Papers presented at the
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Summer Intern Conference

August 17, 1989
Houston, Texas

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**PROGRAM OF PAPERS PRESENTED AT THE FIFTH ANNUAL SUMMER INTERN CONFERENCE**

Lunar and Planetary Institute  
August 17, 1989  
Houston, Texas

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9:00 A.M. - 11:00 A.M.  
Chairman: Dr. Nadine Barlow

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INTRODUCTION

Various geochemical studies indicate that there are significant differences between Antarctic and non-Antarctic meteorites. It has been suggested that these meteorites are derived from different populations [1]. The purpose of this study was to determine whether some of these differences might be due to terrestrial weathering.

Carbonates in eleven carbonaceous and three ordinary chondrites were analyzed mass spectrometrically for carbon and oxygen stable isotope abundances. All except the Murchison and the Allende were Antarctic finds. The meteoritic types are listed in the Table-1.

Isotopic variations in carbonates can arise from mass fractionation during chemical reactions and diffusion. Hence they can provide information on the primary effects of the formation of parent bodies and on secondary hydrous alterations of the parent body or the meteorite itself.

From the data it will be inferred whether the carbonates are endogeneous to the meteorites or terrestrial weathering products.

EXPERIMENTAL PROCEDURE

After recovery the Antarctic finds were stored in a dry nitrogen atmosphere at JSC. The samples obtained for analysis were bulk meteorite powders ground to < 20 µm grain-size. Murchison was obtained from the Field Museum and has a grain-size of < 100 mesh. The Allende samples were obtained from the Smithsonian Institute (position 9 split 13) and Rice University (band saw fines). In the laboratory, all samples were stored in a desiccator.

The samples were digested with 100% H3PO4 in an evacuated reaction vessel (~10⁻⁵ torr) maintained at 25°C with a water bath (2). The evolved CO2 was separated from water and noncondensible gases using cryogenic methods. The acid also released H2S which could not be separated physically and interfered with the analyses. H2S was separated chemically using Pb(O2CCH3)2 (3, 4). The yields of CO2 were determined with a capacitance manometer to ±0.05 torr. After this first extraction, the vessels were then placed in a 50°C water bath, reextracted, and extracted in the same manner.

The isotopic composition of CO2 was analyzed on a Finnigan MAT dynamic mass spectrometer (model 251) using NASA 1 (δ¹³C(PDB) = -7.85‰, δ¹⁸O(SMOW) = 19.17‰) as a working standard. The analytical uncertainty is ± 0.1‰, however, due to sample size and pressure adjustments, the uncertainties varied and are indicated in figure-1 by error bars. The fractionation of the oxygen isotopes that results from the H₂CO₃ = H₂O + CO₂ equilibria was corrected according to Craig (5).

RESULTS

The results of the isotopic analyses are listed in the Table-1 along with the respective yields; also included are the total carbon abundances as measured by Gibson (unpublished results). A comparison of carbonate content of the meteorites to that of the total carbon content is summarized in Figure-3. In order to check the experiments, Murchison and Allende were analyzed as references and the results are compatible with (4).

It is interesting to note the clustering that occurs among the Allan Hills finds at the 25°C determination. The values for δ¹³C(PDB) and δ¹⁸O(SMOW) range from 40.7 to 45.8‰ and 23.1 to 24.7‰, respectively for the carbonaceous chondrites, and for ordinary chondrites the ranges are 2.5 to 3.0‰ and 21.7 to 22.5‰, respectively. The same trend is observed for the Yamato finds, with the two meteorite delta values being within a few per mil of each other for both carbon and oxygen isotopes. The type CO3 chondrite, ALH83108, yielded values representative of weathered chondrites (4). The Japanese find, B7904, also appears to have a similar Antarctic weathering history.

As for the 50°C reactions, the Allan Hills ordinary and carbonaceous chondrite values are not that different from the values obtained from the 25°C reaction. The same is true for the Japanese finds, except for δ¹³C(PDB) of Y793321,8, which is about 13‰ smaller. The interesting changes are those of the Allende (Rice) and the...
Murchison. The oxygen-18 in enriched by about 60% for the Murchison and about 100% for the Allende. However, the 13C abundances have not changed drastically.

Any small discrepancies between repeated analyses are probably due to the inhomogeneity of the bulk samples.

DISCUSSION

The yields of CO₂ obtained are in accordance with the type of carbonaceous chondrite; that is, the type C2 chondrites have a higher concentration of CO₂ than the type C3. In general, the meteorites with greater total carbon abundances released more CO₂ upon acid treatment. The low yields of the ordinary chondrites along with the isotopic composition, indicate that the carbonates are Antarctic weathering products (4). Also, ordinary chondrites do not in general contain carbonate minerals (6,7).

The differences between the delta values at the two temperatures cannot be solely attributed to a kinetic isotope fractionation due to the procedure, because the values for some analyses did not change; moreover, some yields were larger for the 50°C reactions. This would imply there are carbonate minerals different than calcite that require a higher temperature in order for acid dissolution, for instance, ferroan magnesite, dolomite, ankerite, and kutnohorite (8). However, the fine grain-size of the samples might argue against this (9). As a control, NASA 1 was treated in the same manner as the samples; the analysis revealed that the CO₂ isotope values were the same at both temperatures (within experimental uncertainty). As for the anomalously enriched δ¹⁸O of the Allende (Rice), this is more than likely contamination of the band saw fines? Though the Murchison yielded similar results, it was depleted in δ¹³C at 50°C and released equal amounts of CO₂ at both temperatures; and thus, cannot be attributed to fractionation alone.

The similar oxygen isotopic composition of the Allan Hills carbonaceous and ordinary chondrites suggest that the carbonates of the carbonaceous chondrites equilibrated with the Antarctic environment. The differences in the carbon isotopic compositions imply that the carbon isotope of the Allan Hills type C2 chondrites are endogeneous to the meteorites. Furthermore, δ¹³C(PDB) are typically 30 to 70‰ for endogeneous carbonates of carbonaceous chondrites (10). This is the case for all of the Antarctic carbonaceous chondrites except for EET83226, ALH83106, and B7904. The ALH83106 is a type III, as is the Allende, and has a similar isotopic composition; B7904 has a typical Antarctic weathering history with equilibration of ¹³C. Of the Antarctic meteorites only the Yamato finds have oxygen compositions different from the typical value of ~20‰. Combined with data on terrestrial ages (11), it can be presently concluded that the carbonates of the Antarctic carbonaceous meteorites are endogeneous, enriched in heavier isotopes, and are equilibrating with the environment.

REFERENCES

### Table 1: Isotopic Composition of Carbonates in Meteorites

<table>
<thead>
<tr>
<th>METEORITE CLASS</th>
<th>MASS</th>
<th>25°C REACTION</th>
<th>50°C REACTION</th>
<th>BULK TOTAL</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>C-13</td>
<td>O-18 Yield CO₂</td>
<td>C-13</td>
<td>O-18 Yield CO₂</td>
</tr>
<tr>
<td></td>
<td>mg</td>
<td>(%) PDB</td>
<td>(%) SMOW</td>
<td>μmol/mg</td>
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<tr>
<td>Murchison</td>
<td>CM2</td>
<td>102.16</td>
<td>33.3</td>
<td>35.2</td>
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<tr>
<td>Allende (Smith)</td>
<td>CV3</td>
<td>113.31</td>
<td>-4.7</td>
<td>20.9</td>
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<tr>
<td>Allende (Rice)</td>
<td>CV3</td>
<td>101.34</td>
<td>-3.1</td>
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<tr>
<td></td>
<td>CM2</td>
<td>100.95</td>
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<td>31.1</td>
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<td></td>
<td>CM2</td>
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<td></td>
<td></td>
<td>102.85</td>
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<td></td>
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<td></td>
<td></td>
<td>78.09</td>
<td>12.9</td>
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<tr>
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<td>B7904,9</td>
<td>CM2</td>
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<td></td>
<td>CM2</td>
<td>77.9</td>
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<td>Y79321,8</td>
<td>CM2</td>
<td>88.85</td>
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<td>Y791824,7</td>
<td>CM2</td>
<td>80.33</td>
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<td>30.2</td>
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<tr>
<td>AL 77214,28</td>
<td>LL</td>
<td>102.13</td>
<td>2.5</td>
<td>22.5</td>
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<tr>
<td>AL 77230,16</td>
<td>L4</td>
<td>103.05</td>
<td>2.5</td>
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<tr>
<td>AL 77299,28</td>
<td>H3</td>
<td>100.19</td>
<td>3.0</td>
<td>21.7</td>
</tr>
</tbody>
</table>

**Figures:**
- Figure 1: Reaction at 25 degrees
- Figure 2: Reaction at 50 degrees
- Figure 3: Carbon content
GRAVITY MODELING OF THE MANSON STRUCTURE: IMPLICATIONS FOR AN IMPACT ORIGIN; R. A. Brackett, Dept. of Geol. Sciences, University of Texas, Austin, TX 78702, and V. L. Sharpton, Lunar and Planetary Institute, 3303 NASA Rd One, Houston, TX 77058-4399.

Introduction. The Manson Structure, located in North Central Iowa (Figure 1), consists of an ~7 km wide uplifted core of crystalline Proterozoic rocks surrounded by an ~35 km circular zone of disrupted and displaced Paleozoic and Mesozoic strata. The structure is completely buried ~30 m of Pleistocene glacial till, leaving only well logs, drill cores and geophysical methods as suitable techniques for investigation (Hartung and Anderson, 1988). Evidence of shock metamorphism and structural uplift led R. Dietz in 1959 to conclude that the Manson Structure represented a buried complex impact crater. Attention towards Manson was boosted recently by reports that the structure could be associated with the mass extinctions at the Cretaceous-Tertiary boundary (Kunk, et al. 1989, Hartung and Anderson, 1988). Subsequently, the impact interpretation of Manson has been challenged by Officer and Drake (1989) who contend that it hinges solely upon evidence of shock metamorphism and could therefore be endogenically formed. In addition, they claim that geophysical and morphological observations at Manson are inconsistent with an impact origin. Gravity data collected over the structure (Holtzmann, 1970) provide an important source of subsurface information. Holtzmann's study (1970) relied upon residual anomaly maps to provide a generalized structural picture. We have constructed inverse gravity models to enhance constraints on Manson's subsurface configuration and to evaluate the feasibility of the impact interpretation of this structure.

Methodology. The gravity data consist of 725 field stations covering 1600 km² extending 5 km beyond the boundary of the structure (Figure 2). Stations were spaced at roughly 2 km intervals. Several data points were apparently entered into the original data base incorrectly and these spurious points were discarded. Profiles were extracted from the Bouguer data and 2.5 dimensional density models were constructed using standard inverse methods. Modeling was achieved via MAGRAV2, an interactive magnetic and gravity modeling program (Broome, 1986, 1988).

The input parameters to our initial model included density data and general morphological constraints from geological and geophysical observations (Table 1). Density measurements for lithologies involved in the Manson structure were compiled by Holtzmann (1970). Location and width (7 km) of the crystalline core is constrained by well logs and aeromagnetic data (Henderson and Vargo, 1971). Hartung and Anderson estimated the structural relief on the core to be ~6 km based on the depth to crystalline basement outside the structure. However, due to the regional structural complexity (Figure 1) and paucity of deep measurements in the vicinity of Manson, this estimate is subject to substantial uncertainty. The extent of the deformed zone surrounding the core is constrained by well logs and refraction seismic models which indicate a 35 km wide zone of lowered seismic velocity (Smith, 1971). The depth of this zone was assumed to be equivalent to the depth to crystalline basement for the initial model. Half-strike widths were chosen to be consistent with these constraints. All other morphological characteristics of the model density distribution were determined by an iterative process of matching model (g_m) and observed (g_o) gravity values.

Results. Figure 4 illustrates the model density distribution along profile A-A' (Figure 2). Maximum deviation (g_m - g_o) in this model was less than 1 mGal. The floor of the structure is constrained by the gravity to slope gently toward the crystalline core, with the eastern flank apparently deeper and less steep than the western. Although in the initial model, 6 km of structural relief on the core was assumed, a better fit to g_o was achieved with a value of 4.5 km. In addition, we discovered that the core could not be modeled as a single body of uniform density; instead a lower density upper zone appears to be required. This is modeled in Figure 4 as a separate body. This is consistent with a highly brecciated plug of crystalline material with higher densities at the base resulting from hydrostatic closure of fractures.

The data along A-A' are consistent with the addition of two small (~ 6 km along strike, ~ 2 km across strike, ~ 300 m thick) bodies of relatively high density, just below the glacial till. These bodies are located approximately half way from the center of the structure and correspond to features apparent in aeromagnetic data (Smith, 1971).

The Manson Structure and the Impact Model. The morphological features described above all support a meteorite impact origin (Figure 3). The zone of low density is consistent with the intense brecciation characteristic of complex impact structures (Pohl et al., 1977, 1988, Sweeney, 1978). The dimensions of the crystalline core are compatible with those typical of the central uplifts associated with large complex structures. Morphological analysis of other terrestrial impact structures suggests that for a Manson-size crater (D = ~ 35 km) the central peak diameter should be 20 - 25 % D (Pike, 1985). The crystalline core diameter at Manson
Furthermore, the revised value of structural uplift (SU = 4.5 km or 13 % D) resulting from our analysis compares favorably with relationships derived from other terrestrial structures (SU ~ 10 %; Grieve et al., 1980). Occurrence of the shallow dense bodies shown in Figure 4 suggests that remnants of an impact melt sheet (Stoffler, 1981) may have survived erosion prior to burial. Such discontinuous deposits of melt are observed at the Mistassin Lake Structure, Quebec, a 22 km complex crater (Grieve, 1975, Marchand and Crockett, 1977). The asymmetry in the form of the structure could be indicative of oblique impact; however, structural and lithological complexities to the east or differential erosion prior to burial are plausible alternatives. Furthermore, there is no evidence for deep crustal involvement; the gravity characteristics of the Manson Structure are consistent with the shallow, rootless form characteristic of impact craters. In difference with the claim of Officer and Drake (1989), we find no morphological or structural evidence inconsistent with impact, nor any favoring endogenic origin. These observations, coupled with evidence of shock metamorphism, categorically support the impact interpretation of the Manson Structure.


Figure 1. Schematic Structural Cross-section across the State of Iowa (Hartung and Anderson, 1988).
Table 1. List of Regional Densities for Manson

<table>
<thead>
<tr>
<th>Rock Type</th>
<th>Density (g/cm³)</th>
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<tbody>
<tr>
<td>average</td>
<td>2.6</td>
</tr>
<tr>
<td>glacial drift</td>
<td>2.2</td>
</tr>
<tr>
<td>Mesozoic shale</td>
<td>2.8</td>
</tr>
<tr>
<td>Mesozoic limestone</td>
<td>2.6</td>
</tr>
<tr>
<td>Mesozoic sandstone</td>
<td>2.5</td>
</tr>
<tr>
<td>Paleozoic shale</td>
<td>2.5</td>
</tr>
<tr>
<td>Paleozoic limestone</td>
<td>2.8</td>
</tr>
<tr>
<td>Paleozoic sandstone</td>
<td>2.5</td>
</tr>
<tr>
<td>Precambrian rock</td>
<td>2.7</td>
</tr>
</tbody>
</table>

Figure 2. Location of Gravity Stations.

Figure 3. Bouguer Data (top) and Density Model of the Manson Structure (units are g/cm³).

Figure 4. Interpretive Cross-section of the Manson Structure.
GRAVITY, TOPOGRAPHY, AND TRANSIENT MANTLE PLUMES ON VENUS
Mark A. Fischer, Department of Astrophysical Sciences, Princeton University
Bruce G. Bills, Lunar and Planetary Institute

Introduction: A planet's gravity and topography are unique in that they can be remotely sensed, yet tell us something about the internal processes of the body. In recent years, the mechanism behind the production of topographic features on Venus has been the subject of much scientific debate. Crumpler and Head (1988) have suggested that Aphrodite Terra, the largest of the highland regions on Venus, is a divergent plate margin similar to mid-ocean ridges on Earth. Hot spot swells have also been invoked to explain some Venusian topography (Morgan and Phillips, 1983). We propose that analysis of the relationship between gravity and topography on Venus could help resolve this debate and provide important insight into the primary mechanisms of internal heat transport on Venus.

Procedure: Topography and gravity data from Pioneer Venus Orbiter (PVO) in the region from 45° S to 45° N were studied to determine lateral variations in effective Airy compensation depth. The topography was measured using radar altimetry. Line-of-sight gravity observations were acquired by measuring the doppler shift in the data communications link between PVO and Earth. After subtracting contributions from the relative motions of Earth and Venus, the Keplerian motion of PVO around Venus, the Earth's rotation, and solar effects, the residual doppler shift reflects variations in the gravity field of Venus.

The Airy hypothesis assumes that a crust of uniform density but variable thickness is in isostatic equilibrium on top of the mantle, so that mountains are compensated from beneath by "roots", analogous to icebergs floating in the ocean. We have used a local Airy compensation model, in which the effective depth of compensation is allowed to undergo lateral variations, not because we expect that Venus actually behaves in this manner, but because it provides a simple and intuitively appealing parameterization of the relationship between gravity and topography. LOS acceleration profiles were thus modeled using Venus topography, averaged into two-by-two degree bins and initially compensated at a uniform depth of 150 km. A constant mass density of the crustal material (2.9 grams per cubic cm) was assumed. The observed and model accelerations were then differenced to obtain a set of residual LOS gravity anomalies, reflecting lateral deviations from the uniform depth. In order to make the problem linear, we chose to represent these deviations by a set of infinitesimal or "point" dipoles at 150 km depth, one in each bin. The dipole moment was then taken to be the product of the surface mass in that bin (scaled down to reflect changes in bin area with radius) times a smoothly varying parameter with the units of distance. Variations in the magnitude of this parameter were represented by two dimensional quadratic b-splines centered on ten degree intervals and evaluated at bin centers. Figure 1 shows a contour map of the equivalent b-spline representation of Venus topography, and demonstrates their effect of smoothing out the changes in elevation without losing the important topographic trends and features.

It is convenient to think of the variable distance parameter as a correction to the effective depth of compensation in each bin. Strictly speaking, this designation holds true only when the parameter's magnitude is small compared to 150 km, since we have assumed an infinitesimal dipole. Nevertheless, the value of the parameter for a certain bin gives at least a qualitative indication of how deeply the topography in that bin must be compensated in order to reproduce the gravity observations as well as possible. We shall therefore refer to the changes in this parameter from bin to bin as compensation depth excursions about 150 km.

Results: A contour map of the depth excursions is displayed in Figure 2. The solution shown here was obtained by a least-squares fit to the residual accelerations with loose a priori constraints applied to suppress large deviations in depth at high latitudes. These anomalies were a result of the high altitude of PVO, and consequent low signal to noise ratio, at these latitudes. The constrained dipole moments accounted for over 57% of the variance of the residual gravity anomalies. Moreover, the entire model (compensated topography plus the point dipoles) reproduced almost 93% of the original data variance. Our region of interest contains four main positive topographic features (no large negative features exist on Venus). They are, from left
to right in figure 1: Ovda, Thetis, Atla, and Beta Regiones. These four features can be divided into two separate categories. Atla and Beta are regions of high-standing topography underlaid by positive depth excursions, and are therefore rather deeply compensated. In contrast, Ovda and Thetis are positive topographic features compensated at quite shallow depths, but bounded by areas of deep compensation. This latter category was quite unexpected and proved to be especially interesting.

Discussion: An interpretation of Atla and Beta as surface expressions of upwelling mantle plumes has been put forth on several occasions (Kiefer et al., 1986; Kiefer and Hager, 1988). The results of our analysis also support such a model. Under the Airy hypothesis, areas of deep compensation indicate a thick crust, which is equivalent to low-density material located at large depth. This situation is just that of a rising mantle plume. The compensation depth patterns found for Ovda and Thetis, on the other hand, are somewhat more enigmatic. These patterns could possibly be explained by low-viscosity, transient plumes. As noted by Olson and Nam (1986), the topographic and gravitational signatures of such plumes "are determined by the depth and the shape of the leading diapir, not by the conduit." As the spherically-shaped diapir approached the crust, it would undergo lateral spreading and take on a more toroidal appearance. As a result, the surface would be initially uplifted but would broaden in response to the spreading diapir, evolving from a high swell to more of a plateau. Following this interpretation, Atla and Beta are the surface expressions of young, active plumes, whereas Ovda and Thetis are more mature and developed features for which the compensating plumes have spread out into isolated patches around the uplifts. This scenario, though largely speculative, serves to explain a number of facts about these four regions. First of all, Ovda and Thetis are much broader and flatter in appearance than Atla or Beta. In fact, the central region of Thetis is slightly depressed, analogous to the subsiding of a swell crest produced by lateral plume spreading (Olson and Nam, 1986). Furthermore, the location of Ovda and Thetis at the equator is probably no coincidence. Prior to the formation of Atla and Beta, these two positive topographic features would have defined the equator, since a planet prefers to rotate about the axis of greatest inertia. Atla may be older than Beta, since the depth contours under Atla are more elongated. With the emergence of Atla, located at the present axis of least inertia, the equator would have shifted to its present configuration, more or less, in order to accommodate all three features as well as possible. The continuing formation of Beta would also reorient the inertia axes and could account for the present 3° departure of the rotation axis from the axis of greatest inertia.

References:


Figure 1. Venus Topography. Elevations relative to a 6052 km mean radius are represented by 324 two dimensional quadratic b-splines, with centers at 10° intervals in latitude and longitude. Major features on the equator are Ovda (100° E), Thetis (135° E) and Atla (200° E) Regiones. The other major topographic feature is Beta Regio (30° N, 285° E). Contour interval is 1 km, and negative contours are dashed.

Figure 2. Effective Isostatic Compensation Depths. Depth excursions relative to 150 km are reconstructed from quadratic b-spline representation of dipole moment distribution, obtained by least-squares fit to PVO gravity data. Atla and Beta are compensated at great depths, whereas the centers of Ovda and Thetis are compensated shallowly, but are flanked by deep compensation. Contour interval is 50 km, and negative (shallower) contours are dashed.
AN EXAMINATION OF THE ROLE PLAYED BY SURFACE SEDIMENT THICKNESS IN DETERMINING MARTIAN CRATER EJECTA MORPHOLOGY. Eric B. Grosfils, Departments of Physics and Geology, College of William and Mary, Williamsburg, VA. Advisor: Nadine G. Barlow, Lunar and Planetary Institute, Houston, TX.

INTRODUCTION: Using Viking photomosaic images, careful examinations of martian impact craters suggest the surrounding ejecta blankets are emplaced primarily by debris flows. Early experimental attempts to gain insight into flow mechanisms imply the flows result from penetration of a high-energy impactor into a volatile-rich layer (1). Although the presence of subsurface volatiles is strongly suggested, alternate theories now postulate, and experimentally support, that the observed ejecta patterns can result from atmospheric interaction with ejecta material, either with--or without--the presence of volatile-rich subsurface layers (2). Further, recent studies suggest differences in ejecta morphology may result at least partially from varying depths of penetration into both the volatile-poor surface sediments and the underlying volatile-rich layers (3). Initial investigations have attributed differing strengths to the causal link, primarily due to limited sample sizes or areas of investigation as well as variances in method of approach (4,5). Following publication of the "Catalog of Large Martian Impact Craters" (6), an extensive global database now facilitates the current study, which addresses the need to examine the distribution of volatile-poor surface sediment thicknesses on a large scale in order to determine the nature of the relationship between the penetration depths and resultant ejecta morphologies of impact craters on Mars.

METHOD: The study initially employs two separate data sets, one taken from Barlow's catalog, the other generated by Rene De Hon of Northeast Louisiana University (7). De Hon's data consists of roughly 750 thickness values listed by longitude and latitude. The thicknesses were calculated by De Hon using the following rim-height analytical equation developed for fresh impact craters by Pike and Davis (1984):

\[ h = 0.076 D^{0.474 \pm 0.091} \]

where \( h \) is the height of the rim above the surrounding terrain and \( D \) is the rim-to-rim crater diameter (8). This procedure is now thought, due to complications involving erosion of the original rim, to give thickness values which are too large. Barlow's data consists of roughly 3400 buried crater diameters listed by longitude and latitude. In this instance, "buried crater" refers to a crater which has been infilled so that the sediment level approximates the original planar surface prior to impact. By applying the equation

\[ d(a) = 0.212 D^{0.544 \pm 0.094} \]

also from Pike and Davis (1984), relating crater diameter \( D \) to the depth of excavation \( d(a) \) below the pre-impact surface, thickness estimates are obtained at each buried crater location through calculation of the crater's original depth. This method involves errors inherent to the approximation that the crater is filled to exactly the level of the pre-impact surface. As a result, the thickness values generated, like those of De Hon, are also too large. Although both data sets provide thicknesses greater than those which would result if the assumed and actual situations at the surface of the planet were equivalent, the data serve to define the range into which the true average
thickness across a given area is liable to fall.

Each data set is plotted separately as a function of latitude and longitude. For De Hon's data, this is relatively simple: due to the small number of points involved, all the data plot directly onto a single map of the martian surface. Barlow's values, however, are more difficult. Due to the large number of points involved, the data plots most accurately onto 1:2M scale photomosaics, resulting in roughly 100 individual maps, each corresponding to a subquad located within the area studied, between latitudes 30S and 65N.

Once plotting is complete, the data are contoured. Again, for De Hon this is simple to do; however, Barlow's data requires initial contouring at the subquad scale, transferral onto a single map of the martian surface, and then final resolution of any small-scale edge effects as a single isopach map is generated through interconnection of the subquad contour lines.

With both isopach maps complete, the preliminary phase of the study is concluded. For the next phase, a data set containing fresh crater locations by latitude and longitude is generated using Barlow's catalog. A "fresh crater" is defined as one which is surrounded by a pristine ejecta blanket, implying a lack of both later erosion and burial. By using both isopach maps, two sediment thicknesses are determined for each crater location, and the values are recorded.

RESULTS: A commercial statistical and graphical package, Systat/Sygraph, was used to examine the data. The data for thickness as a function of six different ejecta morphologies (single lobe, double lobe, multiple lobe, radial, diverse, and pancake) are normalized and plotted (See Figure One), generating results for both Barlow's and De Hon's data as follows.

1) With the exception of pancake craters, all of the ejecta morphological types occur dominantly in regions covered by the same sediment thickness (See Table 1). Using Barlow's isopach map, this thickness is 1.25km; for De Hon, the value is 0.35km. Due to the small number of pancake craters examined [17], their preferential occurrence in thicker sediment should be viewed with skepticism.

2) A more detailed examination of the largest ejecta morphology subset, the single lobe, reveals that there is no variance of thickness as a function of 10-degree latitudinal bins (See Table 2).

3) The consistency of the peak values within each data set implies that the results are significant, and not purely a question of contouring imprecision. It should be recognized, however, that the process of generating isopach maps by hand generates a quantifiably uncertain degree of error.

4) Though not precisely measured, the peak and other values reflect the distribution of thicknesses as a relative percentage of the total contoured surface. For example, using Barlow's data, more craters of each type may occur at a thickness of 1.25km simply because the relative abundance of this value across the planet's surface is higher than other values.

CONCLUSION: The consistency of the results within each independent data set suggests that there is no global dependency of ejecta morphology upon the thickness of volatile-poor surface sediments. The results of this study, if accurate, suggest one of three major conclusions. First, if subsurface
volatiles exist, they are either located at greater depths than theoretically postulated or fail to play the role expected in determining ejecta morphology. Second, the concept of volatile-poor versus volatile-rich layers may be inaccurate; it is possible that different layers contain the same proportion of volatiles, and therefore have no net effect upon ejecta morphology. Third, subsurface volatiles may not actually be present, suggesting that the theories about atmospheric or other models must be more closely examined as alternate means of explaining ejecta morphology.

| TABLE ONE | BARLOW: SINGLE LOBE, THICKNESS |
| (Peak Thickness By Ejecta Morphology) |
| BARLOW | DE HON |
| SL  | 1.2km | 0.35km |
| DL  | 1.3km | 0.35km |
| ML  | 1.2km | 0.35km |
| RD  | 1.3km | 0.35km |
| DI  | 1.2km | 0.35km |
| PN  | 1.6km | 0.39km |

| TABLE TWO | (Single Lobe Peak By Ten Degree Latitudinal Bins) |
| BARLOW | DE HON |
| 20S-30S | 1.1km | 0.35km |
| 10S-20S | 1.2km | 0.34km |
| 0-10S | 1.3km | 0.35km |
| 0-10N | 1.3km | 0.35km |
| 10N-20N | 1.2km | 0.35km |
| 20N-30N | 1.2km | 0.34km |
| 30N-40N | 1.3km | 0.36km |
| 40N-50N | 1.2km | 0.33km |
| 50N-60N | 1.2km | ---- |
| 60N-65N | 1.3km | ---- |

Figure One is an example of a normalized thickness plot produced by the Systat/Sygraph software. Crater counts per thickness are shown on the rightmost scale.

REFERENCES
GEOCHEMICAL ANALYSIS OF POSSIBLE EXTRATERRESTRIAL PARTICLES

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In this study two rock types suspected of harboring extraterrestrial components were examined for these components. One type, Libyan Desert Glass (LDG) was studied employing Instrumental Neutron Activation Analysis (INAA) while the other, Jurassic limestone, was examined using Scanning Electron Microscope (SEM).

I. Libyan Desert Glass

Introduction

LDG is found scattered over 6500 km$^2$ area on the SW margin of the Great Sand Sea in Western Egypt bordering Libya. Their age is determined to be 29 My and over 1.4 million kilograms of these glass fragments are estimated to be present in the area [1, 2]. Two structures in the vicinity (BP and OASIS) show evidence of a high velocity impact [3]. It has been suggested that LDGs may have formed during such an impact. The bulk composition of LDG is 97-98 wt% SiO$_2$, 1-2 wt% Al$_2$O$_3$ and trace amounts of Fe, Ti etc. [4]. Although the LDG has restricted major element composition their trace element contents (ppm) vary as much as a factor of 5 to 30 and the dark streaked glasses of LDG contain significantly higher siderophile elemental abundances representing meteorite residue [5-7].

Experimental procedure

Six LDG samples with dark streaks (courtesy J. R. Underwood) were chosen for the present study. The samples were washed and broken into 1 cm size pieces. The dark pieces from each sample were carefully handpicked and powdered. Magnetic fractions were separated from +200 mesh portion of these samples using the Franz magnetic separator. Although the magnetic portion obtained from each sample was very small (varied between 0.05 to 0.85 mg), it was analyzed along with the bulk samples.

To ensure the purity of the samples, several precautions were taken. The samples were ground with an agate mortar and pestle that was rinsed with HCl, water, and alcohol. Before use, the Franz magnetic separator was cleaned with acetone and a small portion of the sample was run through it and then discarded. The quartz vials in which the samples were loaded for analysis were soaked in aqua regia for 45 minutes and rinsed with water and alcohol.

The samples were irradiated at the University of Missouri Reactor and counted in the JSC INAA laboratory. The standards for INAA include Flyash, DTS-1 and SARM-7 and Allende meteorite; data reduction was accomplished utilizing an upgraded version of the TEABAGS computer program for INAA [8].
Results and conclusion

The samples of the present study show chemical trends identical to the LDG data reported earlier. The \((\text{La}/\text{Sm})_N\) ratios and the negative Eu anomalies are comparable to the LDG data of Murali et al. (1988). The magnetic fractions of the samples are enriched in Fe, Sc and Cr by a factor of 1.3, in Co by a factor 2 and in Ni by a factor of 5 compared to the bulk samples. Four samples of the present study (BNM, CNM, DNM, and LIB-4) show significant Ir content (~1 ppb) and higher Fe, Ni and Cr values.

The present study clearly indicates that some of the samples with dark streaks have extraterrestrial signatures. Further work is needed to characterize the projectile composition.

II. Jurassic limestone

Introduction

The discovery of the Ir and Co enriched extraterrestrial component in the Cretaceous/Tertiary boundary clay has spawned myriad studies trying to find other evidence of extraterrestrial material. One such study was the discovery of spherules in K-T boundary clay layers in Denmark [9]. Their study demonstrated that meteor ablation spheres can survive 65 million years and that high Ir values serve as useful markers for likely locations of spherules.

A previous INAA study of Callovian-Oxfordian age (169-156 Ma) limestone from a section of the Cracow-Czestochowa Upland of southwest Poland showed that the section was enriched with Ir, Co, and light and middle REEs when compared to the North American Shale Composite (NASC) [10], probably indicating an extraterrestrial component. A sample of this rock was obtained and searched for meteorite ablation spheres.

Method of Analysis

The limestone was etched with dilute HCl and the residue was sieved to a size between 0.105 mm and 0.225 mm. This residue was run through a Franz magnetic separator. Polished grain mounts were made from the magnetic fraction. These mounts were carbon coated and then analyzed using a JEOL 35CF SEM to ascertain internal grain morphology. The presence of barred olivine with interstitial glass and small euhedral magnetite crystals would indicate an object likely to be extraterrestrial in origin. A Princeton Gamma Tech Energy Dispersive Spectrometer (EDS) attached to the SEM was used to determine the bulk composition of the grains. Eight sections were carefully examined. After initial examination, two sections were polished down further to permit viewing of a lower layer of the sample.

Results

No particle that could unequivocally be considered a meteorite ablation sphere was found. Most of the particles present were rich in Si with lesser amounts of Fe, Al, K, and Mg and appeared to be terrestrial sediment.
Conclusions

It seems likely that any meteorite ablation spheres that might have been present originally in the sediment have dissolved in the course of 160 million years. The enhanced levels of Ir and Co detected by the previous INAA study could represent extraterrestrial material that has been adsorbed onto the sediment grains after the dissolution of the original spheres. It is also possible that the spheres did not survive the etching process, but previous investigations indicate otherwise [11].

References


Introduction

The Viking orbiter returned 47,622 images covering the entire surface of Mars at varying resolutions on the order of tens to thousands of meters per pixel, providing a valuable database for photogeologic interpretation of the planet's surface. However, successful interpretation, particularly of small scale geologic features, requires the highest resolution (<30 m/pixel) available [1], and therefore, organization of the high resolution data based on geologic content would be of considerable use. Cataloging of the highest resolution (<20 m/pixel) Viking images has already been done [2,3], to supplement this we have compiled a catalog of the 1748 Viking images which meet the next set of resolutions parameters, those between 20 and 25 m/pixel, according to their salient geologic features.

As a result of this venture a reappraisal of the dark (low albedo) talus streaks on Mars has been possible. We propose an alternative interpretation in which they result from a combination of mass wasting and aeolian processes.

Background

The primary tool used for image cataloging has been the Image Retrieval Processing System (IRPS) [3]. The catalog format includes photometric, geographic, terrain type, geologic information, and has a comments section for the inclusion of enigmatic features (Figure 1).

Dark streaks extending down talus slopes have been found in several locations on Mars [4,5]. These streaks, and those in the present investigation, are shown in (Figures 1, 3, 4). Dark talus streaks in the Olympus Mons aureole were interpreted as erosional products of blocks of dark material distributed randomly throughout the aureole deposit [4], while in Corpates and Candor Chasmata, dark streaks have been interpreted as recent mafic pyroclastic deposits, which have moved down the local slope [5].

Results

The following observations of the dark streaks were derived from photomosaics, stereographic projections, and the plotting of the global distribution of the dark talus streaks identified in this study and other sources [3,4,].

A. Dark talus streaks are always associated with scree slopes, but not all scree slopes have dark streaks.
B. In every case dark talus streaks parallel the local topographic gradient.
C. Dark talus streaks display a variety of contrast intensities.
D. The upslope ends of streaks are not confined to a specific level on the scree slope. Many talus streaks extend from, or lie immediately beneath, distinct outcrops.
E. Some dark talus streaks converge, while others diverge into fan shapes.
F. There is no latitudinal correlation between different locations of dark talus streaks.
G. Dark talus streak occurrence is not dictated by terrain type.
H. Dark talus streaks are often associated with aeolian landforms.

Discussion

Earlier interpretations of these talus streaks were found to be inconsistent with our observations of both the locations and the characteristics of the dark talus streaks [4,5]. For instance, many talus streaks are located in Arabia Planitia in an
area mapped as a Cratered Plains Unit [6]. There is no evidence of recent volcanism in this area, so if the streaks are a result of weathering pyroclastic blocks they would have to be ancient. Further, in this same location, dark talus streaks originate from all levels of the regolith, to depth approximately 2 kilometers below the mean surface. This implies that they are a result of a local process. Other dark streaks which have been attributed to recent pyroclastic origin demonstrate tectonic alignment [5], a characteristic uncommon to the majority of the talus streaks.

Dark talus streaks are similar in appearance to dark aeolian streaks. The later are thought to result from materials with albedos lower than the surrounding terrain- the result of local aeolian erosion [7]. Therefore, it is possible to account for dark streaks without invoking heterogeneous material.

Normal landslips, part of the development of terrestrial scree slopes, may be the cause of the dark talus streaks on Mars. During the landslip event, transport of the scree material could be sufficiently energetic to disrupt the existing dust fallout cover (high albedo) and send it into suspension. This fine, high albedo material would be redistributed during suspension by wind and little would return to the original location. This process would lower the albedo of the area enough to cause a contrast between the recent slip and the surrounding terrain.

If the talus streaks are formed by landslips it should be possible to estimate the rate of talus evolution based on the rate of dust deposition. The present rate of aeolian deposition of fine (high albedo) dust during global dust storms is ~20-25 micrometers per Earth year [8]. At this rate a one meter thick layer of fine dust would be deposited on these features in 25,000 years, a thickness probably sufficient to raise the albedo of the average talus streak significantly, implying that dark talus streaks are recent ephemeral features. In addition, the varying contrast intensities of the talus streaks indicate that a series of landslip events can be seen on given slope; the lower the albedo of a streak, the more recently it has occurred.

Conclusion
The main application of the geologic catalog of 20-25 meter resolution images will be to save the researcher time when searching the Viking Orbiter image data base. In addition, the catalog provided the data set which allowed for the reexamination of the origin of the dark talus streaks by locating more occurrences of the talus streaks, making it possible to plot their geographic distribution. Talus streaks need not be caused by weathering of an heterogeneous material, instead it is quite possible that they appear darker than their surroundings because the locally high albedo dust cover was removed by a relatively recent landslip.

References
Table 1.

<table>
<thead>
<tr>
<th>ID</th>
<th>LAY.</th>
<th>TOP</th>
<th>TWR</th>
<th>GCN</th>
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EMAK = Emission Angle  
PHA = Phase Angle  
IKA = Incidence Angle  
III = Image Height  
IFW = Image Width  
RNG = Range  
GCBN = Geographic Region (by quadrangle)  
TRTP = Terrain Type (from 1:5,000,000 USGS geologic maps)  
VLCN = Volcanics (cones, flow features, craters)  
TCTN = Tectonics (large scarps, small grooves, ring features)  
IPPE = Impact Features (bowl shaped, lobate ejecta, floor features: fracture, fill, rim features: terracing, slumping)  
CHNL = Channels (channel networks, large outflow, fretted, chaotic terrace, small channels, valley fill)  
ENPP = Eolian Features (dune deposits, yardangs, streaks; bright, dark, other eolian)  
TSPP = Terrain Softening Features (lobate debris apron, linedated valleys, concentric crater fill, general softening)

---

Figure 1.

Table 1.

Figure 2. (from Morris) [Viking orbiter image 441B02]  
The arrow indicates one example of a dark talus streak.

Figure 3. (from Lucchitta) [Viking Orbiter image 081A04]

Figure 4. A portion of Viking Orbiter 1 image 713A70.  
The arrow indicates a typical example of the dark talus streaks which we examined (note similarities to talus streaks in Figure 2.)
INTRODUCTION

Eucrite meteorites fall under the general classification of basaltic achondrites, and are composed mainly of calcic plagioclase and calcium-poor (pigeonitic) pyroxene. While many of them are breccias, several eucrites (e.g., Ibitira) exhibit an igneous texture, and are believed to be extrusive basalts (1). On the basis of geochemical, petrologic, and isotopic characteristics, eucrites, diogenites, howardites, mesosiderites, pallasites, and IIIAB iron meteorites are all believed to be pieces of a single parent body, referred to as the Eucrite Parent Body (EPB) (e.g., 2,3). Investigation of the EPB is a unique opportunity to look at the cross section of a body which experienced processes of early planetary differentiation analogous to those experienced by the Earth.

There are basically two models which describe the formation of eucrites; the first proposes that eucrites were produced by extensive fractional crystallization of more magnesian liquids (4), based on the simple progression of major element chemistry. The second model proposes that eucrites formed from primary melts (5); this approach is based on Stolper’s melting experiments of actual eucrite samples, showing that most eucrites cluster around the olivine-pyroxene-anorthite peritectic point on a quartz-olivine-anorthite phase diagram. Stolper’s results suggested that the pyroxene-olivine phase boundary was a reaction line (5), with pyroxene crystallizing at the expense of olivine. This crystallization path moves the residual liquid into the pyroxene field, missing the peritectic point where most eucrite compositions cluster, thus rendering an origin through fractional crystallization unlikely. Later work (e.g., 6,7,8) found the olivine-pyroxene phase boundary to be a cotectic; simultaneous crystallization of olivine and pyroxene would lead a residual liquid to the peritectic point. Ultimately the phase boundary appears to be dependent on pressure effects as well as the Fe/Mg ratio; this complicates the phase relations and makes a final determination between the two genetic models difficult, based on major element chemistry alone.

Trace element concentrations may also be used to constrain the melting and crystallization history of a suite of rocks. Although Rare Earth Element plots of the eucrites are relatively flat, suggesting an originally chondritic source, other elements, such as Sc, show a marked depletion. Scandium is a compatible trace element in mafic systems; it is taken into the lattice structure of both pyroxene and olivine. Earlier work (6), using Sc partition coefficients (\( \text{Sc}_{\text{D}} = \frac{\text{Concentration of Sc in solid}}{\text{Concentration of Sc in liquid}} \)) for olivine taken from the literature indicated that olivine alone as a residual phase could account for the Sc depletion seen in eucrites. Partition coefficients, however, are dependent on composition; because the \( \text{Sc}_{\text{D}} \) was not specifically measured for eucrites, the results were claimed accurate only to within 20% (6). We have experimentally determined Sc partition coefficients for olivine and pyroxene in eucritic melts, and used this new data in the modeling equations of (6) to further constrain the melting history of the eucrites.
ANALYTICAL METHODS

A synthetic eucrite composition (based on Sioux County) and containing 1 wt% Sc, was prepared using oxides; this composition was then doped with 10 wt% olivine, moving the sample into the olivine field and ensuring olivine crystallization. A 125 mg sample was placed on a platinum wire loop and heated in a Deltech furnace to 1250°C at iron-wustite fO2; the charge was held at 1250°C for 48 hours and then air-quenched. This initial run was analyzed on the Electron Microprobe in order to detect the iron lost from the sample to the platinum wire. A 3 wt% iron loss was detected and corrected for in subsequent runs. Compositions which would crystallize pyroxene were created by adding 5 wt% SiO2; this moved the sample from the olivine field into the pyroxene field.

Four experiments were run, involving eight separate charges. The first two charges were held above their liquidus for two hours (see Table 1 for exact experimental parameters), then dropped to liquidus temperatures, where they were held and allowed to grow crystals for between 58 and 68 hours. The samples were then air-quenched; we achieved good crystal growth in all four charges. Analysis of the glass in these first samples indicated that we were still experiencing iron diffusion from the sample to the platinum wire during crystallization of olivine. The next two experiments were held above their liquidus for 24 hours in an attempt to come closer to equilibrium with the wire; they were then allowed to grow crystals at their liquidus for 24 hours, before being air-quenched. All samples were analyzed on the Electron Microprobe for Sc, Fe, Mg, Al, Si, and Ca.

RESULTS

ScD results (see Table 2) for olivine are lower than those used by Jones (6): 0.2 vs. 0.3. Fe and Mg in the olivine crystals show some reverse zoning; however, our Fe/Mg Kp’s for olivine are in good agreement with those of Stolper (5), and we believe that the samples come as close to equilibrium as is possible on a laboratory time scale.

ScD results for pyroxene are lower than predicted by Jones (6), and range in value from 0.72 to 0.83. Our Fe/Mg Kp’s for pyroxene are also in agreement with Stolper; Fe and Mg, however, do show some zoning when plotted vs. distance. Since the pyroxenes were dendritic and it was difficult to probe a traverse across a crystal, a final pyroxene experiment has been run in which we use a four stage cooling procedure to allow slower, more uniform crystal growth. These samples have not yet been analyzed.

CONCLUSIONS

Combining mass-balance equations with bulk partition coefficient equations, and assuming equilibrium partial melting of an eucrite parent body with originally chondritic element ratio’s (Sc/La and Mg/Al), it is possible to solve for the partial melt percentage, as well as the percent olivine and the percent pyroxene remaining in the eucrite source region. Two basic assumptions involved in this procedure are the fact that LaD and AlD are both zero, i.e., La and Al are completely incompatible and will concentrate in the melt.

Using MgD_{ol}, MgD_{px}, ScD_{ol}, and ScD_{px} from analyses of our experimental charges, Sc/La and Mg/Al ratio’s for Sioux County from the literature (8), and Sc/La and Mg/Al chondrite ratio’s from (9), we obtained a
unique solution of 19% partial melt, with 98% olivine and 2% pyroxene remaining in the source region. If the constraint of a chondritic element ratio is relaxed, for either Mg/Al or Sc/La, the equations produce a set of non-unique solutions, which are shown graphically in Fig. 1. The intersection of these two curves is the unique solution mentioned above, and corresponds to 19% partial melt.

If the initial assumptions are accepted, the amount of pyroxene left in the source region is constrained to 2%, or, alternatively, the amount of pyroxene that could have fractionally crystallized and been removed from the eucrite parent magma is 2%. This is a small amount of pyroxene with which to make cumulate pyroxene rocks, such as diogenites. With these results it is difficult to propose a genetic relationship through fractional crystallization for the eucrites and diogenites; it appears that the simpler model of producing eucrites from a primary melt is more likely.

REFERENCES:

Table 1: Experimental Parameters

<table>
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<th>Sample #</th>
<th>Initial Temp (°C)</th>
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All samples run at iron-wüstite oxygen fugacity. All samples air-quenched.

Table 2: Partition Coefficients

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<td>CP-6</td>
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<td>Stolper*</td>
<td>†</td>
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</table>

KD = [(Fe-Mg)_{crystal}]/[(Fe-Mg)_{liquid}]
ScD = [(Sc)_{crystal}]/[(Sc)_{liquid}]
† No analysis
* Stolper, (1977) determined in experiments on Juvinas and Sicor County
EDDIES OF THE WORLD OCEAN -- GEOGRAPHIC DISTRIBUTION AND FUNDAMENTAL CHARACTERISTICS. K. E. Sheetz, Millersville University, Millersville PA. 17551, and C. A. Wood, Johnson Space Center, SN15, Houston, TX. 77058

INTRODUCTION
An eddy is a circular movement of water that embodies a concentration of energy in a relatively small three-dimensional space. The whirlpool motion of ocean eddies is responsible for the exchange of fluid properties such as momentum, mass, heat and suspended matter throughout the seas and oceans of the planet Earth(1). Eddies can range from submesoscale (about 1-50 km in diameter) to greater than 500 km across. The understanding of this energy transfer mechanism is important not only for the study of the physics and chemistry of the world ocean, but also for efficient travel across its surface, and tactical proficiency of naval vessels.

As a result of laboratory experiments, modeling, and research vessel observations, eddy activity is most likely found where strong currents interact with each other or with a stationary obstruction(2). Further studies to delineate the precise mechanism responsible for the formation of eddies have been restrained to observing one eddy or small group of eddies at a given time. From such studies, a great deal of quantitative information has been gathered about a limited number of specific eddies. Our research complements these detailed studies by observing, classifying, and cataloging the worldwide distribution of submesoscale eddies. This approach is unique in that rather than analyzing data on one eddy, it is oriented toward discovering whether there are any global patterns or statistical characteristics of these numerous eddies.

METHODS
The synoptic view necessary for this analysis was obtained by utilizing some of the thousands of photographs taken by the 29 space shuttle missions flown since 1981. From this source, a data base of 182 photographs, showing eddies photographed by shuttle astronauts was compiled by R. Nelson, of the Space Shuttle Earth Observations Office, Johnson Space Center. Each eddy was then classified according to the following: geographic region, latitude, longitude, eddy type, size (diameter), rotation, proximity to land, and time of year.

Eddy types are broken down into six basic groups -- 1) Spiral eddy- an isolated eddy occurrence, that is clearly rotating in one direction and not interacting with other eddy features. 2) Complex eddy field- a complicated integration of two or more eddies, often having a chaotic appearance. 3) Mushroom eddy- two eddies rotating in opposite directions, formed when a current flows into a boundary (either land or another current) and splits. 4) Wake eddy- an eddy that is formed in the wake of an island or some permanent structure 5) Linear spiral eddy series- eddies that are aligned in a straight line rather than randomly. Such alignment is thought to be due to their having just been generated, possibly in conjunction with the lineated shears on one side of a current (3) 6) Headland eddy- formed in a process much like the mushroom eddy against land, with the exception that only one eddy and rotation direction is established.

Eddy diameter is calculated using the simple parallax relationship: Actual size=(measured size on film * orbiter altitude)/ camera focal length.
This method, although exact only for photos taken normal to the Earth's surface, probably only introduces errors of a few kilometers in oblique views.

Rotation is classified as either cyclonic or anticyclonic. Cyclonic motion is counterclockwise in the northern hemisphere, and clockwise in the southern hemisphere.

Visible land is a rough measure of how close the eddy is to some land feature. Each eddy is classified as having one of the following: Land in the photograph, land in the photograph within one eddy diameter, or no land in the photograph. This becomes important when investigating formation mechanisms.

Finally, the time of year, is the three month interval (season) that the photograph was taken.

By cataloguing the eddies in this fashion, we were able to produce graphs and map plots according to any of the above classification fields, thus enabling us to relate the worldwide distribution of ocean eddies to parameters such as geographic location, season, latitudinal variation etc.

Although the space shuttle photographs give us a useful synoptic view, there are precautions that must be taken in the interpretations of the initial data results. The problem lies in selection. Mission times, orbit patterns, cloud cover, sun angle and astronaut attention are all constraints on the selection of photographs available for study. Is what we are seeing a complete record of eddies? Has most of the ocean surface been photographed? Is there a bias in the selection due to surfactant presence (floating debris, oil or biological remnants that suppress surface roughness, allowing sunglint to highlight certain areas with respect to the surroundings)? Some of these potential biases, we have evaluated and dismissed, but a few will remain a question until selection factors are better understood.

RESULTS

Spatial Distribution:

The 221 individual eddies mapped have an apparently irregular distribution across the world ocean. However, 82% of these eddies are found within 320 km of a major land mass. This tendency for eddies to form along continental margins, and in shallow seas, is evidence that shearing motion along coastal currents or sea floor topography may play an important role in eddy formation. The latter mechanism is supported by the concept that potential vorticity, or the tendency for a parcel of water to rotate, is governed by changes in the depth of the water. (4) This may be an important factor in the formation of many shallow water eddies, but detailed bathymetric maps show that many of the open ocean eddies overly regions of little or no submarine topography. When these open ocean eddies are plotted on a current map, however, a remarkable tendency for formation along the horizontal shear zones of current boundaries becomes evident. The North Pacific-North Equatorial, Labrador-Gulf Stream, and the Agulhas-Indian Ocean current boundaries are a few of the major shear zones throughout the world ocean that collectively host over 80% of the open ocean eddy structures photographed by shuttle astronauts.

Size Distribution:

The diameters of the eddies range from 2 km to just under 60 km. Figure 1 shows a histogram for the distribution of these diameters, as grouped in bins of 5 km. The tendency for eddies to form in the 5-10 km bracket is quite clear. This shows us that, although the larger mesoscale and synoptic eddies have drawn most of the scientific attention, these submesoscale features are also significant functions in the energy control process.
**Latitudinal Distribution:**

Figure 2 illustrates the breakdown of eddy sites into 5 degree latitude bins. This shows that nearly all of the eddies on the northern hemisphere (the southern hemisphere follows a similar pattern) lie above 15 degrees. The sinusoidal nature of the spacecraft's orbit allows it approximately the same amount of time over each 5 degree bin per orbit, thus eliminating orbit bias as a possible reason for the lack of eddies in the equatorial regions. One must look instead at the Coriolis force to explain not only the latitudinal distribution, but also the fact that 98% of the eddies exhibit cyclonic rotation. The Coriolis force, which is due to the earth's rotation, balances the horizontal pressure gradient in such a way as to favor cyclonic rotation over anticyclonic(2). The Coriolis parameter is the following: \(2\Omega \sin \phi\), where \(\Omega\) is the earth's angular rotation rate, and \(\phi\) is the geographic latitude(5). As \(\phi\) approaches the equator, the Coriolis force approaches zero, thus reducing the probability of rotation and instability that forms eddies.

**CONCLUSION:**

Through the use of photographs taken from low Earth orbit, a synoptic view of eddies in the world ocean is made available. Most of the photographed eddies have diameters in the 5-10 km range. 82% of the eddies occur at continental margins, where coastal currents and shelf topography govern energy dissipation processes. The open ocean eddies, though fewer in number, tend to occur along the horizontal shear zones associated with major current boundries. Unresolved questions include the controls of eddy diameters and the effects of Coriolis force in eddy formation. With more data and further investigation, we hope to answer these questions.

**REFERENCES:**
