments is dependent on our understanding of the electromagnetic characteristics of the areas under investigation. Electromagnetic properties determine the attenuation of the backscattered radar signal. Attenuation of the radar wave is a function of the dielectric loss tangent, which is defined as the ratio of the real and imaginary parts of the dielectric permittivity.

At radar frequencies the dielectric permittivity of dry rocks is mainly controlled by frequency and the soil’s density and mineralogical composition [1,2]. Previous dielectric measurements [3,4] and radar field studies in Martian geological analog environments [5] show that the first few meters of the near surface and subsurface layers can cause significant attenuation of radar waves, especially if iron oxide minerals are present in Mars-like amounts. Such attenuation can screen subsurface features from mapping by radar sounding or imaging experiments. It is therefore essential to acquire an adequate knowledge of the electromagnetic properties of the Martian surface under varying conditions of density and iron oxide content in order to quantify the signal attenuation in the ongoing observations, and to locate sites for optimal deep radar sounding on Mars.

In this study experimental measurements of the complex permittivity and loss tangent were performed for a variety of synthesized and Mars analog soils over a range of frequencies, compositions and densities that match the Martian soil case. The frequency range of 1MHz-1GHz covers the present and future frequencies of different radar experiments and the density range of 1400-2400kgm$^{-3}$ covers the density range of the Martian surface as deduced from thermal inertial data.

2. Preparation of Martian soil analogs

The surface of Mars has been identified as basaltic [6] with a significant amount of iron bearing minerals. Although the exact identity, global distribution and abundance of iron minerals are relatively unknown, there is evidence of isolated areas of crystalline hematite [7] and also evidence for magnetic minerals such as maghemite and magnetite [8]. A variety of Mars soil analogs were prepared and measured to provide permittivity values over a range of compositions representative of the different scenarios of the Martian surface. Table 1 includes list of the samples prepared and measured so far.

Synthetic soils were prepared with the addition of varying amounts of hematite and maghemite to pure silica sand. The component minerals were ground and filtered to give a grain size of <50µm, as observed for the hematite grain size at the Martian surface [9]. These soils allow a parametric study of the effects of the iron oxide content on the dielectric properties of the dust layer.

For a more complete approach, terrestrial basalts and Mars-like soils with complex mineralogy (previously prepared for the calibration of the TES and Mini-TES instruments) were measured. These soils are a closer match to the observed Martian mineralogy and had a grain size <1mm. The powdered samples were dried in an oven for 24-48 hours at 80°C to eliminate residual moisture that would increase the reading of the complex permittivity.
Table 1. Composition of Mars-like soils used in the laboratory electromagnetic characterization

<table>
<thead>
<tr>
<th>ARES Sample ID</th>
<th>Complex analog soils</th>
<th>Synthetic soils (Mass percentage)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JSC Mars-1</td>
<td>Plagioclase, magnetite, olivine</td>
<td>80% Silica 20% Hematite</td>
</tr>
<tr>
<td>HWMK612</td>
<td>Plagioclase, magnetite, olivine</td>
<td>80% Silica 20% Maghemite</td>
</tr>
<tr>
<td>HWMK740</td>
<td>Plagioclase, alunite, smectite, hematite, kaolinite, magnetite</td>
<td>50% Silica 50% Hematite</td>
</tr>
<tr>
<td>HWMK919</td>
<td>Plagioclase, magnetite, olivine, hematite, glass, mica (trace)</td>
<td>50% Silica 50% Maghemite</td>
</tr>
<tr>
<td>HWMK940</td>
<td>Plagioclase, natrojarosite, magnetite, hematite</td>
<td>50% Silica 25%Hematite25%Maghemite</td>
</tr>
<tr>
<td>Basalt (from the Craters of the Moon, Idaho)</td>
<td>&lt;50µm.</td>
<td>80% Silica 10%Hematite10%Maghemite</td>
</tr>
</tbody>
</table>

and loss tangent have been estimated at a maximum of 5% to include equipment errors, and errors in thickness determination and mass measurements.

4. Results and discussion

The variation of loss tangent with density and frequency for a selection of the samples measured is shown in Figures 1 and 2 respectively. Data from below 200 MHz where the error exceeds 8%, has been corrected for band-edge resonance, and in figure 1 data points have been added at a density of 1000 kgm⁻³ for HWMK919 and JSC-Mars1 from loose powder measurements performed using an open APC7 coaxial cell.

The highest loss tangents of 0.02-0.045 are observed for the complex soil samples, and synthetic soils with high concentrations of iron oxide. The high complex permittivity of oxides and the presence of Mars-like amounts (10 to 25%) of magnetic materials increased the loss tangent above that of ordinary basalt.

There is a clear and significant increase in loss tangent with density occurring for all samples at 500 MHz (Figure 2). The behavior with density differs with each sample from being almost constant (JSC-Mars1), close to linear (synthetic soils) to extremely non-linear (basalt) and is thus strongly dependent on the composition. Variation with density is also frequency dependent, will be shown by other figures in the presentation.

The differences in the permittivities of the materials are a response to the different polarizations mechanisms induced by the interaction with the alternating electric field. There are very few materials with a permittivity that is even approximately constant over a frequency range [10] and for combinations of many minerals, each exhibiting different forms of polarization, the total dielectric permittivity is a complicated function of frequency and density and shows no simple relationship with the mineral composition.

Synthetic soils show the expected decrease of loss tangent with frequency due to the linearity of the interfacial polarization. However, complex soils show an increase in loss tangent with frequency caused by an increase in the imaginary part of the dielectric constant. The lack of consistent trend with frequency over the range of samples shows that complex mineralogy is an important factor in determining the variation of loss tangent with frequency.

This investigation shows the spectral complexity of the dielectric properties of Mars analog soils and their variation with density.

Figure 1. Loss tangent as a function of frequency at a compression of 10 tons/cm² (Average density ~2000kgm⁻³)

Figure 2. Loss tangent as a function of density at 500 MHz with ± 5% error bars.
5. Implications

The high values of loss tangents and its dependency on the density shown by this investigation has significant implications for radar investigations of Mars in terms of increasing the signal attenuation in the first few meters in iron-oxide rich consolidated sediments. Loss tangent values used in models to test the performance of Mars radars [11,12] are as much as 10 times lower than those measured here. When converted into 2-way losses and penetration depth, the achievable depth will be much less than predicted even before including scattering effects (which can reduce the range by a half again).

In light of the recent MARSIS data, the need for adequate understanding of the electrical properties of the Martian surface is even more urgent. Further investigations into electrical and also the magnetic properties of more analog soils over a wider range of density and frequency and also with varying conditions of temperature and grain size are required and represent the continuity of this study.

Improved estimates of the Martian dielectric properties are vital to make better models for the design and testing of radar instruments, and also in the interpretation of radar data. Surface dielectric maps could be produced to allow location of the best sites for radar exploration that present the least attenuation for each frequency. The increase in loss tangent with density and iron oxide content makes detection of subsurface water at several kilometers depth [13] beneath these areas with these radar systems unlikely.

References

Mapping and Analyzing Martian Valley Networks
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Advisor: T. F. Stepinski, Lunar and Planetary Institute, 3600 Bay Area Boulevard, Houston, TX, 77058, USA (email: tom@lpi.usra.edu)

I. INTRODUCTION

Numerous missions, including Viking, Mars Global Surveyor and Mars Odyssey, have gathered a wealth of Martian surface images. A common geomorphic feature being studied using this data is the networks of valleys that are found predominantly in the cratered uplands of the southern hemisphere of the planet. By studying these networks we hope to better understand the processes under which they formed, whether by precipitation-fed fluvial erosion or by groundwater sapping [2]. This in turn reveals to us the hydrological history of Mars.

Traditionally, the channels composing valley networks have nearly all been mapped manually from the Viking images. Unfortunately, the relatively low resolution, 1/256 of a degree (0.2314km/pixel), of the images suggested that the valley networks were low-stream-order, immature drainage systems. Hynek (2003) used a combination of data from the Mars Orbiter Camera (resolution also 1/256) and Mars Orbiter Laser Altimeter (MOLA) digital topography to find that many previously mapped segments are actually part of larger, more mature drainage networks [1]. The problem with manual mapping is that it becomes impractical to do so for large geographic regions. Therefore, we are currently developing an algorithm to extract these channels by automation.

In this paper, we manually determine the drainage density of channels of eight sites on Mars, listed in Table 1, using a combination of Thermal Emission Imaging System (THEMIS) and MOLA data sets. We then compare this to the drainage density obtained by automatically extracting the channels from MOLA data sets. Additionally, for one site, we compare our results to those of Hynek.

II. DATA

Our data consists of daytime infrared images of 100 meters/pixel resolution from THEMIS, which has ten different filters between 6.78 and 14.88 micrometers available. Out of convenience, our mapping was done using only the first band (6.78 micrometers). Additionally, we used digital topography from MOLA, at a resolution of 1/128 of a degree. The use of the THEMIS data set allows for better resolution and better quality images than the data used by Hynek, while the MOLA data set allows for entire site coverage and easier interpretation of the THEMIS images. The MOLA data set also allows automated extraction of the channels to be performed.

III. METHODS

To manually map the channels, we first created a mosaic of the image strips from THEMIS covering the area of interest using ISIS, free software provided by USGS. Then, the MOLA digital topography was overlaid onto the THEMIS mosaic in ArcMap. The channels were mapped using principally the THEMIS images and the MOLA digital topography as support. Once the valley networks had been determined, the individual drainage basins were delineated computationally using the digital topography. Finally, for each basin, we calculated the basin area ($A$) and total length of channels ($L_T$) in that area to determine the drainage density ($D_D$).

$$D_D = \frac{L_T}{A}$$ (1)

Manually mapped valley networks are then compared to those obtained by an automated process that is still in development. For each major basin located in the site of interest, we use the same process as for the manual mapping to determine the drainage density of the valley networks extracted by automation.
IV. RESULTS

Manual and automated mapping were performed for eight sites. Of these, Naktong Vallis has also previously been mapped by Hynek, for which he determined a drainage density of 0.082 km\(^{-1}\) [1]. It is necessary to mention, however, that he did not determine drainage densities for individual basins, but rather, he used the entire site area and the total length of channels found in the site as variables in equation 1. Comparisons between manual and automated mapping of the Dawes East region are shown in Figure 1. Comparisons between our drainage network and the one obtained by Hynek for Naktong Vallis are shown in Figure 2. Finally, Table 1 lists the drainage density obtained for all 8 sites.

<table>
<thead>
<tr>
<th>Site Coordinates</th>
<th>Manual DD (km(^{-1})</th>
<th>Automated DD (km(^{-1})</th>
</tr>
</thead>
<tbody>
<tr>
<td>Locras (6,45,11,49)</td>
<td>0.094</td>
<td>0.049</td>
</tr>
<tr>
<td>Pollack (-11,28,-7,35)</td>
<td>0.069</td>
<td>0.033</td>
</tr>
<tr>
<td>Evros Vallis W. (-16,11,-10,18)</td>
<td>0.055</td>
<td>0.035</td>
</tr>
<tr>
<td>Dawes E. (-8,43,-3,48)</td>
<td>0.077</td>
<td>0.040</td>
</tr>
<tr>
<td>Millochou W. (-23,85,-17,92)</td>
<td>0.076</td>
<td>0.034</td>
</tr>
<tr>
<td>Evros Vallis N. (-11,12,-5,19)</td>
<td>0.068</td>
<td>0.045</td>
</tr>
<tr>
<td>Millochou S. (-26,85,-20,91)</td>
<td>0.079</td>
<td>0.035</td>
</tr>
<tr>
<td>Naktong Vallis (1,28,6,36)</td>
<td>0.043</td>
<td>0.038</td>
</tr>
</tbody>
</table>

Table 1: Site characteristics and drainage densities obtained by both manual and automated methods. Coordinates are in the form (lower left latitude, lower left longitude, upper right latitude, upper right longitude).

V. DISCUSSION AND CONCLUSIONS

This study found a good general agreement between the manual and automated mapping of drainage networks. Most differences between the two are found simply in the details. However, the drainage density was systematically lower for channels extracted by automation than for channels mapped manually. This is easily understood since the extraction algorithm uses the lower resolution MOLA digital topography compared to the THEMIS images, which were used for the manual mapping. Use of higher resolution digital topography, once it becomes available, would therefore help improve the automated extraction procedure.

In the case of Naktong Vallis, we also compare our results to those of Hynek. Once again, there is good general agreement between the two and differences only in the details, though he finds a drainage density that is nearly twice as large as ours. However, since Hynek used lower resolution images and did not use the same method to determine the drainage density, we should have obtained an equivalent or higher drainage density than he did. Although we are not aware of his selection criteria, we can still agree with his conclusion that using higher resolution images than the Viking images allows us to detect larger, more mature drainage networks. However, using images of resolution as high as THEMIS images is seemingly not as important as having good quality images that have optimal site coverage.

ACKNOWLEDGMENTS

I would like to thank Dr. Stepinski for his support and guidance and Ian Molloy for all of his hard work and patience.

REFERENCES

Figure 1: Comparison of valley networks determined by manual and automated methods. A: Manually mapped valley networks for the Dawes East region. B: Automated extraction of valley networks for the Dawes East region. Fewer channels are in evidence from the automated method.

Figure 2: Comparison of valley networks for the Naktong Vallis region. A: Manually mapped valley networks from this study. B: Manually mapped valley networks as determined by Hynek. This figure is adopted from [1]. Note that the extents are not quite the same. More channels are in evidence for the same region as mapped by Hynek.
NEW OBSERVATIONS OF THE NORTH POLAR ICE CAP OF MARS USING THEMIS-VIS. D. M. Hurwitz and D. C. Nunes. Lunar Planetary Institute, Houston, TX.

Introduction

The northern Martian polar ice cap extends from approximately 78°N to 90°N and is made up of two major units that likely differ in composition, appearance, and age (Byrne and Murray, 2002). The upper, younger unit is approximately 2500 m thick and is composed mainly of water ice and dust. This unit appears as finely layered deposits with variable layer thickness, steep scarps and smooth outcrops. The lower, older unit is at least 600 m thick and has a platy appearance with jagged outcrops. This unit acts as at least part of the source of the material that makes up the nearby polar dunes, though the ratio of silicate particulates to ice in this unit is not constrained. The thickness estimates of these two units are derived from observations focused at the margins of the cap and are approximations, especially with the lower unit, the basement of which is not visible (Byrne and Murray, 2002). The lower unit is believed to extend beneath part of the upper unit only, leaving a portion of the upper unit in direct contact with the Hesperian-aged surface unit, the Vastitas Borealis Formation (VBF) (e.g., Fishbaugh and Head, 2005).

The surface of the northern polar cap has previously been identified as smooth, lacking impact craters, but possessing a series of troughs that spiral outward from the center of the cap (Herkenhoff and Plaut, 1999). These conclusions were made using Viking Orbiter and Mariner 9 images, which cover approximately 70% of the cap with relatively low resolution (231 m/px with selected areas of 58 m/px, covering about 70% of the Martian surface). Our study uses the more recent THEMIS–VIS data, which includes images with resolutions between 20 and 40 m/px and cover nearly the entire northern ice cap (except latitudes >88°). Our goal is to i) verify Herkenhoff and Plaut’s conclusions on the dearth of craters on the perennial ice, and ii) to generate updated observational constraints on the deformation of the perennial ice that can support more updated thermal and mechanical models of the ice cap. Previous geophysical models of the Martian polar caps did not incorporate the recently-identified lower unit, which can possibly alter the thermo-mechanical state of the overlying ice.

Methods

A map of the northern Martian ice cap was generated in ArcMap using MOLA gridded elevation data (MEGT_N_512) with resolution 115 m/pixel. THEMIS-VIS images with resolution between 20 and 40 m/pixel and illumination angles between 16° and 23° were mosaicked and superposed onto the elevation map; Figure 1. We used ISIS to calibrate the THEMIS data and to place the dataset in a polar stereographic projection. When processing the images, we used band #3 of the 5 available visible bands because it offers the greatest brightness contrast. Some of the resulting mosaics had geographic alignment errors of a few hundred meters.

We surveyed both datasets in search of impact craters and other morphological features that could
possibly constrain the deformation of the ice regionally and lead to better characterization of the thermo-mechanical state of the ice cap.

**Discussion**

Our investigation of the THEMIS images on the MOLA gridded elevation map covers approximately 60% of the ice cap, and our observations support Herkenhoff and Plaut’s conclusions that craters of all sizes are mostly absent from the surface of the perennial ice cap. Initial observations of the MOLA grid alone suggested the presence of many small (< 1 km) craters on the surface of the ice cap. However, the THEMIS–VIS images show that many of these “craters” are only artifacts of the elevation grid. We did observe additional topographic elements in two forms: impact craters on the floor of polar troughs and possible fissures on the surface of the ice cap.

**Craters**

The craters that we found are mostly concentrated on the floors of the spiraling troughs between the mapped longitudes of 0-20°E and 130-140°E. Investigation of troughs in other locations might reveal craters as well. Troughs at the specified longitudes theoretically overly the VBF unit directly, and they might be deep enough to penetrate the upper layer of ice and reach the underlying unit. Fishbaugh and Head [2004] made similar observations of a concentration of craters on the floor of Chasma Boreale. It is not clear, however, if the floor of Chasma Boreale corresponds to the bottom unit of the cap or the VBF.

The lack of craters on the upper unit of the ice cap supports Herkenhoff and Plaut’s estimate for the surface age of at most 100 kyr. The craters in the troughs might be used to approximate an age for the underlying units.

**Possible Fissures**

Lineaments were found between the spiraling troughs on the surface of the ice cap between longitudes 135°E and 150°E. These lineaments occur on the crest of the surface undulations. Elevation profiles of these lineaments indicate approximate depths of 10 m and widths of 100-500 m, and measurements of these lineaments indicate approximate lengths of 5-10 km. In some cases, as in Figure 2, several segments appear in a row and form groups nearly 100 kilometers long that follow the crests of undulations.

We interpret these lineaments as normal faults resulting from brittle deformation under tensile stresses. It is not clear, however, if the brittle deformation is a result of the processes forming the undulations or if the undulations serve as a focusing mechanism for regional stresses. Previous works have suggested a combination of insolation and katabatic winds behind the formation of troughs and undulations (Pelletier, 2004; Thomas et al., 1992). In any case, brittle deformation in conjunction with its relatively small width and depth is suggestive of ductile deformation occurring beneath a thin competent layer.

In our observations, these fissures are found in longitudes where the bottom unit underlies the ice, but they are absent from longitudes 0-90°E that overlie the VBF. Further mapping should validate or refute this as a cap-wide trend.

![Figure 2](image-url)
Future Work

More images from this data release should be studied, and identified craters and fissures should be verified in order to support the stated observations. In addition, troughs that theoretically overly the lower unit should be searched for craters to compare crater concentrations with the troughs that overly the VBF. In addition, images should be searched for more fissures with the goal of investigating their distribution to characterize the mechanical state throughout the cap. These observations may be used to generate constraints for future mechanical and thermal models of the cap.

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$^{53}$Mn-$^{53}$Cr DATING OF CHONDRULES AND CHONDRITES. K. Jogo¹, L.E. Nyquist², C.-Y. Shih³ and Y. Reese⁴, *E-mail address: kaori@geo.kyushu-u.ac.jp. Department of Earth and Planetary Science, Kyushu University, Hakoizaki, Fukuoka 812-8581, Japan, ²Mail Code KR, NASA Johnson Space Center, Houston, TX 77058, USA, ³Mail Code JE-23, ESCG/Muniz Engineering, Houston, TX 77058, USA.

Introduction: Chondrules and chondrites are interpreted as objects formed in the early solar system, and it is important to study them in order to elucidate the evolution of the early solar system. Here, we report the study of the Mn-Cr systematics of individual Chainpur (LL3.4) chondrules, the chondrite Tagish Lake, the R-Chondrite NWA753, and the CO-chondrite ALH83108, and compare the results to present chondrule and chondrite data.

The first goal was to verify and improve the present chondrule formation age as dated by the $^{53}$Mn-$^{53}$Cr method [1]. The present Chainpur chondrule formation age using the Mn-Cr method appears to be nearly as old as CAI ages. However, the CAIs have been interpreted as the oldest solar system objects by other chronological studies of CAIs and chondrules [2][3][4][5][6].

Fig. 1 shows that measured $^{53}$Cr/$^{52}$Cr correlates with $^{55}$Mn/$^{52}$Cr to give initial ($^{53}$Mn/$^{55}$Mn)$_i$ = (9.4 ± 1.7) × 10$^{-6}$ for Chainpur chondrules [1]. Fig. 2 shows that $^{53}$Cr/$^{52}$Cr correlates with $^{55}$Mn/$^{52}$Cr to give initial ($^{53}$Mn/$^{55}$Mn)$_i$ = (5.8 ± 1.9) × 10$^{-6}$ for Semarkona chondrules [7]. The values of ($^{53}$Mn/$^{55}$Mn)$_i$ are obtained from the slopes of the isochrons in Figures 1 and 2. Although these values are expected to be the same, they are slightly different. There is a corresponding difference in the ordinate intercepts of the isochrons as dated by the Mn-Cr method [1]. The present Chainpur chondrule formation age using the Mn-Cr method is nearly as old as CAI ages. However, the CAIs have been interpreted as objects formed in the early solar system. Here, we report the study of the Mn-Cr systematics of individual Chainpur (LL3.4) chondrules, the chondrite Tagish Lake, the R-Chondrite NWA753, and the CO-chondrite ALH83108, and compare the results to present chondrule and chondrite data.

The second goal was to determine Cr isotopic and age variations among chondrite groups with different O-isotope signatures (Fig. 3). The $^{53}$Mn-$^{53}$Cr method is based on the assumption that $^{53}$Mn was initially homogeneously distributed in the solar nebula [9][10]. However, different groups of chondrites were formed from regions of different O-isotope composition (Fig. 3). So different types of chondrites may also have different initial $^{53}$Mn abundances and Cr isotopic compositions. Thus, it is important to determine Cr isotopic ratios among chondrite groups. We are studying Tagish Lake, R-Chondrite NWA753, and CO-chondrite ALH83108, which have very distinct O-isotope compositions as shown in Fig. 3.

Procedure (Chainpur Chondrules): We separated chondrules from a bulk chondrite sample using the Freeze-Thaw-Sonication method. We started with 0.5 g of bulk Chainpur sample. First, we froze it in liquid N$_2$ and thawed it on a hot plate. Then, we sonicated it. After

Fig. 1. $\epsilon^{53}$Cr vs. $^{55}$Mn/$^{52}$Cr evolution diagram for Chainpur chondrules, bulk LL, L, and H chondrites. The slope of the isochron defined by the chondrules and bulk chondrites corresponds to initial $^{53}$Mn/$^{55}$Mn$_i$ = (9.4 ± 1.7) × 10$^{-6}$ and $\epsilon^{53}$Cr$_i$ = -0.15 ± 0.13 at the time of formation of these chondrules or their precursors [1].
repeating this cycle 3 times, 24 chondrules were separated. Most are greater than 1 mg and are suitable for Mn-Cr dating. We broke open all of the chondrules and picked up 4. Following the previous Chainpur study, the interiors of the 4 chondrules were characterized by a scanning electron microscope.

Chondrule Classification: Figure 4a shows the broken surface of Chainpur Chondrule C-39, a porphyritic olivine (PO) chondrule, Fig. 4b shows the broken surface of Chainpur Chondrule C-44, a radial pyroxene (RP) chondrule, and Fig. 4c shows the broken surface of Chainpur Chondrule C-46, a barred olivine (BO) chondrule. The subequant, dark gray crystals in Fig. 3a are olivine phenocrysts, and the lighter gray, interstitial material is low-Ca pyroxene and mesostasis.

Procedure and Results (Bulk Chondrites): We powdered the bulk chondrite samples and dissolved them. Cr was chemically separated by a cation exchange method. We took 10% of the solution for Mn and Cr concentration determination by ICP-MS, and 90% of the solution for Cr-isotopic measurement by a mass spectrometer. $\varepsilon^{53}$Cr for NWA753 is given in Table 1. We measured $\varepsilon^{53}$Cr = 0.48 ± 0.13 for NWA753. Cr-isotopic measurements for Tagish Lake and ALH83108 (CO) are in progress, as are Mn-Cr measurements for all of these samples. $^{55}$Mn/$^{52}$Cr = 0.71 for NWA753 was calculated from literature values of Mn and Cr [11]. The data were corrected for mass fractionation using the terrestrial $^{50}$Cr/$^{52}$Cr and $^{54}$Cr/$^{52}$Cr ratio.

**Table 1.** $\varepsilon^{53}$Cr and $^{55}$Mn/$^{52}$Cr data for NWA753 (this investigation), CM, CI, CV, CR, H chondrite [7], Tagish Lake [12] and Kainsaz[11].

<table>
<thead>
<tr>
<th>Sample</th>
<th>Type</th>
<th>$\varepsilon^{53}$Cr</th>
<th>$^{55}$Mn/$^{52}$Cr</th>
</tr>
</thead>
<tbody>
<tr>
<td>NWA753</td>
<td>R</td>
<td>0.48 ± 0.13</td>
<td>0.71</td>
</tr>
<tr>
<td>Murray</td>
<td>CM</td>
<td>0.23 ± 0.04</td>
<td>0.68 ± 0.03</td>
</tr>
<tr>
<td>Murchison</td>
<td>CM</td>
<td>0.17 ± 0.14</td>
<td>0.69 ± 0.04</td>
</tr>
<tr>
<td>Orgueil</td>
<td>CI</td>
<td>0.27 ± 0.11</td>
<td>0.81 ± 0.04</td>
</tr>
<tr>
<td>Allende</td>
<td>CV</td>
<td>0.10 ± 0.05</td>
<td>0.49 ± 0.03</td>
</tr>
<tr>
<td>A881595</td>
<td>CR</td>
<td>-0.04 ± 0.20</td>
<td>0.38 ± 0.02</td>
</tr>
<tr>
<td>Tieshitz</td>
<td>H</td>
<td>0.23 ± 0.05</td>
<td>0.80 ± 0.02</td>
</tr>
<tr>
<td>Tagish Lake</td>
<td>Cl or CM</td>
<td>0.25 ± 0.10</td>
<td>0.67</td>
</tr>
<tr>
<td>Kainsaz</td>
<td>CO</td>
<td>0.20 ± 0.13</td>
<td>0.54</td>
</tr>
</tbody>
</table>

Discussion (Bulk Chondrites): Yamashita et al. [12] obtained $\varepsilon^{53}$Cr and $^{55}$Mn/$^{52}$Cr of 0.25 ± 0.10 and 0.67 from Tagish Lake and Shukolyukov et al. [13] obtained $\varepsilon^{53}$Cr and a $^{55}$Mn/$^{52}$Cr of 0.20 ± 0.13 and 0.54 from Kainsaz (CO) (Table 1). The data points for Tagish Lake[12] and Kainsaz[13] fall on an isochron for C-chondrites as analysed in the JSC laboratory (Fig. 5). The new $^{55}$Mn/$^{55}$Mn ratio is $(7.2 ± 3.0) \times 10^{-6}$ with an initial $^{53}$Cr/$^{52}$Cr ratio at the time of Mn/Cr fractionation of -0.21 ± 0.17. The corresponding time of Mn/Cr fractionation is $T_{LEW} = 4563-4569$ Ma, relative to the time of igneous crystallization of the Lewis Cliff (LEW) 86010 angrite. This time interval includes the 4567.2 ± 0.6 Ma Pb-Pb age of CAI [14]. That data for Tagish Lake and Kainsaz plot on the same isochron may suggest that different types of carbonaceous chondrites (CI, CM, CV, CR, CO, Tagish Lake) have the same initial $^{53}$Mn abundances and Cr isotopic compositions.
The data point for NWA753 falls on the isochron for the C-chondrites [7] (Fig. 6). Data for L and LL chondrites also appear to fall slightly above the C chondrite isochron, whereas data for H chondrites are in good agreement with it. Thus, each chondrite group may have its own correlation line. Also, different types of chondrites appear to have different Cr-isotopic compositions corresponding to different O-isotope compositions (Fig. 3).

Conclusions: Each chondrite group may have its own correlation line. This suggests that different types of chondrites have different Cr-isotopic compositions corresponding to different O-isotope compositions. This finding could result from an initially heterogeneous distribution of $^{55}$Mn in the solar nebula as favored by Lugmair et al. [9] and Lugmair and Shukolyukov [10]. Alternatively, initial $^{53}$Cr$/^{52}$Cr may have been heterogeneous. Distinguishing these possibilities requires that complete isochrons be determined for each O-isotope group.

Future Works: We will get the mass spectrometric data of Chainpur chondrules after this summer internship is completed. We will also get INAA data for the chondrules (courtesy of D. Mittlefehldt). After all of the analyses are complete, we will know whether the previous Chainpur chondrule isochron is confirmed, or whether the Semarkona isochron is more reliable. Additional data for Tagish Lake and ALH83108 (CO), as well as for ALH77307 (CO, requested) should refine the C chondrite isochron, and determine how well it agrees with the chondrule isochrones. A value age of $8.5 \times 10^4$ [10][13] for $^{55}$Mn/$^{56}$Mn of C chondrites implies an absolute age of 4571 Ma, slightly greater than that of CAIs. Finally, additional measurements of NWA753 and other R chondrites should firmly establish whether these chondrites have high $^{53}$Cr/$^{52}$Cr causing from to plot above the C chondrite isochron, in contrast to E chondrites which have low $^{53}$Cr/$^{52}$Cr [15].


Acknowledgements: We thank Sue Wentworth for the SEM data for the chondrules.
CALIBRATION OF THE EU OXYBAROMETER FORNAKHLITES. J. Makishima¹, G. Mckay² and L. Le³
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Introduction: Martian meteorites have various characteristics, which are clues to understand the petrogenesis of Mars rocks. Recently Wadhwa [1] and Herd et al. [2] reported some important characteristics of Martian basalts. They concluded that Martian basalts were formed under a range of oxidation states, suggesting complex petrogenesis processes. Also they reported that the correlation between oxidation states and certain geochemical parameters (initial Sr and Nd isotopic compositions, La/Yb ratios) indicates the presence of long-term incompatible-element enriched and depleted reservoirs, which are oxidized and reduced, respectively. Therefore the variation in oxidation state among the Martian meteorites must have important implications for redox conditions of Martian crust/mantle and overall differentiation on Mars. The nakhlites, which have rather different characteristics from basaltic shegottites, may give us additional clues to Martian petrogenesis.

The oxidation states of meteorites are usually described by the oxygen fugacity (fO₂) under which the meteorites crystallized. One of the methods to estimate the oxygen fugacity is the depth of Eu anomaly. Eu²⁺/Eu³⁺ is determined by the oxygen fugacity and partitioning is different for Eu²⁺ and Eu³⁺. Then the depth of Eu anomaly in a mineral is regarded as a function of the oxygen fugacity. The calibration of the Eu oxybarometer has been done with the basaltic shergottites before [4]. However it has never been applied to nakhlites (Oe.et al. measured the depth of Eu anomaly in the synthetic pyroxene only at QFM[5]). The composition of the nakhlite pyroxene is rather different from the basaltic shergottite pyroxene and partition coefficients are strongly affected by phase compositions, especially pyroxene Ca content. It may be impossible to apply the calibration for the basaltic shergottite [6] to the Nakhlite. Thus we report in this abstract the experimental calibration of the depth of Eu anomaly in pyroxene vs. oxygen fugacity for nakhlites.

Experimental Methods: In order to measure the Eu anomaly in the nakhlite pyroxene, we used the synthetic nakhlite parent melt composition, which our group has studied previously [7,8]. In this study we used the NJ4 composition, which crystallizes pyroxene that are most similar to the pyroxene composition in the nakhlites. This composition was doped with 1.0% Sr, 0.7% Sm, 1.0% Eu and 0.7% Gd. ~125mg pellets of these composition were put on the Pt loops and held in CO-CO₂ gas mixing DelTec furnaces at 1300ºC for 48 hours in order to be homogenized, and then quenched. Homogenized charges were placed back into the furnaces slightly below the liquidus of pyroxene, and held at constant temperature or cooled down at 0.5 ºC/hr to various temperatures, growing pyroxene crystals. Experiments were conducted under the 3 oxygen fugacities, QFM, IW+1.5 and IW. We analyzed quenched charges with the Camaca SX-100 Electron Microprobe at the NASA/JSC.

Results and Discussion: One of the charges which we grew in our experiments and an example of its line-analyses are shown in Fig.1. They contain some pyroxene crystals and glass. Crystals are about 50-100µm across.

Fig.1 The BSE image of an experimental NJ4-817 charge, which was grown at IW. We can see the pyroxene crystal in the charge. The small image shows an example of line-analyses that we conducted in our experiments. The charge is about 4mm across.
After we grew the synthetic pyroxenes so that they resembled the cumulus augite in the nakhlites, we conducted the line-analyses of each charge with the electron microprobe, measuring the amount of major and minor elements in the pyroxene and glass.

We analyzed charges from three experiments, at oxygen fugacities of IW, IW+1.5 and QFM (IW+3.5). Experimental conditions are shown in Table 1. Also pyroxene compositions of our experiments (NJ4-817, 818 and 819) are shown in Fig. 2, in which we compare to the composition of Nakhla pyroxene cores. Experimental pyroxene compositions are uniform, varying generally from En_{37-40} Fs_{19-23} Wo_{38-42}. Although we have analyzed only ~20 pyroxene crystals and we do not know yet whether we have correct synthetic starting composition of nakhlites, we can see a general agreement between experimental pyroxene compositions and Nakhla pyroxene core compositions. We also compared our partition coefficients with the REE partition coefficient pattern of Oe et al. (Fig.3) [9]. Our D values of Sr, Sm, Eu and Gd are all higher than the those of Oe’s. We do not know the reason for this and further experiments will be required to investigate. We can see the deeper Eu anomaly in NJ4-817 and 818 than in NJ4-819, as expected under more reducing conditions.

Table 1. Cooling histories for experiments

<table>
<thead>
<tr>
<th>Expt.</th>
<th>Oxygen fugacity</th>
<th>Start cooling</th>
<th>Quench temperature</th>
<th>Cooling rate</th>
</tr>
</thead>
<tbody>
<tr>
<td>817</td>
<td>IW</td>
<td>1155°C</td>
<td>1150°C</td>
<td>0.5°C/h</td>
</tr>
<tr>
<td>818</td>
<td>IW+1.5</td>
<td>1155°C</td>
<td>1150°C</td>
<td>0.5°C/h</td>
</tr>
<tr>
<td>819</td>
<td>QFM</td>
<td>1159°C</td>
<td>1152°C</td>
<td>0.5°C/h</td>
</tr>
</tbody>
</table>

Table 2 D values

<table>
<thead>
<tr>
<th>Expt</th>
<th>f(O2)</th>
<th>D(Sr)</th>
<th>D(Sm)</th>
<th>D(Eu)</th>
<th>D(Gd)</th>
<th>D(Eu/Gd)</th>
<th>D(Eu/Sm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>817</td>
<td>IW</td>
<td>0.099</td>
<td>0.137</td>
<td>0.131</td>
<td>0.206</td>
<td>0.639</td>
<td>0.960</td>
</tr>
<tr>
<td>818</td>
<td>IW+1.5</td>
<td>0.098</td>
<td>0.132</td>
<td>0.137</td>
<td>0.180</td>
<td>0.761</td>
<td>1.034</td>
</tr>
<tr>
<td>819</td>
<td>QFM</td>
<td>0.100</td>
<td>0.117</td>
<td>0.150</td>
<td>0.175</td>
<td>0.856</td>
<td>1.276</td>
</tr>
</tbody>
</table>

Table 2 shows the D values of the trace elements and their ratios which we calculated from the experimental results. Fig. 4 shows the result of the calibration of the depth of the Eu anomaly in pyroxene vs. oxygen fugacity for nakhlites. In our result the D(Eu)/D(Gd) and D(Eu)/D(Sm) have somewhat different trends. We regard the D(Eu)/D(Gd) as more reliable than the other. There are two reason for this. One is that we doped Eu and Gd together in one charge in our experiments, whereas doped Sm in another chage. Eu and Gd were more similarly affected in their crystallization than Gd. Another reason is the comparison with the calibration of the Eu oxybarometer for the shergottites [6]. McCanta conducted similar experiments to ours, using the synthetic shergottite melt compositions. The trend of the D(Eu)/D(Gd) in Fig. 5 is more similar to those of [6]. Thus, We estimated the oxygen fugacity of the
nakhlite parent melt using the D(Eu)/D(Gd) trend we have obtained in this experiments.

Table 3 shows a comparison of our and Oe’s D(Eu)/D(Gd) values and calculated D(Eu)/D(Gd) values of the nakhlite melt from the most reliable analyses of the Nakhla meteorite. The D ratio of Nakhla melt is between that of NJ4-817 (grown at IW) and NJ4-818 (grown at IW+1.5). Thus, we can say that Nakhla has a possibility that it was crystallized at IW~IW+1.5 rather than at QFM.

**Summary and Conclusion:** One of the most important characteristics of a planet is the oxidation states of its mantle. To re-produce a rock of the planet in experimental methods is very useful to know the oxidation state. One method to estimate the oxidation state of rocks is the Eu oxybarometer. The depth of the Eu anomaly is a function of oxygen fugacity. In this study we conducted experiments to calibrate the Eu oxybarometer for Nakhlite, one of the martian meteorites. We actually grew the synthetic pyroxenes resembling the cumulus augite in nakhlitites under a series of oxygen fugacities, using a most similar starting composition to that of nakhlite melts which we have studied before. From our results, we suggest that the nakhlite may have been crystallized under IW~IW+1.5. This is because of the results of the Eu oxybarometer. That is to say, the D(Eu)/D(Gd) value of Nakhla melt, which we calculated from reliable analyses of a Nakhla meteorite, is between the D(Eu)/D(Gd) values of our experimental rocks grown at IW and at IW+1.5. However, this conclusion is only preliminary. We need to check reproducibility by running more experiments. We should also check why there is a difference in some values of our results, especially the values related to Sm, and modify our experimental methods to obtain more precise data. In addition to this, we are going to investigate Al effects on the calibration because there is clear effects on partition coefficients as our group proved previously [9].

<table>
<thead>
<tr>
<th>Expt</th>
<th>fO2</th>
<th>D(px)</th>
<th>Nakhla melt</th>
</tr>
</thead>
<tbody>
<tr>
<td>817</td>
<td>IW</td>
<td>0.639</td>
<td></td>
</tr>
<tr>
<td>818</td>
<td>IW+1.5</td>
<td>0.761</td>
<td>0.693</td>
</tr>
<tr>
<td>819</td>
<td>QFM</td>
<td>0.856</td>
<td></td>
</tr>
<tr>
<td>Oe</td>
<td>QFM</td>
<td>0.859</td>
<td></td>
</tr>
</tbody>
</table>


**Acknowledgements:** I thank Gordon McKay and Loan Le for their support and guidance through my thesis. Gordon McKay calculated the D(Eu)/E(Gd) of nakhlite melts. Loan Le conducted almost all the electron microprobe analyses. She will also conduct more experiments to further this project. I also thank people who helped me to conduct experiments and analyses.
Classification and Orientation of Martian Gullies to Determine Origin

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1. Introduction

Malin and Edgett [2000] discovered gully landforms (Fig. 1) on steep slopes of Mars, using imagery from the Mars Orbital Camera (MOC) on the Mars Global Surveyor (MGS) spacecraft. Their gullies are composite landforms that represent movement of solid debris from the top of a slope, through a channel, to be deposited near the base of the slope.

Malin and Edgett [2000] inferred that gullies represent recent movement of debris lubricated by groundwater seeping from inside the slope. However, there is little agreement on the source of the water, or even if water is involved in gully formation. Malin and Edgett [2000] seem to suggest that the groundwater exists in long-term aquifers; Hartmann [2001] suggested that groundwater forms as lava melts ground ice, while Gaidos [2001] argues that the water ‘erupts’ from depth. These hypotheses implicitly suggest relatively pure water, while other studies infer that the water must have been highly saline brines (Doran and Forman [2000], Knauth et al. [2000], Wynn-Williams et al. [2001], and Knauth and Burt [2002]). On the other hand, several workers claim that solar heating is adequate, at the proper Martian obliquity, to melt ground ice to form gullies (Mellon and Phillips [2001]; Costard et al. [2002]: Heldmann [2003]). Similarly, Christensen [2003] would have the gullies form from melted snowpack rather than ground ice. Other workers, conversely, claim that liquid CO₂ is the fluid lubricant of choice in gully formation (Musselwhite et al. [2001], Draper et al [2000], Hoffman [2000, 2001]). And Treiman [2003] suggested that gully landforms could form without liquid water or CO₂, by flow of fine wind-deposited sediment.

If the gullies formed by liquid water, no matter what mechanism, it would have significant implications for astrobiology, planetary evolution, and human exploration. The earlier studies, with widely differing results, have relied on data from individual gully occurrences or on global-scale averages of properties. In this work, we focus on gullies at a regional scale, and seek relations among the properties of gullies with the local and regional geographic, geologic, and environmental contexts.

2. Methodology

We investigated gullies in the region of 154°W - 172°W and 30°S - 65°S. This area exhibits gullies on a wide variety of terrain types and elevations. Narrow angle images from the MOC were obtained from Malin Space Science’s website (www.msss.com). After searching these images for gullies¹, the individual features of the gullies were classified. Malin and Edgett [2000] classified the anatomy of gullies into three primary structures: alcoves, “theatre-shaped depressions within the wall or slope on which they have formed”; channels, “V-shaped channels emanating from the down slope apex of the head alcove”; and depositional aprons (referred to as fans from now on), consisting of a fan shaped deposit of sediments at the base of a channel. Malin and Edgett [2000] also described four types of alcoves: lengthened, widened, occupied, and abbreviated. We added two more types: badlands and bare slope (no alcove). We also classified gullies by alcove location (top slope, rock layer, smooth), channel type (short, long, straight, crooked, widened, narrow, meandering, multiple channels/tributaries, braided), depositional structure (fan, digits, nodule, not present), gully direction (8 compass points), and pole/equator facing orientation.

Fig. 1. Pole Facing gullies on a crater wall from 167°W, 38°S.
A. Alcove  B. Channel  C. Depositional Fan

After the initial set of gullies was located, raw images were processed in ISIS for map projection and georeferenced for ArcGIS. A base map, provided by USGS Astrogeology PDS Map-A-Planet from a mosaicked Viking dataset, was sinusoidally projected in ArcGIS. Initial images included 31 sets² of gullies. Seventy-three more sets of gullies were added later, as well as a Mars Orbital Laser Altimeter (MOLA) elevation map. Symbols

¹Structures are classified as gullies if (1) they posses a well defined channel and/or (2) they posses both a depositional fan and an alcove structure. Structures containing one of the three parts are not classified as a gullies.
²A set of gullies refers to the presence of a gully or gullies, regardless of number, in a single locality having the same orientation.
were added to gullies so that classification information was available in a graphical format.

3. Observations
3.1 Spatial Relationships
Most gullies in this area are on crater walls. Seventy-five of 114 sets were on crater walls, while 31 sets were on graben walls in Gorgonum Chaos, and the remainder were found on central peaks, rings, and other sites. In the northern portion of the study area (N of 46°S) most gullies are pole facing (i.e. south facing). In the southern portion (S of 46°S) gullies are mostly equator facing. In addition to this dependence of gully direction on latitude, we also find a dependence on longitude: in mid-latitudes gullies in the west are mostly equator facing, while those in the east are mostly pole facing (Figure 2).

Figure 2. Gullies in our study area from 154°W - 172°W and 30°S - 65°S. The white north facing arrows indicate equator facing gullies while the black south facing arrows indicate pole facing gullies. The ratio of pole facing to equator facing in this study is 68:32.

Malin and Edgett [2000] first noted the quantitative difference of pole facing vs. equator facing gullies: most are pole facing. Our dataset included 68 pole facing sets of gullies, 32 equator facing sets, 11 due east sets, and 3 due west sets; which is in the near the Malin and Edgett [2000] 71% (PF: EF) ratio. However, this ratio neglects to identify the terrain gullies form on. All 31 sets of gullies in Gorgonum Chaos were pole facing. This skews the results of this study, as without this locality the ratio is near 1:1 (PF: EF) (Edgett [2003]).

With MOLA elevation data, we compared gully properties to regional and local elevations and slopes. In Gorgonum Chaos and Newton Basin, gully sets faced both uphill and downhill. Gullies seem more abundant in basins than on high ground, however, this may be due to the heavy coverage of MOC imagery in basins compared to mountains.

We did not note meaningful or strong correlations among gully properties (except facing direction) in elevation, geologic setting, or environmental influences.

3.2 Age Relationships
Craters containing more than one gully commonly show evidence of multiple formation ages. One depositional fan in our study possessed craters; however, in several locations sand dunes lie on top of depositional fans. Depositional fans are often superimposed on other fans. This shows that gully formation was not a one-time event and possibly, an ongoing process. Gullies interpreted to be younger geologically are sharper and more clearly marked, while older gullies are smoother and occasionally have boulders on the surface of the crater wall, between channels or superimposed on a depositional fan.

4. Discussion
Malin and Edgett [2000] interpreted gullies as being from recent groundwater seepage, released from solar melting of ground ice on pole facing slopes. Our work identifies two problems with this model. In our study area, gullies are not preferentially on pole-facing slopes, but are more or less evenly distributed (Edgett [2003]). Similarly this model does not predict the latitude dependence of gully facings or its longitudinal dependence (Fig. 2). In addition it is difficult to understand how groundwater could form and persist on isolated highlands, like central peaks and peak rings (Treiman [2003, 2005]).

Geothermal heating and melting of near surface ice (Hartman [2001]) should produce even distribution of gully facing directions, however we found few west facing gullies. In addition, there are no signs of recent volcanoes or lava flows expected from near surface geothermal heating (Treiman [2003]).

Gaidos [2001] model also depends on geothermal heating for cryovolcanism (water deep beneath the surface moves surface-ward and freezes, then under great enough pressure creates a large outburst of water). Regional ground ice and aquifers aren’t required, but this hypothesis fails to explain the lack of telltale signs of cryovolcanism anywhere except steep walls. Gully formation theories involving water also fail to explain how alcoves form at different elevations, on the same crater wall. Water models also cannot to explain how
gullies can form repeatedly over short periods of time to show different formation ages.

Gully formation by melting of ground ice by solar insolation (Mellon and Phillips [2001]; Costard et al. [2002]; Heldmann [2003]) is unlikely for many of the same reasons. This model also requires a high obliquity angle, which is much greater than Mars’ current axial tilt. Equator facing gullies are prominent at all latitudes, even though pole-facing gullies outnumber equator-facing gullies. Our work implies that specific slope angles and directions are not required for gully formation. Therefore, solar insolation cannot account for the regional variation of gully locations.

Christensen [2003] suggests water-rich snow melts on crater walls are coated by a layer of insulating dust, to form gullies at high obliquities. This theory, like solar insolation, fails to explain why both pole facing and equator facing gullies form at the same latitudes, when Christensen [2003] acknowledges that the proper warming would only take place on pole facing slopes at high obliquities. The Christensen [2003] model also fails to explain how different ages of gullies occur as the period of one high obliquity to another would presumably be long enough to erode and/or cover up existing gullies. The differences in solar heating cannot explain the spatial distribution of gullies, while the premise of water rich snow at mid latitudes is not widely supported.

Liquid CO₂ (Musselwhite et al. [2001], Draper et al [2000], and Hoffman [2000]) must become pressurized in a similar manner to ice for it to explode out of the crater wall, so this falls into the same geologic pitfalls as ground ice theories.

Eolian processes (Treiman [2003]) are able to avoid many of the drawbacks of water and CO₂ models. Eolian sedimentation and deposition would allow gullies to form at different times and locations in a single crater due to variation in Martian winds. The Treiman [2003] model does not require anything like a regional aquifer, which perhaps allows terrain to play a significant role in gully formation. The Treiman [2003] model doesn’t explain the regional shift from pole facing gullies in the northern portion of the study (30°-46°S) to equator facing gullies (south of 46° S).

So, no hypothesis of gully formation that invokes liquid water can explain how gullies form without a host of anomalies and special circumstances. However, the eolian sedimentation and depositional processes that are dominant in the Treiman [2003] model seem like a possible mechanism for gully formation. However, the Treiman [2003] model needs to explain the regional variation in gully direction for it to be legitimately considered as plausible.

5. Acknowledgements
I would like to thank Allan Treiman for all of his help, knowledge, and support over the past ten weeks, Brian Fessler for always being able to solve our technical problems, and everyone at LPI for making me feel at home this summer.

6. References


Latitudinal/Seasonal Variation of Hydrocarbon Abundances on Uranus and Neptune
Renee Naphas1, Julie Moses2 and Thomas Greathouse2, 1Embry-Riddle Aeronautical University, 2Lunar and Planetary Institute

Introduction:

The photolytic destruction of methane in giant-planet atmospheres initiates a photochemical network of reactions which results in the production of C$_2$H$_2$ and C$_2$H$_6$. Currently, limited information exists concerning atmospheric composition and temperatures on Uranus and Neptune. Observations alone have not provided sufficient information because the low flux from these objects has required averaging across large areas. Using the Caltech/JPL Kinetics code [Allen et al. 1981], we have developed photochemical models that account for seasonal variations in solar insolation to make predictions of C$_2$H$_2$ and C$_2$H$_6$ abundances. These predictions are then compared to observational data to verify model accuracy and to identify future observational strategies.

C$_2$H$_2$ is of interest because it has a relatively short photochemical lifetime in gas giant stratospheres [e.g., Moses et al., 2005], and therefore is less affected by stratospheric dynamics than the long-lived C$_2$H$_6$. The comparison of C$_2$H$_2$ and C$_2$H$_6$ abundances with latitude will dictate the level of influence of solar flux versus meridional transport on hydrocarbon abundance variations with latitude and season.

Photochemical Model Inputs:

The models are run for two full Uranus and Neptune years (84 and 165 Earth years, respectively) — the same two planetary years are repeated until the results converge to make sure that high-altitude variations have had a chance to diffuse through the stratosphere. The 11-year cycle of the solar ultraviolet flux is incorporated in the model. The required solar flux covers a large time period, for which we do not have actual solar flux measurements. We use a normalized solar flux, which consists of the combination of the daily/monthly International Sunspot Number and actual/predicted F10.7-cm values. Ring shadowing is found to be negligible for both planets and is ignored. The hydrostatic equilibrium equation is solved to create an altitude-pressure-temperature-density grid for every 8 degrees of latitude. Heliocentric radii are acquired from the online JPL Ephemeris calculator at <http://ssd.jpl.nasa.gov/cgi-bin/eph>.

Observations:

Data for Neptune were obtained by Therese Encrenaz, using TEXES, the Texas Echelon cross-dispersed Echelle Spectrograph, mounted on the NASA Infrared Telescope facility on June 22nd and 23rd 2003. The spectrograph slit was oriented along celestial N/S, while Neptune’s rotational axis was tilted by 2.96°. Neptune had an equatorial diameter of 2.29 arcsec and was at L$_s$ = 251.58° (southern spring). The sub-Earth point was ~29.32° planetographic latitude. The spectral regions of interest included 1229.4-1234.5 [cm$^{-1}$] for the $\nu_4$ band of methane, 819-825.2 [cm$^{-1}$] for $\nu_9$ band of ethane, and 729-731 [cm$^{-1}$] for $\nu_5$ band of acetylene. A slit width of 1.4 arcsec was used for the CH$_4$ observations giving a resolving power of ~100,000, while a 2.1 arcsec wide slit was used for the C$_2$H$_2$ and C$_2$H$_6$ observations resulting in a resolving power of ~80,000.

Figure 1: An observation of Neptune’s southern hemisphere between –29° and –90° latitude. The data (crosses) are compared to our best-fit radiative-transfer model (solid line). The dashed line is the telluric (Earth) transmission scaled by 0.1.

Observational Results:

The observations are modeled using a line-by-line radiative transfer code assuming plane-parallel geometry and LTE throughout (Greathouse et al., 2005). Figure 1 is an example of model output compared to observational data. By assuming that the methane abundance profile from the globally averaged photochemical model is invariant with latitude, we are able to infer a temperature profile constrained by the observed CH$_4$ emission spectra. Using the derived temperatures, we then vary the C$_2$H$_2$ and C$_2$H$_6$...
abundances to fit the 730 cm$^{-1}$ and the 822 cm$^{-1}$ data, respectively.

**Photochemical Model Results:**

A contour plot of mean daily solar insolation, for both Uranus and Neptune, is shown in Figure 2. Note that the time period in which a particular location receives no solar flux during the day (e.g., polar night) extends to very low latitudes on Uranus as compared with Neptune because of Uranus’s high obliquity. In addition, yearly averaged solar flux values are greater at high latitudes than low latitudes on Uranus; the opposite situation occurs for Neptune. Therefore, one might expect the latitudinal/seasonal variation to be different on the two planets.

Figure 3 shows how the mole fractions of CH$_4$, C$_2$H$_6$, and C$_2$H$_2$ vary with season on Uranus and Neptune. Note that our 1-D photochemical model predicts that Neptune should exhibit marked seasonal variations, whereas Uranus does not. This difference is largely caused by the fact that atmospheric mixing (as represented by an eddy diffusion coefficient in 1-D models) is greater on Neptune than on Uranus.

When methane is photolyzed near the 0.1-mbar region on Uranus, the lifetimes of the products C$_2$H$_2$ and C$_2$H$_6$ are already approaching a Uranus year. These lifetimes only increase as altitude decreases; therefore, seasonal changes have no effect on ethane and acetylene abundances. On Neptune, methane is photolyzed in the ~1x10$^{-4}$ mbar, where the C$_2$H$_2$ and C$_2$H$_6$ photochemical lifetimes are very short. Although the photochemical lifetimes of these species increase with decreasing altitude, vertical diffusion time scales are short down to ~0.1 mbar, the C$_2$H$_2$ and C$_2$H$_6$ produced at higher altitudes diffuse down to ~0.1 mbar, where seasonal variations can still be seen. Below the 0.1 mbar level seasonal influence decreases as the pressures increases and vertical diffusion time scales for both hydrocarbons approach a Neptunian year.

**Conclusions:**

Radiative-transfer modeling of the observational data shows no latitudinal variation in temperature or C$_2$H$_2$ or C$_2$H$_6$ abundances for Neptune, within error. The photochemical models show a slight variation with latitude, but the photochemical models and observations agree to within observational uncertainty. If future observations of Uranus show latitudinal variation in these abundances, then either the eddy diffusion coefficient is changing with time or 2-D photochemical models are needed to explain observations.

Given the similarity in the photochemical lifetimes for C$_2$H$_2$ and C$_2$H$_6$, we would expect these molecules to have a similar latitudinal distribution in the stratospheres of Uranus and Neptune. Some differences in C$_2$H$_2$ and C$_2$H$_6$ lifetimes do exist in the 2-8 mbar region on Uranus and in the 0.01-0.1 mbar and 0.3-2 mbar regions on Neptune. If meridional transport time scales are greater than ~40 years in the 2-8 mbar region on Uranus, then C$_2$H$_2$ and C$_2$H$_6$ might exhibit different latitudinal variations in this pressure region. Similarly, if meridional time scales on Neptune are greater than 100 years in the 0.01-0.1 mbar region or greater than ~500 years in the 0.3-2 mbar region, then C$_2$H$_2$ and C$_2$H$_6$ might exhibit different latitudinal variations in these pressure regions.

No observations are currently available to help constrain meridional wind/diffusion speeds or transport time scales on Uranus and Neptune. The spreading of the Shoemaker-Levy 9 debris after the 1994 impact of the comet with Jupiter allowed meridional diffusion coefficients to be derived for the Jovian stratosphere (e.g., Friedson *et al.* 1999, Moreno *et al.* 2003). If meridional diffusion coefficients are similar on the other giant planets as on Jupiter, then we might expect meridional transport time scales on Uranus and Neptune to be ~2 years at 0.3 mbar and ~20 years at 30 mbar. These transport time scales are considerably smaller than the photochemical time constants for C$_2$H$_2$ and C$_2$H$_6$, suggesting that meridional transport might clear away any latitudinal variations caused by solar insolation changes in the lower stratospheres of Uranus and Neptune. By the same token, C$_2$H$_2$ and C$_2$H$_6$ should exhibit a similar latitudinal variation that is controlled by transport effects rather than by photochemistry.

**References:**

Sromovsky *et al.* 2001. Icarus 150, 244-260.
Figure 2: Daily mean solar insolation for Neptune (Left) and Uranus (Right) [W*m\(^{-2}\) per planetary day]. Shaded regions denote polar night.

Figure 3: Mole fraction profiles for -80° planetocentric latitude on Neptune (Left) and Uranus (Right). Southern summer (\(L_s = 270°\)) is represented by the dotted lines, winter (\(L_s = 90°\)) by the dashed lines, spring (\(L_s = 180°\)) by the solid lines, and fall (\(L_s = 0°\)) by the dot-dashed lines. All points indicate observational data.
Opaque Assemblages in CK and CV Carbonaceous Chondrites
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Introduction:
CK carbonaceous chondrites are named after the first observed fall in Karoonda, South Australia in 1930. The CKs are the only group of carbonaceous chondrites that exhibit thermal metamorphism. The CKs are largely defined by Antarctic and southern hemisphere finds. Thermal metamorphism is indicated by petrologic type numbers 4-6 increasing in degree of thermal alteration. As a result, CKs display features of metamorphism such as silicate darkening, recrystallization and shock veins. Silicate darkening is created by small grains of pentlandite and magnetite forming inside silicates. Calcium Aluminum Inclusions and Fe-Ni metal are rare. The most common chondrule texture is porphyritic consisting of 99% of chondrules in the CK group [6], but a few barred olivine chondrules also exist. Chondrule rims are reported to not be in CKs although olivine rims around BO were found in this study and opaque rims are common [7].

<table>
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Table 1: Modal Abundances. Data for CV and CK groups from [6] Scott and Krot.

The CV and CK carbonaceous chondrite groups have been compared to each other often because of petrographic similarities, such as overlapping oxygen isotopic ratios. Scientists have suggested the two groups of carbonaceous chondrites formed from the same parent body and CKs are equilibrated CV chondrites [1, 2]. The oxidized CV group has been most closely related to CKs.

This study examines the petrology and mineralogy of CKs and CVs focusing on opaque minerals found in the meteorites. Using the oxide, metal and sulfide assemblages, constraints can be placed on the temperature and oxygen fugacity at which the meteorites equilibrated. The temperature and oxygen fugacity of the CK and CV chondrites can be compared in order to help define their formation history.

Experimental Methods:
The polished CK thin sections used in the study include; MET01149,11, EET90007,11, EET99430,5, QUE99679,7, QUE99680,8, EET87860,15, LEW87009,13 from JSC and Maralinga. The polished CV3 thin sections include the Bali type, ALH85006,23, the Allende type; LEW86006,2, ALH84028,11 and ALHA81258,4 and the reduced type, QUE93429,5 from the JSC collection. The thin sections were studied on an optical microscope using transmitted and reflected light and the 5910LV JEOL Scanning Electron Microscope at JSC. The mineral composition data was obtained using the JSC Cameca Electron microprobe using natural and synthetic minerals as standards and PAP reduction scheme for ZAF corrections. The probe was operated with a beam current of 20 nA and an accelerating velocity of 20kV.

Modal abundances were determined using SEM back scattered electron mosaic images of sections and referencing to an optical microscope.
**Results:**

The CK thin sections in the study represent normal CK carbonaceous chondrites. The published data on modal abundances for the CK group (Table 1) show the matrix is 75% and the chondrules are 15% for the CK group [6]. The average matrix in this study of Antarctic CKs is 71.2% and the average amount of chondrules is 13.9%.

The opaque assemblages are found throughout the matrix, spherical inside chondrules and in rims around chondrules. The assemblages consist of sulfides, metal and magnetites.

![Figure 1: Barred olivine chondrule surrounded by rim of opaques from EET9007](image1.jpg)

Magnetite is the most abundant opaque mineral in CK chondrites. However, magnetite was not found in MET01149, petrologic type 3. Exsolution lamellae ilmenite between 5-10 µm and alumino spinel were found in the magnetite of the CK chondrites. Magnetite was less abundant in the CV sections and found in oxidized and reduced subgroups.

![Figure 2: The image is a magnetite grain with exsolution lamellae of ilmenite from EET99430.](image2.jpg)

Ilmenite was found as lamellae in CKs petrologic type 5-6. Ilmenite also existed as grains in LEW87009, petrologic type 6. Ilmenite was not found in the CV thin sections.

The sulfides in the CK chondrites primarily are Fe or Fe,Ni sulfides. One grain of ehrlichmanite, OsS₂, 2 µm in size was found in LEW87009. The most prevalent sulfide is pentlandite (Fe,Ni)₉S₈ and is often found in association with magnetite in CK chondrites. The Fe sulfides are pyrite, pyrrhotite (Fe(1-x)S) and troilite. Millerite, NiS was not found in this study, but has been reported in Maralinga by Geiger and Bischoff [5].

![Figure 3: Fe-Ni-S Diagram](image3.png)

Fe, Ni metal was common in the CV thin sections, but not found in the CK sections. Metal in volume percent of the CV group is 0-5% and <0.01% for the CK group [6].

**Discussion:**

Exsolution of ilmenite within magnetite has been found in CK thin sections. The oxygen fugacity and temperature for the magnetite and ilmenite pairs were calculated using the Fe, Ti thermometer/oxybarometer [3,4], based on the distribution of Ti between ilmenite and magnetite.

The CV thin sections in this study did not contain ilmenite, but analyses of the Fe, Ni metal can be used with magnetite compositions to calculate oxygen fugacity. The temperature and the oxygen fugacity for the CV thin sections were calculated using thermodynamic equilibria between metal and magnetite. The Gibbs free energy data was obtained from Robie et al [8].
3Fe + 2O₂ = Fe₃O₄
ΔG = -RTlnK
\[ fO_2 = \sqrt{\left(\frac{aFe}{aFe_3O_4}\right) \left(\frac{aFe}{e}\right)^{-\Delta G/RT}} \]

The temperature cannot be calculated using the same method used with the magnetite-ilmenite pairs. Therefore, the oxygen fugacity was calculated using a temperature range of 573-1173 Kelvin (300-900 °C) for each metal composition. The results shown in the graph of oxygen fugacity vs. temperature (Figure 5) show all types of CVs are more reduced than the CK chondrites. The CK temperature and oxygen fugacities plot in a linear trend.

Figure 4: Oxygen Fugacity vs. Temperature

The calculated temperature range from Figure 4 is 550-926 Kelvin (277-653 °C). Geiger and Bischoff have estimated the temperatures for CK to be in the range of 823-1273 Kelvin (550-1000 °C) [9]. Pentlandite, a mineral found in the highest petrologic types of CK chondrites, is not stable at temperatures greater than 883 K (610°C) [10]. This falls within the calculated temperature range from this study.

Petrologic evidence for an oxidized formation of CK chondrites includes the absence of Fe, Ni metal and the abundance of magnetite [5]. The CV chondrites contain Fe,Ni metal in the reduced and oxidized subgroups. Even the oxidized CV’s are less oxidized than the CK chondrites since they contain metal.

The largest difference between the CK and CV chondrite groups is the amount of oxidation. The average ΔFMQ for a CV in this study is -2.36 while the average ΔFMQ for CK is 3.70. Metamorphism is an explanation for the high degree of oxidation of CKs, but the question if a CV would become as oxidized as a CK chondrite if a CV went through the same metamorphism remains.

Acknowledgements:
I want to thank Kevin Righter for providing me with a fascinating project and working with me, but most of all for being a friend and a mentor. I would like to thank Mike Zolensky and Lindsay Keller for their time, expertise and for extreme generosity with loaning me thin sections and resources, which helped shape my project. I would like to give a special thanks to Loan Le for many probe calibrations and LPI and JSC for this incredible opportunity.

References:
Mineralogy and Mineral Chemistry of Primitive Interplanetary Dust

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1. Introduction

Interplanetary dust particles (IDPs) are collected in the stratosphere at 20-25 km altitude and are typically 5-15 µm in diameter. There are two major types of IDPs, chondritic porous (CP) IDPs and chondritic smooth (CS) IDPs. IDPs have bulk compositions that are chondritic within a factor of 2 for most major elements. Most chondritic porous IDPs are anhydrous, while the chondritic smooth IDPs are generally hydrous. CP IDPs are believed to have a direct link to cometary sources [1]. They have escaped the severe thermal metamorphism and aqueous alteration that has influenced even the most primitive meteorites. This is demonstrated by their high-carbon and nitrogen abundances, unequilibrated mineralogy [2], presence of presolar hydrogen and nitrogen isotopic signatures and abundant presolar silicates [3]. Indeed, CP IDPs are the most primitive materials that are available for laboratory studies. Detailed studies of CP IDPs may reveal new insights into the nature of the early solar system.

Trace and minor element abundances in IDPs are different from those of primitive meteorites. For example, Klöck et al. (1989) reported the presence of low-iron, manganese-enriched (LIME) forsterite in IDPs and concluded it was a condensate from the early solar nebula [4]. Weisberg et al. (2004) also reported such forsterite in primitive meteorites [5]. In this study we systematically investigated the compositions of both olivine and pyroxene grains in several CP IDPs with emphasis on their minor chemistry and comparisons with similar phases in some primitive meteorites.

2. Samples and methods

We analyzed microtome thin sections of individual IDPs using a JEOL JEM-200FX transmission electron microscope (TEM) equipped with a thin-window energy dispersive x-ray spectrometer. EDX spectra were collected until the Si peak reached 5K counts full scale which allowed for counting statistics of <1% for major elements. Counting statistics for the minor elements were typically 10-50% depending on concentration.

3. Results

We analyzed thin sections of 9 different interplanetary dust particles and collected data on 143 pyroxene grains and 98 olivine grains. Figure 1 shows FeO vs. MnO abundances (wt. %) of pyroxene grains in IDPs and some primitive classes of meteorites. Most of the meteorite data show little increase...
of Mn with increasing Fe concentration for both olivine and pyroxene. However, many of the IDPs show a different trend. They have much higher Mn content than typical meteorites and tend to be much more Fe-poor. Some show Mn contents that are even higher than that of Fe. It is very unusual characteristic of IDPs and similar to the trend identified by Klöck [4]. Figure 2 shows FeO vs. MnO abundances (wt %) of olivine grains in IDPs and some meteorites. As seen in pyroxene grains, some olivine grains in IDPs also have higher content of MnO than FeO.

Table 1 and 2 shows the typical compositions of pyroxenes and olivines in interplanetary dust particles. Pyroxenes and olivines in IDPs typically contain minor amount of Mn. In this study the highest MnO content in pyroxenes and olivines are 3.84 and 1.51 wt. % respectively. Most of the “High-Mn” data shows higher Mn content than Fe.

We have discovered unusual enstatite compositions in several of the IDPs that have very high contents of Al, but no detectable Na. In Figure 3, which shows the content of Al2O3 vs. CaO abundance (wt %), a lot of IDPs which have low-Ca, high-Al content were observed. The highest Al2O3 content is 9.10 wt. %, while CaO content is 0.67 wt. %. Only some of the high-Ca pyroxenes in carbonaceous chondrites show such high-Al contents. Another characteristic of Al in pyroxenes in IDPs observed in this study is the crystal chemical site of Al. Typically, Al resides in the tetrahedral sites or as the coupled substitution ViAlIVAl for VIM2+IVSi4+ (Tschermak’s substitution.) However, for many of the IDP pyroxene grains, much, if not all, of the Al resides in the M2+ cation site with Mg, Fe, Cr, Mn and Ca.

4. Discussion

In this study, we analyzed 143 pyroxene grains and 98 olivine grains in 10 thin sections. Most of these anhydrous IDPs are pyroxene-dominated (En>>Ol), except for only one IDP thin section which contain no pyroxene grain (W7013D1). IDPs are typically far more pyroxene-rich than any of the primitive chondritic meteorites. The high pyroxene-olivine ratio we observed in most of the IDPs is consistent with astronomical observations of the silicate mineralogy of solar system comets [e.g. 17]. As shown in Figures 1 and 2, there are obvious difference between the Mn content of pyroxenes and olivines in IDPs and meteorites. These differences may reflect from different histories.

The origin of the highly aluminous enstatite is poorly known. Mg-Tschermak’s enstatite is a high-pressure phase observed in experiments on upper mantle rocks [e.g. 18]. One possibility is that the high temperature alteration of enstatite may produce aluminous enstatite with “excess SiO2,” or Mg/Si <1, shown substitution by a hypothetical Mg Eskola pyroxene, Mg0.5Al[Si2O6], with vacancies in the large (M2) cation sites [19].

References

Most of these data show higher Mn content than Fe.

Table 2. Typical compositions of olivine grains in IDPs. “High-Mn” shows the data that shows the highest Mn content in each thin section. Most of these data show higher Mn content than Fe.
LAYERS, BRIGHT-RIM CRATERS, AND FAULTING IN THE ARABIA TERRA, MARS. E. M. Venechuk¹, C. C. Allen², and D.Z. Oehler³, ¹Scripps College, Pomona College Dept. of Geology, Claremont, CA 91711, elizabeth.venechuk@gmail.com, ²NASA Johnson Space Center, Houston, TX 77058

Introduction: The atmosphere above Arabia Terra has comparatively high methane concentrations [1]. Methane has a lifetime of ~300 years in the Martian atmosphere [2], suggesting an active source in Arabia Terra. Possible sources include volcanic outgassing, bacterial methanogenesis, hot springs, and release of thermally matured organic deposits [3]. Arabia is also a region noted for its layered outcrops, suggestive of sedimentary rocks[4-6]. By identifying where in the region layers occur, past areas where water may have been present can be identified. Bright rings around craters, observed in Meridiani Planum, resemble terrestrial red bed bleaching due to hydrocarbons (Fig. 6) [3]. These bright rings may indicate where hydrocarbons underlay the region. Numerous small faults run through Arabia, providing potential conduits for hydrocarbons to be released to the surface (Fig. 2). Mapping where layers, bright-ring craters, and faults occur and intersect in Arabia, a process analogous to “New Ventures” projects in the oil and gas industry, may indicate possible areas from which the methane is being sourced.

Arabia Terra is characterized by heavily cratered Noachian plains, as well as a steady slope from -4000 m with regard to the Mars Orbital Laser Altimeter (MOLA) datum in the NW to 4000 m in the SE. This slope may have provided a constraint on sediment deposition and thus layer formation. Most of the region is Noachian in age; however, a significant percentage of the region’s area is identified as Hesperian. Although the history of the Arabia Terra initially seems to be straightforward – cratered plains with several younger units atop them – detailed mapping may reveal a more complex history.

Methods: 1011 high-resolution (5-12 m/pixel) Mars Orbital Camera (MOC) pictures from 0 to 40 N and 20 W to 60 E (spanning all of Arabia) were examined for the presence of layers, bright rings and faults. Three image sets, differentiated by date, were used: AB1-M04, R03-R09, and R10-R15 [7]. These three sets covered several years and all seasons on Mars, so any seasonal variation in visibility was accounted for. A map was generated from the data indicating where sediments and other features were located in Arabia and analyzed for trends. Sites where layers, bright-ring craters, and faults overlapped were identified, and the occurrence of layers was compared to regional MOLA data and Viking-era geologic maps.

Results: 80.2% (811) of the MOC images examined showed layers, which were present in varying concentrations across the entire region. No trend associated with latitude or longitude was seen; however, certain trends can be identified for each half of the map. The East and West halves of the map do not show matching trends, especially regarding altitude.

In the West half, certain areas show a noticeable lack of layered outcrops while other areas have numerous examples of layers (Fig. 1). When compared with the Viking-era geologic maps, those areas with very few layers are often in the same areas identified as Ridged Units (Nplr and Hr), mapped as volcanic in origin [8,9]. There is also a correlation between a lack of layers and the -2000 m altitude line in the western half of the map (region A, fig. 1). Layers occur above the -2000 m line and below the -2500 m line. Both of these altitude changes are relatively steep. In the East half, layered deposits are fairly evenly distributed, with a few exceptions: ridged units tend to have fewer layers, and altitudes above 1000 m tend to have fewer layers. Although there are regions between -2000 m and -2500 m in the East half of the map, no corresponding lack of sediments has been identified.

Bright rings surrounding craters are confined between 0 and 20 W, and faulting is concentrated in the SW corner of the region (Figs. 2, 6). In the area defined by 2 to 12 N and 5 to 12 W (Fig. 1), layers, faults, and bright rings all occur, suggesting an area from which methane may be leaking. Further methane investigations in this small region are recommended.

Three distinct types of layers have been identified. Type one consists of alternating layers of bright and dark rock (Fig. 3). Type two consists of layers that cannot be visually distinguished from each other, but still erode at regular intervals (Fig. 4). Type three consists of thin, usually dark layers that do not form an even cliff face, and may have shed boulders that can be seen downslope (Fig. 5).

Discussion: The lack of layers between -2000 m and -2500 m in the West half of the region suggests an altitude-dependent erosion or concealment of the layers seen in the surrounding terrain. Both contour lines mark areas of very rapid change: a drop of at least 150 m and up to 300 m in topography (Fig. 1). The -2500 m altitude lines marks the boundary of the northern lowlands. The few instances of layers in region A are either exposed in crater walls or are within the craters themselves. Several possible scenarios in which region A developed are possible.
In the first scenario, sediments were deposited over the entire region above -2000 m. As the north-south dichotomy formed its sharp cliff at what is now -2500 m, sediments were deposited below -2500 m. The rare layers seen in region A are windblown deposits atop an older surface. In the second scenario, sediments were deposited over the entire region regardless of altitude, after which a depositional event exclusively in region A, such as a lava flow or massive dust layer, or burying the layers. The layers seen in crater walls reflect this second event. The formation of the north-south dichotomy cut away at the deposit covering the sediments underneath region A and exposed them below -2500 m. In yet a third scenario, the entire region was covered with layers, which were then eroded off of region A, possibly creating the sharp topographic boundaries [10]. The layers seen are remnants of the original layers. Many more possible scenarios undoubtedly exist; these are presented to suggest some of the more likely mechanisms by which the lack of sediments in region A relative to the rest of Arabia may be explained.

The three types of layers may be distinguished by composition. The first type, noted for its color contrasts, may represent a varying depositional environment such as a terrestrial near-shore environment as sea level rises and falls, depositing alternating sandstone and limestone (Fig. 3). Another possibility involves the sediment supply. Regular volcanic events may provide large amounts of ash that interrupt the usual deposition of sand and dust, creating contrasting layered deposits. The second type, which often forms distinctive ‘stair-step’ mesas, could represent an environment that is not constant, with unconformities between layers causing the ‘stair-step’ erosional pattern to emerge (Fig. 4). The third type, a thin rubbly layer, could form when a sheet of lava flows across a sandy region (Fig. 5). The crystalline nature of the lava would make it more resistant to erosion than the particulate material surrounding it, allowing large coherent pieces to fall off and roll downslope. Identifying the distribution of these three types of layers may provide important information regarding the history of the Arabia Terra.
Future Work: The interactions between the Noachian and Hesperian units of Arabia and the distribution of layers suggests a complex history for the region. Studying the morphology and distribution of the three layers types may lead to important insights regarding the formation of the observed layers, and therefore the history of Arabia Terra. Polygonal cracks, potentially indicating permafrost [11], and possible pseudocraters, potentially indicating lava flows [12], have been observed in the region. Mapping these occurrences may provide other clues to the geologic history of Arabia. The last comprehensive geologic mapping used Viking data [8,9]. However, many higher-resolution datasets are now available. Performing a comprehensive geologic survey of the region using data from MOLA, MOC, and other instruments would be highly beneficial to this study.

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