## PROGRAM

**Morning session - 9:00 to 11:30 am**  
Chairman: David Gust

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<td>Claire Duncan</td>
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**Afternoon session - 1:00 to 3:10 pm**  
Chairman: Jim Zimbelman

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<tr>
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<tr>
<td>1:00</td>
<td>Laurel Kaczor</td>
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<td>Rachel Lerner</td>
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This year each of the LPI Summer Interns will give a presentation on the results of their 10-week research project at a "mini-conference" to be held Wednesday, August 14th in the Berkner Room. In past years these presentations were given during the noon seminars on the last five Fridays in the program. Dave Gust and Jim Zimbelman have originated this new procedure in the intern program in the hope that by having all the presentations given at the same time, the interns would all have the same time to devote to their projects prior to their presentations.

Anyone interested in attending the conference is welcome. A program for the conference is attached.

Attachment
Papers presented at the
SUMMER INTERN CONFERENCE

August 14, 1985
Houston, Texas

1985 Summer Intern Program for Undergraduates
Lunar and Planetary Institute

Co-sponsors: Lunar and Planetary Institute
NASA Johnson Space Center

Compiled by the
Lunar and Planetary Institute
3303 NASA Road One
Houston, Texas 77058
PROGRAM OF PAPERS PRESENTED AT THE SUMMER INTERN CONFERENCE
Lunar and Planetary Institute
August 14, 1985
Houston, Texas

MORNING SESSION
9:00 - 11:00
Chairman: David Gust

<table>
<thead>
<tr>
<th>Time</th>
<th>Author and Title</th>
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<tr>
<td>9:00</td>
<td>Gordon Alexander</td>
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<td>U and TH Partitioning between accessory minerals and a carbonic fluid phase: Implications for granulite genesis.</td>
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<td>Jonathan K. Bareng</td>
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<td>Investigation of mantle metasomatism in an oceanic alpine Massif-Trinity Peridotite.</td>
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<td>9:40</td>
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<td>Thermal model for Mars.</td>
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<td>James M. Brenan</td>
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<td>Shock induced melting of granitic rocks: Experiments to 754 kilobars.</td>
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<td>Volcano census of the Central Andes.</td>
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<td>Debra Colodner</td>
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<td>Localized melting in the Shergotty meteorite: Implications for Shergotty's shock history.</td>
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<td>Claire D. Duncan</td>
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<td>Ashflow caldera fields: A synthesis and interpretation.</td>
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AFTERNOON SESSION
1:00 - 3:10 p.m.

Chairman: James Zimbelman

1:00 Laurel Kaczor
    Stratospheric particle size and chemistry distributions
    for flat-plate collection surface from 1984, with
    comparison to data from the past decade.

1:20 Rachel Lerner
    Does CO₂ play a role in the igneous petrogenesis of
    anhydrous granites?--A study of proterozoic granites from
    the Nain Province, Labrador.

1:40 Laurie A. Leshin
    Thermal inertias for the Elysium and Aeolis quadrangles
    of Mars.

2:00 Martin L. Lieurance
    Solar Max; micro-material analyses.

2:20 COFFEE BREAK

2:30 Glenn M. May
    Lake Bonneville: A model of isostatic adjustment.

2:50 Eric B. Rehwoldt
    Survey of early earth volatile inclusions within Archean
    sediments.
Geochemical analyses of both low and high grade metamorphic rocks indicate that metamorphism at the granulite facies (usually taken as 6-12 Kb, 700°-1000°C, conditions appropriate to the lower crust) often results in the loss of radiogenic elements, primarily K, U, and Th (1). Two mechanisms have been proposed to account for this depletion.

Since these large ion, lithophile elements (LILE) partition strongly into the liquid phase during partial melting, it has been suggested that partial anatexis and extraction of a granitic melt has left behind LILE-depleted residual granulites (2). Evidence for this granite-granulite link consists of petrographic indicators of anatexis in some granulites, field relationships indicating cogenesis of some granite-granulite pairs (3), reciprocal and Rare Earth Element (REE) fractionation patterns observed in these suites (4,5).

Alternatively, the removal of LILE may be explained by metasomatic stripping by a CO₂-rich fluid phase. The existence of CO₂-rich fluid inclusions found in many granulites (6,7,8) supports this theory, and granulites which show little or no LILE depletion lack CO₂-rich inclusions (9). Amphibolite facies rocks commonly associated with granulites but show no depletion in LILE (10,11) and contain H₂O-rich, CO₂-poor inclusions (12). This mechanism is in agreement with the observed depletion of only the highly mobile LILE, especially K, Rb, Th, U, and Cs (11), while purely magmatic processes would result in depletion of all the incompatible elements, notably the light REE.

U and Th reside almost exclusively in highly refractory accessory minerals in granulites (10) and although large fractions of partial melt could entrain and remove these minerals the fluid inclusions suggest that CO₂-rich fluid is the primary depleting agent. To test this hypothesis a set of experiments has been initiated to determine the partitioning behavior of U and Th between their host minerals, primarily zircon, sphene, apatite, allanite, and monazite, and a CO₂-rich fluid phase.

Natural minerals ground to 60 mesh will be exposed to temperature-pressure regimes ranging from 700°C to 1000°C and 5-8 Kb for 48 to 192 hours. The composition of the fluid will be varied from pure CO₂ to a 50/50 mol% CO₂-H₂O mixture and finally since many inclusions contain trace amounts of halogens a dilute solution of HCl will be used.

The U and Th concentrations in natural samples of these minerals
are low, on the order of 50-600 ppm. Assuming strong partitioning of U and Th into the fluid, their post-run concentrations will be much lower. Experimental data (13,14) indicate that these concentrations may be near the lower limit of Henry's Law, which states that the partition coefficient \( D \) (a measure of the relative preference of an element for a crystalline phase and its coexisting liquid) is constant with respect to changes in the total concentration of the element in the system. To ensure the applicability of Henry's Law and thus the validity of the calculated partition coefficients, synthetic zircon, sphene and apatite have been prepared as follows. Zircon was synthesized by sintering stoichiometric amounts of baddleyite \((\text{ZrO}_2)\) and cristobalite \((\text{SiO}_2)\) in air at 1450°C for 48 hours. This produced euhedral crystals of zircon 6-8\( \mu \) in size according to the reaction:

\[
\text{ZrO}_2 + \text{SiO}_2 \rightarrow \text{ZrSiO}_4 \text{ (zircon)}.
\]

Sphene was produced from its constituent oxides by complete melting and subsequent cooling along the liquidus in the pseudobinary system albite-sphene (fig. 1) using a starting composition \( \text{Sp}_{0.08}\text{Ab}_{0.2} \). This technique yielded elongate prisms of sphene in a matrix of 3% albitic glass. Apatite was synthesized from a starting mixture of dibasic calcium phosphate, fluorite and calcium oxide according to the reaction:

\[
6\text{CaHPO}_4 + \text{CaF}_2 + 3\text{CaO} \rightarrow 2\text{Ca}_2\text{F}(\text{PO}_4)_3 + 3\text{H}_2\text{O}.
\]

(apatite)

The starting mix was heated in air at 1400°C for 36 hours, reacting in the solid state to form fluorapatite crystals up to 1500\( \mu \) in size.

A fundamental criterion which must be satisfied to establish the validity of these data is the demonstration of equilibrium. Varying the duration of the experiments is a direct test, since time-invariant data indicate that equilibrium or a near-equilibrium state has been reached. Another approach is to reverse the process, starting with U,Th free minerals and a radioisotope rich fluid phase. The data thus obtained will bracket the equilibrium partition coefficient.

A set of low mineral/fluid partition coefficients will firmly
establish the capability of this mechanism to account for the observed depletion of granulites in these LILE. Conversely, a set of high coefficients effectively rules this out as a viable process. In view of the strong evidence for CO₂ metasomatism and against anatexis, it is certain that high partition coefficients indicate that the LILE removed by metamorphism were not resident in the accessory minerals but occupied another site, possibly along grain boundaries within the rock (10). If this proves to be the case, a second set of tests will be conducted to confirm the ability of CO₂ metasomatism to remove LILE from grain boundaries, accounting for the depletion of granulites in these mobile LILE and thus the low radiogenic heat production of the lower crust.

REFERENCES

Mantle metasomatism (Bailey, 1982, Wilshire, 1984) is a proposed explanation for the decoupling in incompatible element concentration and isotopic composition of some volcanic rocks, i.e. alkali basalts, basanite and kimberlites. A popular view holds that fluids rich in incompatibles percolate through and interact with the upper mantle, causing compositional changes, particularly enrichment in incompatible elements. Work on ultramafic xenoliths found as inclusions in basalt (for example Wilshire and Shervais, 1975) has produced evidence supporting this metasomatic process, in the form of veins and interstitial secondary minerals in lherzolite.

The purpose of this project is to investigate a similar phenomenon in peridotite massifs. Investigation of rare element distribution in veined oceanic and continental alpine massifs may provide corroborating evidence of mantle metasomatism as viewed in xenoliths. A prime motivation for using peridotite massifs in this study is that they provide a much larger scale than xenoliths in which one can examine the extent of chemical modification due to metasomatic fluids.

A matter of significant importance is the extent of alteration. Most peridotite massifs, including the areas studied in this project, have been at least partially altered to serpentine and magnetite. The serpentine form cross-fiber veins or mesh texture (Quick, 1981).

The majority of samples studied are from the Trinity Peridotite, which underlies an extensive area in the eastern Klamath Mountains of northern California. Regional mapping and gravity and magnetic modeling suggest that the peridotite forms and easterly dipping sheet that is sandwiched between and tectonically related to the overlying eastern Klamath Subprovince to the north and east and the underlying central Metamorphic Belt (Quick, 1981). The Trinity Peridotite is inferred to have ascended through the upper mantle from a depth of not less than 30km. En route to the base of the crust, the peridotite deformed plastically, reacting with transient melts generated at greater depth and partially melted. The massif is dominantly harzburgite and lherzolite, with dunite and clinopyroxene lithologies. Multiple pulses to transient melts formed clinopyroxene-rich dikes, ariegite dikes, dunite bodies and gabbro plutons and dikes. Minerologic and field constraints suggest that the melts were initially picritic basalts and that they fractionated in the direction of tholeiitic basalts.

Samples were collected in the vicinity of Vicki Bluff, 3km south of China Mountain. Sequences of samples were chosen such that they represented traverses across veins or dikes. In addition vein and dike materials were sampled. Quick performed microprobe analyses which showed the olivines in the peridotite are all Mg-rich, ranging in composition from Fo88 to Fo93. Orthopyroxenes also proved to be Mg-rich, showing a range of En91Wo4Fs8 to En87Wo4Fs9. Clinopyroxenes showed a narrow range of compositions from En54Wo47Fs9 to En49Wo48Fs3. The An-contents of most plagioclase in the plaglherzolite range from An83 to An93, though some grains were as sodic as An81-82.

To measure possible effect of metasomatism, instrumental neutron activation analysis was implemented to observe variation of rare earth element abundances at increasing distances from the vein. Samples were crushed and least altered chips separated for powdering. In addition to
whole rock analyses, mineral separation of clinopyroxene was attempted for samples with sufficient modal proportions of that mineral. Details of sample preparation and data reduction are discussed by Jacobs and others (1977). These samples, along with USGS standards ECR, BHVO, and DTS were irradiated in the reactor of University of Missouri, at a flux rate of $10^{14}$ neutrons/cm$^2$/s.

At this time analyses are still in progress. Data collection on vein REE composition has been very recently completed, while whole rock analyses continues. Mineral separation of clinopyroxene is in progress. Preliminary work on REE abundance data of the veins has begun. Figure 1 demonstrates variations in altered and unaltered vein material of sample W8. Figure 2 compares two samples of W16, one containing secondary mineral inclusions and the other uncontaminated. These two diagrams demonstrate that hydrothermal alteration produces substantial changes in the REE composition, and must be considered when attempting to model the consequences of metasomatism. Figures 3 and 4 both display apparently little or no variation across the veins represented by samples W5 and W8. There is also little variation in REE content for samples of different veins in the massif, suggesting that little variation in the melt composition existed. Once data collection is completed, the question of mantle metasomatism as evidenced in Trinity Peridotite can be properly addressed.

REFERENCES

Over the years, a variety of computer-based thermal models have been developed for the Moon and other planetary bodies (e.g., Linsky, 1965, 1966; Morrison, 1969; Kieffer et al., 1977). Unfortunately, despite the sophistication of many of these models, their utility to other investigators has been hindered by limited distribution and often enigmatic code. The Mars Thermal Model (MARSTHERM) is an attempt to develop a program that can be more easily understood, modified, and distributed; providing a tool that others may use and hopefully improve upon. With this in mind, corrections and improvements will be actively solicited, and revised versions of both the thermal model and the documentation will be issued on a periodic basis.

MARSTHERM is a FORTRAN 77 program that uses the method of finite differences to compute surface and subsurface temperature variations throughout the martian year. For the sake of computational efficiency, the program is designed to utilize non-constant depth and time intervals, increasing program execution speed with only a small loss in numerical accuracy. The program is divided into five mutually dependent subroutines, each of which solves a specific part of the problem. The default values for most of the constants (the albedo, thermal conductivity, soil density, etc.) are those adopted for the Standard Viking Thermal Model (Kieffer et al., 1977).

The program is based on a model of the martian regolith consisting of a vertical column that has been sub-divided into compartments of varying thickness (Fig. 1). The transfer of heat through the regolith is calculated via the one-dimensional time-dependent heat conduction equation:

\[
\frac{dT}{dt} = \frac{k}{(\rho c)} \frac{d^2T}{dz^2}
\]

where \( k \) is the thermal conductivity, \( \rho \) is the density, \( c \) is the specific heat, \( T \) is the temperature, and \( dz \) is the thickness of the compartment. The equation is solved by substituting finite difference expressions for both derivatives. Expanding both \( T(t+dt) \) and \( T(t-dt) \) in a Taylor series and then solving for \( dT/dt \) gives:

\[
\frac{dT}{dt} = \frac{(T(t+dt)-T(t))}{dt}
\]

(If \( T(t) \) represents the temperature of a particular compartment at the present time "t", then \( T(t-dt) \) and \( T(t+dt) \) represent the temperatures of this same compartment during the previous and subsequent time steps.) Sundqvist and Veronis (1979) developed the expression that was used for the second derivative:
The surface temperature at a given latitude is a function of both the time of day and the position of Mars in its orbit. The program finds the temperature at the surface by solving the energy balance equation:

\[ S * (1 - A) * \cos(i) + k * \frac{dT}{dz} + FF + L * \frac{dm}{dt} = e * \sigma * T^4 \]

where \( S \) is the solar flux (W m\(^{-2} \)) at the current position of Mars in its orbit; \( A \) is the local surface albedo (which changes to 0.65 when CO\(_2\) is present on the ground); \( i \) is the angle of incidence of incoming sunlight; \( k \) is the thermal conductivity of the regolith; \( dT/dz \) is the temperature gradient evaluated at the surface; \( FF \) represents the downward component of heat that is radiated from the atmosphere; \( L \) is the latent heat of vaporization of CO\(_2\); \( dm/dt \) is the mass of CO\(_2\) that condenses out of the atmosphere per unit time; \( e \) is the emissivity of the surface; and \( \sigma \) is the Stefan-Boltzmann constant.

The program uses a finite difference expression for the surface derivative \( dT/dz \). Manipulating the Taylor expansions of the temperatures in the second and third compartments \( T(z + dz) \) and \( T(z + 2dz) \), respectively, gives:

\[ \frac{dT}{dz} \Bigg|_{z=0} = \frac{4T_2 - T_3 - 3T_1}{2dz} \]

where \( T_1, T_2, \) and \( T_3 \) are the temperatures in compartments one, two, and three, and \( dz \) is the thickness of each compartment.

The results of MARSTHERM are in good agreement with those of the Standard Viking Thermal Model (Kieffer et al., 1977), although at high latitudes there is an apparent discrepancy of several weeks in the time predicted for the disappearance of the seasonal polar cap. Since the amount of condensed CO\(_2\) predicted by MARSTHERM corresponds well with the amount calculated by Leighton and Murray (1966), the apparent discrepancy may actually be an artifact that arises from the way the
results of the Viking Thermal Model were presented (Zimbelman, personal communication, 1985).

References


Figures

Fig. 1. Relative compartment sizes and time intervals down to a depth of two meters.

Fig. 2. Temperatures as a function of depth at six hour intervals.
Fig. 3. Daily temperatures as a function of time at depths of 0, 5, and 10 cm.

Fig. 4a. Annual variation of mean diurnal temperatures at 0 degrees latitude.

Fig. 4b. Annual variation of mean diurnal temperatures at 65 degrees latitude.
SHOCK INDUCED MELTING OF GRANITIC ROCKS: EXPERIMENTS TO 754 KILOBARS; James M. Brenan, 1985 Summer Intern, Lunar and Planetary Institute, Houston, Texas 77058

Shock metamorphism in crystalline material is a phenomenon resulting from rapid acceleration of that material due to passage of a shock front. Such a phenomenon, characterized by high strain rates and extremely high pressures, is most commonly associated with hypervelocity meteorite impacts. The purpose of this study is to characterize shock-metamorphic effects in samples of both Hardhat granodiorite and Westerly granite at pressures ranging from 310 to 754 kilobars. Particular emphasis was put on modal abundances of fusion products in an effort to calibrate incipient melt production as a function of peak shock pressure.

The pressure-volume response of a crystalline material to a shock front is illustrated in figure 1. The shock front represents a discontinuity whereby the initial pressure ($P_0$), volume ($V_0$) and internal energy ($E_0$) ahead of the shock wave suddenly increase to $P_1$, $V_1$ and $E_1$ respectively. The transition to the shocked state (as described by the Raleigh-Line in figure 1) is non-adiabatic and yields irreversible work which manifests itself as heat upon adiabatic decompression. Coupled with other thermodynamic considerations, observation of the upward concavity of the Hugoniot curve in figure 1 suggests that entropy, and therefore residual heat, increases with increasing pressure. Thus the amount of incipient melt produced (once fusion has commenced) should vary directly as a function of peak shock pressure.

![Diagram of pressure-volume response](image.png)

Both granite and granodiorite samples were shocked using an apparatus similar to that described by Hörz (1970). Peak pressures for shock experiments on these materials were obtained by the "impedence matching" method described by Hörz (1970), Gibbons and Ahrens (1971) and Stöffler (1972). Modal abundances of phases of interest were obtained using a plane-table digitizer on photomosaics of experimental charges. Quartz and feldspar dominate the mode of the unshocked samples (47.5% quartz, 43% feldspar for Hardhat; 29% quartz, 66% feldspar for Westerly), with lesser amounts of biotite (6% Hardhat, 4% Westerly), opaque oxides (2%
Hardhat, 0.75% Westerly), titanite (1.2% Hardhat, 0% Westerly) and hornblende (<0.5% Hardhat, 0% Westerly). Quartz grains in shocked samples typically exhibit well-developed planar elements, mosaicism and fracturing. Increased pressure (>360 kb) results in nearly complete reversion of quartz to diaplectic glass. Feldspar also fractures at low pressure (<360 kb) and, like quartz, rapidly inverts to diaplectic glass (maskelynite) at higher pressure. Shocked biotites in these rocks typically exhibit kink-banding at pressures less than 360 kb and a general decrease in birefringence as pressure increases. Progressive thermal decomposition of biotite takes place over a range of pressures. Breakdown of biotite begins along grain boundaries and is characterized by the presence of Fe-Ti oxides set in a weakly birefringent mesostasis (possibly some combination of hercynite and/or pyroxene and/or feldspar glass [Chao, 1967; Stöffler, 1971]). Accessory phases in shocked Hardhat samples, such as apatite and zircon, appear to be unaffected. Shocked Hardhat titanite, however, exhibits well developed twinning on (221) which Borg (1970) suggests is the result of lattice translation induced by the passage of a shock wave.

Glass produced by incipient melting (fusion glass) is first observed in the Hardhat sample at 357 kb and in the Westerly sample at 408 kb. Criteria for the recognition of fusion glass are based on the observations of Kieffer (1971), Stöffler (1972), Keiffer et al. (1976) and Schaal and Hörz (1977). Such criteria include vesicularity, green or brown coloring and flow features such as elongate vesicles and schleiren. Fusion glass is typically spatially associated with biotite grain margins and/or in fractures propagating from biotite grains. This association suggests that the difference in compressibility between biotite and tectosilicate grains is sufficient to cause local focusing of energy deposition and enhancement of melting. Figure 2 depicts the modal proportion of melt produced as a function of peak pressure for both Hardhat and Westerly samples. Also depicted are the theoretical melting curves obtained from data calculated by Cintala (pers. comm.).

**Figure 2.**

Dissipation of energy into other forms of work beside melting is suggested to explain the large discrepancy between the theoretical and
experimental melting curves. Energy used for the initiation and propagation of fractures, solid-state phase transformations and lattice distortions may thus decrease the total amount of energy available for fusion. In addition, peak shock pressures were achieved experimentally through multiple-reverberations as opposed to a single-shock event. Gibbons and Ahrens (1971) show that post-shock temperatures associated with multiple-reverberation experiments are 200-1000 °C lower than those associated with single-shock events (for peak shock pressures of 300-500 kb, respectively). As a result, theoretical modelling assuming a single-shock event results in higher calculated post-shock temperatures associated with a given peak shock pressure and greater melt production.

The experimental melting curves portrayed in figure 2 could provide an additional tool in deciphering the complex P-T history of impact structures preserved in granitic terranes. In addition, comparison of the experimental and theoretical data, as in figure 2, facilitates a test of the validity of theoretical shock-melting models for granitic rocks. Such a test clearly shows that further refinement of current shock-melting models is necessary. Thus, combined efforts of experimentation and numerical modelling can ultimately lead to a better understanding of the nature of shock-metamorphic processes.

References


Chao ECT (1967) Shock effects in some rock forming minerals. Science 156: 192-200


BACKGROUND

At the present time, documentation of active volcanism within the Central Andes is inadequate at best. The most complete documentation available is the Catalogue of the Active Volcanoes (L. Casertano, 1963, and G. Hantke and A. Parodi I., 1966). While this catalogue lists a few of the major active volcanoes in Peru and Chile, it does not cover all the active volcanism there, and no mention at all is given of volcanic activity in Bolivia and Argentina, where recent volcanism is abundant.

A detailed census of active volcanism in the Central Andes is urgently needed, and is therefore our long-term project goal. Using high resolution (30x30m) Landsat Thematic Mapper (TM) images, a detailed overview of Andean volcanism can be obtained for the first time. This study will ultimately lead to a more up to date and complete inventory of the active volcanism in the Central Andes, and provide a basis for further petrologic and tectonic studies.

The immediate goal, however, was to use the lower resolution (57x79m) and less expensive Multispectral Scanner (MSS) imagery to locate important volcanoes, major volcanic trends, and to create a data base for use with the subsequent TM studies. Furthermore, two MSS scenes were digitally mosaicked to facilitate image analysis of the most active volcanic segment.

PROCEDURE

The initial step in creating the data base was to mosaic MSS 1:1 million black and white photographic prints of the Andes from approximately 11°s to 39°s latitude, and from 75°w to 64°w longitude. The major volcanoes in this region were located, outlined, and assigned a relative age from one through five (1 = recent, 5-10 Ma) using geochronological and geomorphological data from Baker (1977). Classification was further aided by Shuttle and aerial photographs. The volcanic centers were then color coded by age to reveal possible trends in the volcanic arc's history. Each volcano was systematically entered into the data base using the following parameters: name; type; age; location; summit and edifice height; basal diameter; obvious flows and length of the longest flow; absolute nearest neighbor and nearest neighbor of plus or minus one relative age; and eruption date if recorded in the Catalogue of the Active Volcanoes.

Concurrently with the data base formation, two MSS images were mosaicked together using various computer programs. This process consisted of first finding ground control points (GCPs), or recognizable geographic features or landmarks whose actual geodetic positions could be measured in existing maps (J. Moik, 1980). The GCPs' positions relative to the universal transverse Mercator (UTM) projection were then recorded and stored using the LIPS program GCP. Missing lines and bright borders were corrected before the images could be sampled, or "rubber sheeted" to the UTM coordinates, to remove distortions. Using the image processing program MOSAIC, the two restored images were merged together, to form a continuous image of an important volcanic region. This mosaicking process created an image of almost twice the original longitudinal variation, which will enable the viewer to construct geological maps of the area with more accuracy and convenience.
RESULTS

Previous work by Carr (1983), Ben-Avraham and Nur (1980), and Sillitoe (1974) had shown that major volcanic arcs could be divided into segments characterized by volcanoes of different heights and spacing. Latitudinal segmentation of the Central Andes has been directly observed from the volcanic trends in the color coded mosaic and indirectly supported by plotting various database parameters versus latitude. The approximate latitudinal boundaries of the proposed nine segments are shown in the figure below.

Segment one initiates the volcanic arc with dominately age five volcanoes, and is terminated to the south at the 1-2 segment boundary by the first active volcanism. This boundary may have been formed by a tear in the descending Nazca Plate, as proposed by Barazangi and Isacks (1979) using seismic data. Along with a change in relative age, segment two is further characterized by an increase in volcano basal diameter. The 2-3 segment boundary is dominated by a gap and offset to the east. This offset changes the strike of segment three's volcanic front, which is emphasized by a dramatic decrease in basal diameter. The 3-4 boundary transition, as well as the boundaries of 4-5 and 6-7, is relatively wide and terminated to the east by a large ignimbrite field. The relatively large basal diameter of the 3-4 boundary volcanoes, and a slight N-S change in strike of the volcanic front, help to emphasize segment four. A unique feature of segment five is
its high concentration of age two volcanism. This trend is abruptly terminated at the 5-6 boundary by a change to dominantly age one volcanoes with large basal diameters. Section seven is defined by a large transverse offset to the east and a decrease in the general basal diameter, while retaining a similar age trend as segment six. Furthermore, segment seven has peculiar lines of young volcanism parallel to its strike which fade until very thin (-35km) at the 7-8 transition. The thinness of lower segment seven is contrasted with a very wide 7-8 boundary (-200km), which is retained throughout much of the relatively older segment eight. The southernmost segment shows a change to more recent volcanism which tapers off around 28°40’s. The cross strike age trends of segments 1,2,3,6,7, and possibly 5 show ageing to the east, while the opposite is true of segment nine and possibly four. Segment eight shows no apparent age trend.

Changes in summit and edifice height are also related to segment boundaries. The general summit height of the volcanic arc varies only slightly, with a maximum in segment two, dropping to a minimum in segment four, and roughly increasing to a maximum in segment nine. Maximum summit heights at the arc ends, and a minimum roughly in the mid section corresponds to the pattern of volcano summit heights in the Central American volcanic arc (Carr, 1980). With the exception of boundaries 3-4 and 4-5, changes in mean edifice height (MEH) tend to emphasize segmentation. For example, segment two has a MEH of approximately 1538m, while segments three, four, and five are significantly lower at 1178m, 1258m, and 1190m respectively. The MEH of segment six is approximately 1612m, which drops to 1094m in segment seven, then increases in segment eight to 1556m. The MEH of segment nine (1652m) is not significantly different than that of segment eight, but the two segments are separated by a MEH of 995m between 25°45’s and 26°30’s latitude. These trends in summit and edifice height are opposite to those proposed by Z. Ben-Avraham and A. Nur (1980) for continental volcanism.

An attempt was made to investigate the variation of volcano spacing with latitude. However, a problem emerged in the distance measurements between individual volcanoes and members of composite volcanic complexes, e.g. linear ridges or clusters of related volcanoes. Even though this problem occurred, a good approximation can be made from direct observation of the MSS photo mosaic. It appears that the spacing of volcanic vents increases at segment boundaries.

CONCLUSIONS

The geomorphic characteristics of the Central Andes volcanic arc may be controlled by the subducting Nazca plate. A composite cross section of seismic activity for the 1-2 segment boundary clearly defines a break in the Nazca plate by contrasting the flat Benioff zone to the northwest to the more steeply dipping (25-30 degree) Benioff zone to the southeast. The large variation at this boundary may be explained by the subduction of the Nazca ridge, which is speculated to be a dormant transform fault and therefore a zone of weakness between the more buoyant, younger (by ~5my) oceanic crust to the north, and the more dense, older oceanic crust to the south. The shallow subduction under segment one may cause the lack of recent volcanism in this segment by not allowing the subducting slab to be surrounded by hot asthenosphere (M. Barazangi and B. Isacks, 1976). The absolute convergence rate between the South America and Nazca plates varies between 9.0 cm/yr at 17°s to 9.3 cm/yr at 30°s, and back to 9.0 cm/yr at 40°-20°s latitude (Minster and Jordan, 1978). This high rate of convergence,
combined with the trench's distinct bend, may contribute to the initiation of faults within the subducting plate, as similarly proposed by G. Roger (1985) using the Juan de Fuca plate system. These fractures may result in varying subduction rates of individual segments, and therefore changes in strike, age trend, and geomorphology of the volcanic arc. Seismic hypocenter cross sections of segment nine show a slight leveling in the Benioff zone, which possibly explains the greater width of this segment (D. James, 1971; M. Barazangi and B. Isacks, 1976). Furthermore, the latitudinal variations of earthquake epicenters, located less than or equal to 60km deep, show gaps in activity at all segment boundaries, excluding 6-7 (Chin and Isacks, 1983). Although the nine segments have been defined using geomorphological, geophysical, and geochronological data, ancillary data from geochemical studies would also be helpful in clarifying the nature of segment boundaries.

This research has demonstrated the usefulness of the relatively lower resolution MSS images for locating important volcanoes and defining major volcanic trends. A data base was generated using the characteristics of individual volcanoes from approximately 14°s to 29°s latitude. The resulting data base proved helpful in characterizing each proposed volcanic segment and in preparing the basis for more detailed petrologic and tectonic studies.

REFERENCES
LOCALIZED MELTING IN THE SHERGOTTY METEORITE: IMPLICATIONS FOR SHERGOTTY'S SHOCK HISTORY; Debra Colodner, Summer Intern, Lunar and Planetary Institute

Introduction

The Shergotty meteorite is one of a group of achondrites that has been dubbed the SNC meteorites, for Shergottites, Nakhlites and Chassignites. These meteorites differ substantially from other achondrites in crystalization age, chemistry, petrology and physical and magnetic properties. Much debate ensues over the origin of the SNC's and one suggestion is that they are pieces of Mars. Their young crystalization age of 1.3 by (4.5 by for other achondrites), evidence for a complex magmatic history, including crystal accumulation in a gravity field, the possibility of garnet in the source region, oxygen isotope compositions, and the presence of trapped gases similar to the Martian atmosphere are some of the SNC characteristics consistent with a Martian origin [1,2]. The dynamics of ejecting a piece large enough to form any of the SNC's without destroying their igneous textures are not well understood however, and this has led some investigators to consider other explanations. Vickery and Melosh (1983) argue that it is possible that the SNC meteorites crystalized in the central, deep portions of an impact crater on a large asteroid. The resolution of this genetic question depends in part on developing a better knowledge of the meteorites' shock history.

The results of isotopic studies done on the SNC's are not well understood. Shih, et al (1982), report a "best estimate" upper limit for the crystalization age at 1340±60 my, based on Sm-Nd whole rock data. Their "best estimate" lower limit is 1050 my, from 39Ar-40Ar dating of the low-temperature (feldspar) phase. Other ages reported are 650 my [5] from Sm-Nd and 39Ar-40Ar, and 637±20 my [6] from Ar-Ar. There is also disagreement on the age of the shock event. Rb-Sr ages reported range from 165 my [5] to 180 my [4]. However, Ar-Ar and Pb-U ages place the shock event at 250-260 my and 190 my ago, respectively [4,7]. In order to determine the shock age more precisely, it is important to understand the resetting mechanisms of these isotopic systems. Small, localized melt areas which occur in Shergotty may provide part of the answer. If these melt areas are impact-generated, they will be concentrated near iron-oxides and the mesostasis due to shock reverberations at density discontinuities between these phases and the less dense pyroxene and plagioclase around them. Since the mesostasis is a site of important radiogenic isotopes (in particular, Rb and Sr), concentration of stress and temperature there may have disturbed these isotopic systems. This study looks at melt regions in Shergotty and addresses two major questions: 1) What is their origin - are they igneous liquids, samples of fusion crust, or shock melts? 2) What are some of the implications for isotopic studies - to what extent are major isotope reservoirs (mesostasis, oxides, phosphates) involved in melting?
Results

Petrographic and microprobe analyses of ten melt areas in Shergotty show that they are a mixture of pyroxene, plagioclase, Ti-rich magnetite, whitlockite ($\text{Ca}_3(\text{PO}_4)_2$), fayalite, pyrrhotite and mesostasis. Melt compositions are variable, both within and between melt areas. For example, spots can have 2-15 wt% MgO, 2-35 wt% FeO, 2-25% Al$_2$O$_3$, 0-15 wt% P$_2$O$_5$ and 0-4 wt% TiO$_2$.

Analyses done along straight line traverses across melt areas show irregularly spaced highs and lows of both major and minor elements, demonstrating that melting did not completely homogenize mineral components (Fig. 1). This heterogeneity, and variation in the ratio of pyroxene/plagioclase components in the melt also proves that these melts are not igneous in origin (Fig. 2).

The possibility that these melts are actually injected samples of fusion crust cannot be ruled out as easily. Three sections of Shergotty were analyzed, one with a sample of the actual fusion crust, one from an area near the fusion crust, and one from an interior region of the meteorite. On the average, the fusion crust and melts near the fusion crust have lower Al (plagioclase component) than melts from the interior section (4.46 to 10.74, Fig. 3). Individual areas within the interior section, however, have compositions very similar to that of the fusion crust. Since fusion crust should not preferentially include oxides, mesostasis, or associated minerals, like phosphates, one might expect to see much less P and Ti in the fusion crust than in the interior melts, if they are impact melts. This is true for P (.74 in fusion crust to 2.49 in melts) as an average, but it is not true for Ti (.77 in fusion crust to .74 in melts). Until a more extensive sampling of the fusion crust is done, I can only say that the interior melts seem to include slightly more plagioclase and whitlockite than the fusion crust. Chemical differences do not rule out the possibility that the interior melts are injected fusion crust, but there is no petrographic evidence for this, such as long stringers of melt that would connect these areas to the surface. This, of course might be due to the angle at which the section was cut, but one might expect to see some connection to fusion crust out of the ten areas studied.

Melts occur both as short (<700 micron) veins through host pyroxenes and more commonly, as rounded areas (<600 micron diameter) along phase boundaries. Their texture generally consists of small iron-sulfide spherules [6] dispersed in the heterogeneous silicate melt. Some of the melt areas are themselves extensively fractured, while fractures in host minerals end at the margins of other melts. In a single melt area, some fractures may end at the melt edge, while others cut it. This may be evidence for more than one shock event, as proposed by Lambert (1985), or for different stages of a single event. This observation alone is not sufficient to resolve this question.

Microprobe analyses indicate a large degree of chemical
involvement of the mesostasis in melting. Compared to pyroxene and plagioclase, which constitute over 90% of Shergotty, the melts are all enriched in the components of the mesostasis and the late-stage crystallization products such as whitlockite and magnetite. These components include K, Ti, P, Fe (Fig. 4). As discussed before, this is most likely due to density discontinuities at grain boundaries with high-iron phases. Whitlockite is included in the melt, probably because of its close association with these high-density materials.

Conclusions

On the basis of the heterogeneity of the melt areas, it is clear that they are not igneous in origin. Because of compositional differences (less incorporation of plagioclase and whitlockite in fusion crust), and lack of petrographic evidence to support fusion crust origin of interior melts, I have concluded that these melts are impact-generated. Perhaps there are other ways to differentiate between these two possible origins, and these should be explored in the future.

The role of impact melting in the interpretation of Shergotty's shock history may be most significant in relation to the resetting of the Rb-Sr isochron, 165 my ago [5]. The coincidence of concentration of these trace elements in the mesostasis, and the concentration of shock-induced heat in and around the mesostasis may have initiated reactions between the mesostasis/melts and their host minerals. Sr may have been particularly susceptible to diffusive exchange between melt and host. This would influence the calculated age of re-equilibration of the Rb-Sr system and thereby, the proposed age of the shock event. Preliminary traverses with the microprobe across pyroxene rims adjacent to melt areas did not reveal any exchange of Fe, Mg or other elements across the pyroxene-melt boundary. Future studies should include more careful studies of possible reactions between melt and host minerals.

References

VARIATION IN MELT COMPOSITION

Fig. 1

SHERTOTT MELT VARIATION

Fig. 2

SHERGOTT MELT VARIATION

INTERIOR VS. FUSION CRUST

Fig. 3

PYROXENE VS. MELT COMPOSITION

Fig. 4
ASHFLOW CALDERA FIELDS: A SYNTHESIS AND INTERPRETATION

Claire D. Duncan, 1985 Summer Intern, Lunar and Planetary Institute, Houston, TX

Ashflow calderas, the largest calderas on earth, are associated with voluminous (100-1000 km³) eruptions of ignimbrites, typically of dacitic to rhyolitic compositions. (1,2) Previous workers have mapped and described various aspects of individual ashflow caldera fields (ACF) and their associated calderas including; topographic expression and dimension, estimation of ashflow/field volumes, ignimbrite geochemistry and composition, resurgence, relative and radiometric ages and interpretation of pluton/batholith properties. (3,4,5,6) This preliminary study attempts to synthesize existing ACF data dispersed throughout published literature to study the mechanics of ACF as a discrete type of volcanism.

Twelve ACF have been researched and at least eleven of these have yielded sufficient descriptive information to include them in this study. These fields are: Jemez Mountains, New Mexico; San Juan, Colorado; Marysvale, Utah; Long Valley/Mono Basin, California; Yellowstone Plateau, Wyoming; Mogollon-Datil, New Mexico; Timber Mountain/Oasis Valley, Nevada; McDermitt, southeast Oregon/north central Nevada; Kane Springs Wash, Nevada; Tibesti, Chad and the Younger Granite Province, Nigeria. For each of these fields, the following characteristics have been compiled: field area and volume, caldera number and diameters, average distance from a caldera to its nearest neighbor (NND) and total age range (lifetime) of the fields (Table I).

Subsequent to tabulation, several of these characteristics were graphed to observe any existing trends in ashflow field properties. A number of the plots failed to yield satisfactory results. However, positive graphic relationships observed for others substantiated some general assumptions concerning ACF. For example, a graph of the number of calderas versus field volume (km³) indicates that caldera number increases with field volume (fig. 1). Caldera diameter (km) plotted against ashflow volumes (km³) displays that with increasing caldera diameter, the ashflow increases as well (fig. 2).

Manipulation of ACF parameters by computing ratios and averages provides additional information and yields clearer graphic relationships. Eruption rate, total field volume per million years, graphed versus field volume is one example. This graph (fig. 3) suggests that the greater the total volume of an ACF, the greater the eruption rate. Another plot of average caldera diameter against field volume clearly demonstrates that
<table>
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<th>VOLUME (CU KM)</th>
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<th>MND (KM)</th>
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**FIGURE 1**

![Figure 1](image_url)

**FIGURE 2**

![Figure 2](image_url)
regardless of field volume, all calderas have approximately equal diameters (fig. 4). Since field volumes reflect pluton/batholith volumes, this graph suggests that the batholith cupolas, the magma chambers that the calderas collapse into, are also approximately equal in size. An additional graph of number of calderas versus field lifetime yields many points which may be approximated to one of four lines (fig. 5). These lines represent caldera formation per unit time. This graph may also be interpreted as indicating the period of time required to establish or replenish magma cupolas/chambers. Intervals between caldera formation range from about .17 to 2 m.y., with .7 m.y. the most common period.

It is interesting to note that for many of the graphs, Yellowstone Plateau ACF parameters such as field/ashflow volumes and caldera diameters greatly exceed those of other ashflow fields. As a result, Yellowstone often differs substantially from the general trend of the graphs defined by these parameters (see fig. 4). In addition, the Yellowstone caldera has two separate resurgent centers. Based on this evidence, we suggest that previous workers may have failed to recognize that Yellowstone caldera and an adjacent, unnamed one actually represent multiple rather than single calderas.

This study has yielded some additional ideas about the properties and systematics of ashflow caldera volcanism and the batholithic sources associated with them. Data from additional ACFs must be compiled to further constrain preliminary relationships. Subsequent research should also compile crustal information such as structure and composition for each field to establish any correlations with other ACF parameters.

REFERENCES
Examination of stratospheric particles is important because they have proven to be an important source of extraterrestrial material, providing samples of comets, the most pristine components of our solar system [1-5]. Much thought has been given as to what should be found in the stratosphere, and examination of the particles provides data as to what actually is found there. Stratospheric dust particles have a relatively short lifetime (<2 years), so there must be a continual influx of this particulate matter [5]. There are several possible origins for the particles. Some particles are extraterrestrial, being derived from comets, asteroids or the interstellar medium. Some particles are terrestrial, forced into the stratosphere by volcanic eruptions, violent dust storms or nuclear testing. Also present are artificial contaminants, from rockets and satellites, or chemical contaminants from processing.

Two previous examinations of collection surfaces (from flags U2-9 in 1976 and W7017 in 1981) involved filter-processed particles, rather than particles that were picked from the collection surfaces by hand [5]. This study of a third flat-plate particle collection surface (U2024) also involves particles processed by means of a filtering system. A third examination has been undertaken in order to complement and provide comparison to the first two studies.

The particles on the collection surface U2024 were collected on silicone oil-coated, flat-plate dust collectors that were mounted on pylons on the underside of the wings of a NASA-U2 plane. The particles were collected from the lower stratosphere over the North Atlantic area in 1984, at an altitude of approximately 20 kilometers. The duration of the flight was approximately 14 hours. The particles were washed free from the collection surface by successive rinses of freon, through a series of sequentially stacked filters, 5 µm, 1 µm and .4 µm in size. A final rinse with hexane dislodged any remaining particles [5&6].

Two size fractions were separated, >1 µm and >5 µm. The two Nucleopore filter surfaces were examined using the JEOL 35CF Scanning Electron Microscope, where particle size, shape and chemistry were noted. Bulk elemental analyses were obtained using the PGT EDS attachment to the SEM. One quadrant of the >5 µm filter surface was examined, where all particles (533) were analyzed. Approximately 1% of the >1 µm filter surface was examined, the lower percentage being due to the much higher concentration of particles. To date, 310 particles on the >1 µm filter surface have been analyzed.

The particles are classified according to a system devised at the Johnson Space Center [667]. The system is based on major element chemistry and gross particle morphology, with classes including chondritic, silicate, aluminum (consisting solely of aluminum or aluminum oxide), aluminum-prime (predominantly aluminum with lesser amounts of other elements), iron, iron plus sulfur, calcium aluminum silicate, low-Z and "other". Two tables for flag U2024 are given; the first accounts for all particles and the second has sulfur-rich particles deleted. It is believed that the sulfur-rich particles formed by the oxidation of sulfuric acid droplets during the filtering process. Tables are given for flags U2-9 and W7017 for comparison. The total stratospheric number...
density of particles larger than 1 μm in diameter from the U2024 flag is calculated to be 1.1 particles/m³ air. This figure is not significantly changed after the deletion of the sulfur-rich particles. Compared with the past two studies (U2-9 = 0.58 x 10⁻¹ particles/m³ air for 1976, and W7017 = 1.0 x 10⁻¹ particles/m³ air for 1981), flag U2024 reflects a much higher particle density [5]. This difference may be due to randomly distributed particle populations, or may be due to a greater influx of particles at the time of collection of flag U2024 in 1984.

From Table 2, it can be seen that 51% of the particles are aluminum-rich, with a substantial amount of silicate, low-z and "other" particles. The majority of the aluminum-rich particles are believed to be rocket exhaust and satellite ablation debris. With increasing grain size, there is an increase in the relative abundance of aluminum-prime and "other" particles and a concurrent decrease for silicate and low-Z particles. In comparison with the two past studies, there appears to be a decrease in the relative abundance of chondritic and silicate particles, and an increase for aluminum-rich particles. The increase in the relative abundance of aluminum-rich particles may reflect an increase in man-made debris or may just reflect a non-uniform stratospheric distribution of particles.

References:
Table 1  Stratospheric dust surface U2024 (1984)

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Table 2  Stratospheric dust surface U2024, after deletion of sulfur-rich material

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Table 3  Stratospheric dust surface W7017 (1981), after deletion of volcanic material

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<td>11</td>
<td>13</td>
</tr>
</tbody>
</table>

Table 4  Stratospheric dust surface U2-9 (1976)

<table>
<thead>
<tr>
<th>size</th>
<th>chondritic</th>
<th>silicate</th>
<th>Al</th>
<th>Al'</th>
<th>Fe-S</th>
<th>Fe+S</th>
<th>CAS</th>
<th>low-Z</th>
<th>other</th>
</tr>
</thead>
<tbody>
<tr>
<td>&gt;10um</td>
<td>17%</td>
<td>24</td>
<td>5</td>
<td>7</td>
<td>10</td>
<td>2</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>35</td>
</tr>
<tr>
<td>6-10um</td>
<td>&lt;1%</td>
<td>34</td>
<td>24</td>
<td>10</td>
<td>10</td>
<td>3</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>19</td>
</tr>
<tr>
<td>1-5um</td>
<td>12%</td>
<td>34</td>
<td>5</td>
<td>20</td>
<td>8</td>
<td>6</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>15</td>
</tr>
<tr>
<td>entire</td>
<td>12%</td>
<td>34</td>
<td>5</td>
<td>20</td>
<td>8</td>
<td>6</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>15</td>
</tr>
</tbody>
</table>
Charnockites are pyroxene-bearing anhydrous granitic rocks which are common in granulite terranes. They are believed by some\(^2,3\) to have formed in response to the influx of CO\(_2\)-rich fluids through the rocks during granulite-facies metamorphism. The common presence of CO\(_2\)-rich fluid inclusions trapped in the quartz grains of charnockites and granulites supports this idea\(^1\). Since granulites typically form at depths of 25-30 km, this implies that carbonic fluids are also present at these depths.

This research concerned the study of some unmetamorphosed anhydrous pyroxene-bearing granites in an attempt to discover whether CO\(_2\) played any role in their formation. These granites are from the 1.4 Ga Mistastin Lake Batholith, a granitic complex located in the Nain Province of Labrador. Similar rocks occur throughout this province, where they are associated with major unmetamorphosed anorthosite massifs. The presence of CO\(_2\) in the rocks under study would indicate that a carbonic fluid may have in fact had a direct role in their igneous petrogenesis, perhaps providing an explanation for their anhydrous mineralogy.

Pure CO\(_2\) fluid inclusions in the samples were indeed found and observed using fluid inclusion microthermometry and quadrupole mass spectrometry. Melting and homogenization temperature data were obtained for three samples; microthermometry data for one more sample exists from last year's work. Melting temperatures of the inclusions ranged from about -54°C to -57.5°C; homogenization temperatures generally fell between +22°C and +30°C, with a few slightly lower or higher. This is consistent with the behavior of pure CO\(_2\) inclusions with a density of 0.75-0.80 gm/cm\(^3\). Of 155 inclusions for which homogenization temperatures were measured, about 30% were in planes (fractures), 15-20% were in trails, about 15% were in groups, and 15-20% were apparently isolated. About 20% could not be classified. Of 130 inclusions for which melting temperatures were measured, roughly 15% were in planes, 30-35% were in trails, 25-30% were in...
CO₂ IN ANHYDROUS GRANITES
Rachel Lerner

(cont.) groups, 5-10% were isolated, and about 15% were unclassifiable.

Mass spectrometry was performed on seven specimens of four different samples--three samples were run twice. The results showed H₂O to be the most abundant constituent of the rock, with lesser amounts of CO₂ and CH₄. The mole ratio CO₂/H₂O was on the order of equal to or less than 10⁻³, with one sample showing a significantly large ratio (about 0.2). The mole ratio CH₄/H₂O varied between 10⁻² and 10⁻⁵. Two samples contained too little CO₂ to construct a ratio; one of these two showed a similar situation with CH₄.

Results obtained from the research generally support the hypothesis that CO₂ was present during the formation of the Mistastin granite, though the data is too limited to be conclusive. Up to 20% of the inclusions studied are isolated; isolated inclusions are more likely to be primary. The study also shows a correlation between the amount of pyroxene in the samples and the number of CO₂ inclusions seen; the greater the pyroxene content, the greater the number of CO₂ inclusions in the sample. This would be expected if a carbonic fluid was in fact present during the granites' genesis, driving off the water to form an anhydrous mineralogy. This trend is strongly supported by work done by Madsen(4) on a Norwegian granitic intrusion. Furthermore, the two samples examined under thin section showed not only abundant pyroxene but also texturally igneous (interstitial) carbonate (siderite?); this represents further evidence that carbonic fluids were present in the granitic melt at the time of its crystallization. The presence of CH₄ in the samples and its relationship (if any) to the CO₂ is as yet unexplained and would require further study.

Many thanks are owed to Dr. Mike Sommer and Eric Rehwoldt for running my samples through the mass spectrometer in Dr. Sommer's lab; to Dr. Mike Zolensky for technical aid and helpful remarks; and to my advisor Dr. Lew Ashwal for guidance, information, and advice.

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Rachel Lerner

(cont.) phism, granulites, and crustal growth"  Nature, V.288, pp.45-50


THERMAL INERTIAS FOR THE ELYSIUM AND AEOLIS QUADRANGLES OF MARS.

Laurie A. Leshin, 1985 Summer Intern, Lunar and Planetary Institute, Houston, Texas, 77058

Low-resolution infrared data from the Viking infrared thermal mapper (IRTM) have been used to produce a global map of Martian thermal inertias (1,2), however, little of the high-resolution data have been examined thus far. This work presents, for the first time, results and interpretations obtained from mapping high-resolution thermal infrared data in the Elysium and Aeolis quadrangles at equatorial latitudes on Mars. Comparing thermal inertias to surface features and geologic units leads to a better understanding of the geologic processes, particularly aeolian processes, which take place in this highly diverse region of the planet.

The two quadrangles studied are distinctive both thermally (Fig. 1) and geologically. Elysium (0° to 15° N, 180° to 225° W) contains the volcano Elysium Mons and the Cerberus low albedo feature as well as several craters with accompanying low albedo patches (3). Aeolis (15° S to 0°, 180° to 225° W) shows very distinctly the transition from the smooth plains of Elysium Planitia to the older, elevated, more heavily cratered areas of the southern hemisphere.

The thermal inertias mapped in Elysium and Aeolis range from a low of 1 (x 10^{-3} cal cm^{-2} sec^{-1/2} K^{-1}) to a high of 14. The lowest values were recorded at Elysium Mons and the highs correspond to dark patches in craters and channels in the south (e.g. 18° S, 197° W) which appear to have trapped the low albedo, high thermal inertia material. The range of thermal inertias determined from this work differs slightly from the global map for this area which has a low of 1.5 and peaks at about 10 (2). Overall, the high-resolution thermal inertia values and their correlation with albedo agree very well with the global trends. The larger range of values in the high-resolution data is due primarily to the diminishing effect of averaging data (global mapping determined an average thermal inertia value for 2° x 2° areas while the high-resolution data are averaged in 1/8° x 1/8° bins). Thus, more localized extreme thermal inertia features can be recognized.

Craters often appear to have collected aeolian material (4,5,6) and show markedly different inertia patterns than their surrounding terrain, and these patterns are clearly distinguishable in the high-resolution data. One 60 km diameter crater (located at 3° N, 198° W) has low albedo material on the floor of the crater and in an adjacent SW-oriented streak. Orbital track A648-2 crossed directly over the interior deposit and the thermal inertia values increased from 3 for the surrounding plains to greater than 9 for the crater interior. This crater represents one of the few features in the Elysium Planitia area that has not been completely covered by fine grained material (interpreted to be aeolian dust) which blankets these plains (1,7).

Directly north of this crater is the Cerberus low albedo region. This collection of dark material also has high thermal inertia values (~8 or 9) when compared with the Elysium Planitia region. This is consistent with the global correlation between low albedo and high thermal inertia (1). Two lower resolution orbits were mapped across Elysium Mons and the thermal inertias found (1-3 on and around the volcano) agree with the global data.

The two quadrangles have geologic units ranging from some of the oldest areas on the planet to much younger plains (8). The thermal signature of the younger units is quite distinct; for example, thermal
inertia values for the Elysium Planitia region average around 2-3. This corresponds to a surface covered with fine grained material (<.1mm; 1). Since the material overlays all geologic features and is thought to be aeolian in origin (1), the thermal inertia values are consistent with aeolian activity which post-dates all geologic terrains.

The thermal inertias obtained for the older geologic units are not as consistent. Most of the older units occur south of Elysium Planitia in a region that does not appear to be mantled by dust, as the younger plains are. Collections of dark material in this high crater density area (9) would lead to an expectation of a distinctly high thermal signature (1). Though there is a concentration of thermal inertias of 6.5-8 in this region, a peak of values also occurs at 3. It should be stressed that the higher thermal inertias are much more pronounced in the south than the lows. However, the fact that there is a considerable amount of low inertia material is significant. Again, aeolian dust must be blanketing portions of the surface, although not as completely as on the younger units. This material is probably a thinner covering than the young plains where the geologic units are virtually completely masked by the thick deposits (11).

The low albedo, high inertia material which has the most pronounced thermal signature in the older regions is also proposed to be aeolian in origin (10,12). The thermal inertia is indicative of course materials (>1mm; 1) and the dark deposits generally dominate the thermal signature in the older regions.

Although it may be expected that the geologic units dictate an area's thermal signature (1), one significant exception is noted in the studied area. The geologic unit described by Scott and Carr (8) as "knobby material" is interpreted to be older than the young plains material which surrounds it. Thus it might be expected to show high thermal inertia values like other comparitively older units of the southern hemisphere. However, this is not the case. Thermal inertias for this terrain were found to concentrate around 3, without a peak at a higher value as with the other older units. This indicates that its surface characteristics may be completely masked to the IRTM by the fine grained material which covers the surface.

Using Christensen's model (13) for determining areal block abundance (where 10 cm sized blocks are assumed to have a thermal inertia of 30), a comparison of thermal inertia to the calculated block abundance can be made. The abundances were mapped only for the subquadrangle of MC-23 NE (15°S to 0°, 180° to 202.5°W), where a wide variety of geologic terrains are present. The values were quite uniform, averaging about 5% block cover on most terrains. The only major deviations from this value occurred in some isolated high thermal inertia (-5) patches located at 12°S, 190°W where the block abundance dropped nearly to 0. This indicates that the high inertias observed there are due completely to the collection of dark sand-sized material and not a mixture of a wide variety of particle sizes. Thus, the high inertia material here would appear to be aeolian materials that have become size sorted through saltation and surface creep (12).

Between the thick dust deposits of Elysium Planitia and the dark, cratered terrain of the south, aeolian processes dominate the distribution of thermal inertias in this area of the planet. Mapping the rest of the equatorial regions of Mars should reveal even more about interaction between the wind and the surface materials.
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Figure 1. High-resolution thermal inertias for the Elysium and Aeolis quadrangles of Mars. Data correspond to the Viking orbits and represent the only place data are available.
SOLAR MAX; MICRO-MATERIAL ANALYSES
M.L. Lieurance; Lunar and Planetary Institute, Advisor: D.S. McKay (NASA/JSC)

Introduction:
The Solar Max satellite was launched on February 14, 1980. On April 10, 1984 Solar Max was repaired by a shuttle crew (STS 41-C). The thermal insulation pads brought back from the satellite show impact craters from high velocity material. The layering of the aluminum coated plastic material within the structure of the pads provides an excellent opportunity to analyze the structure, physics, and chemistry of high velocity impact material and craters. This paper concentrates on the elemental compositions of particles found associated with some of the craters, and the composition of particles thought to be unassociated with impact material (pre- and/or post impact contaminants).

Past and current work on the impact craters of the Solar Max shows two major types of particles: meteoritic, and man-made debris. Because of the potential destructiveness of high velocity particles, it is especially important to evaluate the percentage of man-made impacts versus meteoritic impacts. Results from the Solar Max can be compared to similar future projects to determine the magnitude and rate of debris accumulation in space. Many other valuable conclusions may eventually be extracted from further research on the Solar Max micro-impacts with results giving interest to groups ranging from the defense department to geologists.

Procedure:
Six impact sites and contaminants from eight samples were examined on the outward facing side with a JSM-35CF scanning electron microscope (SEM) equipped with a PGT System IV energy dispersive x-ray analyzer (EDXA). Particles analyzed ranged in size from .5 microns to 20 microns. A counting time of 100 seconds was used for the EDXA analyses. A beryllium window was used on the detector which screens out energy spectrums less than about 1 kilo-electron volt (elements with Z values less than sodium were not detected). Particles were randomly chosen when the population of the area investigated was larger than about 300, otherwise all particles were examined. Particle spectrums which only contained low Z elements (less than Na) were not recorded. A few procedures of averaging particle compositions were explored. For some particles and particle groups, background subtraction was used to enhance element detection. Photographs of the SEM images were taken to record impact structure, particle orientation and density, and anomalous particles (anomalous is based on the degree of deviation from typical composition and/or shape). Contamination analysis began by taking samples from the thermal pads which had no craters on or near them (blanks). Blanks of the first layer are of two types, exposed and unexposed. The unexposed blanks were shielded from exposure to space. One exposed blank, one unexposed blank, one partially exposed/unexposed blank, and particles on sample 4-1-6-300-1 (contains 3 craters) that were
greater than 500 microns from the impact sites, were used in the first layer investigation. Four blanks were analyzed from the second layer.

Results:
Results from the first layer contamination investigation indicate that many of the debris particles previously thought to be impact related may be explained as contamination. Specifically, the paint type particles (Si, Ti, Zn, Cl and K) (Kessler and others, 1985) were observed as occasional contaminants. Approximately one particle per 1,000,000 square microns is a conservative estimate of the average density of paint particle contamination on the exposed portions of the first layer (estimate made from an immature data base). This density was observed to vary locally from no presence of paint type particles, to an increase of about 100 times that of the estimated overall average. Titanium-rich particles were common to the exposed blanks but not the unexposed. Although usually found with just silicon, titanium-rich particles often had minor amounts of aluminum, zinc, chlorine, and/or potassium. The titanium-rich particles may have the same origin as the paint type particles; an origin which is not known with any certainty at this time.

Particles which have compositions similar to urine (K, Na, Cl, P, and/or S) were found both on the unexposed and the exposed first layer contamination samples. Some of these particles may be perspiration residue left from human handling of the sample. Other particles of urine composition, which were found on the exposed blanks, may have originated as waste in space, but did not impact the thermal pad with enough velocity to penetrate the first plastic (mylar) layer of the thermal pads. The first layer of mylar is 75 microns in thickness before exposure to the atomic oxygen present in orbit.

Silicon-rich and calcium-rich particles were frequently encountered on all the blanks. The silicon-rich material may have originated as an outer coating. Scratches in the mylar, possibly from the application of silicon oil, can often be seen near the denser deposits of silicon particles. The calcium may be from the composition of the mylar (type of plastic that the layers of the thermal pads are made of). Iron particles were found on all blanks, usually directly associated with the dag (carbon material used to prevent electrical charging of the sample). Silver-chlorine particles were observed on an unexposed first layer blank, probably having a photographic origin and being deposited by handling of the sample.

The overall composition of the contaminants on the exposed samples by particle averaging is silicon, chlorine, and potassium as major constituents, and titanium, aluminum, calcium, and iron as minor. For the unexposed first layer samples the overall contaminant composition is silicon, chlorine, and calcium as major constituents, and the minor elements are aluminum, sulfur, and silver. Also in comparing the two types of first layer samples, it should be noted that the particle density (high Z and low Z) for the exposed blanks is about 25 particles per 1,000,000 square microns and about 5 for the unexposed.
For the blanks of the second layer, the particle density averaged about two particles per 1,000,000 square microns. The second layer EDXA analyses show major amounts of aluminum, and minor amounts of silicon, phosphorous and calcium. These elements are explainable as being part of the composition of the mylar and its coating. Therefore, it is thought that the second layer has only trace amounts of contamination that is not related to the thermal pad composition. It is important to consider though, that pre-impact contaminants from the first layer could be carried through to the second layer during the impact.

Of the six impacts analyzed, three were conclusive as to the composition of the impacting material. The most probable impact composition for sample 4-1-6-300-(1,2) (130 micron crater) is sulfur with minor chlorine. This is proven best in the EXDA analysis of the second layer particle and vapor deposit thought to be created during the impact. The apparent composition of sample MEB-1-6-300-(1,2) (20 micron crater) is potassium and phosphorus with trace chlorine. Sample MEB-3-2-0-1 is a blank (no craters), but has a ring feature about 100 microns in inside diameter that is made of crystalized potassium and chlorine. Inside the ring are particles with compositions of potassium, sulfur and phosphorus with minor and trace amounts of sodium and chlorine. Bladed crystals of potassium and phosphorus were found inside and out of the ring feature for a radius of about 400 microns. The overall composition (in decreasing abundance) appears to be potassium, chlorine, phosphorus, and sulfur with trace sodium. This composition and impact site might be expained by low velocity (no penetration), human waste material originating from the shuttle (McKay, 1985). No explanations are offered for the other impacts that were analyzed at this time including the inconclusive impacts. Sample 4-1-6-300-1 (300 micron crater) is one of these inconclusive impacts, but it should be noted that a strong counting bismuth particle was found on the crater lip and is apparently impact related. Bismuth has not yet been observed to be a contaminant. The other inconclusive impacts were samples MEB-1-23-280-(1,2) and MEB-1-4-300-(1,2).

Conclusion:
It is apparent that contamination can interfer with impact material analysis. It is not yet clear to what degree. More research is needed on contaminant compositions and populations. It could be argued at this time that the paint type particles are not truly contaminants in the perspective that they may have been deposited on the thermal pads while in orbit. They may have settled on the thermal pads at a relatively low velocity, similar to the mechanism that may be allowing the waste particles to be deposited. Even so, there is evidence to show that paint and waste type particles can be found with no association to craters on the exposed portions of the first layer. A large data base of contaminant particles will give future crater analysts a background to determine if a crater has enough suspected impact material to be anomalous. Since contamination was seen to vary locally, it is suggested that particle analyses be done away from the crater site to establish
a local contamination background. A procedure of averaging compositions of particles found directly associated with the impact and averaging particles that are local to the sample but unrelated to the impact (background), along with subtraction of these two averages, may be helpful in determining the composition of the impact material. In general, the Si-rich, Al-rich, Ca-rich and the calcium-phosphorus particles should be assumed contaminants. To avoid contamination from lab procedures, such as the iron-rich, sodium-chlorine and silver-chlorine particles, great care should be taken in the handling and storage of the samples.

References:


LAKE BONNEVILLE: A MODEL OF ISOSTATIC ADJUSTMENT; Glenn M. May, Bruce G. Bills, Lunar and Planetary Institute, 3303 NASA Rd 1, Houston, TX.

Lake Bonneville, a Late Pleistocene lake, was the largest lake which occupied the Great Basin of Utah, Idaho, and Nevada of the Basin and Range Province. During its maximum extent, from around 15,000 to 14,000 years ago, this lake covered an area greater than 50,000 km² and had a maximum water depth of 340 meters.

This lake level was followed by a catastrophic decrease in water level around 14,000 years ago during a collapse of unconsolidated material near Red Rock Pass in the northeast corner of the lake. Lake level rapidly lowered about 96 meters stabilizing at the Provo shoreline at the 1,445 meter level. Lake Provo at its maximum extent covered an area greater than 38,000 km² and had a maximum water depth of 213 meters.

Lake level remained at this elevation until 13,500 years ago. At this time water level began to decline, a decline which lasted for some 2,000 to 3,000 years. Between 11,000 and 10,000 years ago lake level again rose forming the Gilbert level at 1,295 meters. This lake at its maximum extent covered an area of roughly 17,000 km². The maximum duration of this lake was only some 500 to 1,000 years and marked the end of the Lake Bonneville cycle and the beginning of its successor, the Great Salt Lake cycle.

This region has been examined through the use of remote sensing techniques; the primary objective to look for discernable traces of Great Basin lakes that occupied this region. The most recognizable shorelines as determined through field work are the Bonneville level, the Provo level, and the Gilbert level.

Remote sensing was executed using Heat Capacity Mapping Mission (HCMM) images sampled at 500 meters resolution and Landsat Multi Spectral Scanner (MSS) images sampled at 100 meters resolution. The HCMM images were geometrically corrected using the Universal Transverse Mercator grid system. A false color composite was produced using day visible, day infrared, and night infrared as red, green, and blue respectively. This color composite demarked a region very nearly coincident with the Gilbert shoreline. The Bonneville and Provo shorelines were not as easily discernible.

Six Landsat MSS images were mosaicked to cover the entire Lake Bonneville region. Each image was first prepared by correcting missing lines, destriped to remove scan lines, and geometrically corrected to the Universal Transverse Mercator grid system. This process was performed on all 4 bands of each image. The six images were then mosaicked and a false color composite was produced using band 7 (0.8-1.1 micrometers), band 5 (0.6-0.7 micrometers), and band 4 (0.5-0.6 micrometers) as red, green, and blue respectively. Like the HCMM image, the Gilbert shoreline was the most discernible of the three shorelines. It was determined that remote sensing techniques can aid in the mapping of Lake Bonneville shorelines in this region.

Contoured deflection plots of the Bonneville, Provo, and Gilbert shorelines were produced at a scale of 1:750,000 to coincide with a shoreline map already in print. Both the Bonneville and Provo deflections were plotted with a contour interval of 10 meters; the Gilbert deflection being plotted with a 5 meter contour interval. These plots show that isostatic rebound in this area produced by the removal of previous water loads was greatest near Lakeside and Grassy Mountains where the water depth was greatest. Isostatic rebound of the Bonneville shoreline has been measured at around 70 meters, the Provo shoreline circa 60 meters, and the Gilbert shoreline circa 20 meters. An area surrounding Pavant Butte in the southern region of the lake
basin shows less isostatic rebound than immediately surrounding areas suggesting that the load imparted by these volcanics added significantly to the crustal load thereby reducing the amount of isostatic rebound of the Bonneville shoreline.

The first real attempt at a geophysical model simulating this rebound was made by Crittenden (1963). He made a thorough re-evaluation of the Bonneville shoreline; measuring both location and elevation of 90 shoreline features. This report showed that the isostatic uplift in this region was slightly more than 200 feet (61 meters).

Crittenden's model assumed an elastic crust 50 km in thickness of mean density 2.80 floating on highly viscous but fluid substratum of mean density 3.25. It was determined that the relaxation time ($T_r$) of the Bonneville shoreline was between 4,000 and 10,000 years. In 1967, Crittenden re-evaluated this relaxation time by placing the previously dated Lake Bonneville (57,500 years B.P.) at 16,000 years B.P. The $T_r$ in this later model was less than 4,000 years.

From this relaxation time the viscosity for a viscous half space can be calculated using the expression:

$$ N = \frac{DgT_r}{2f} $$

where $N$ is the viscosity in poises, $D$ the density of the substratum, $g$ the gravitational constant, $T_r$ the relaxation time in seconds, and $f = (1/m/(l^2 + m^2)^{1/2})$, where $l$ and $m$ are the horizontal dimensions of the load in centimeters. The viscosity of the substratum was calculated to be $0.96 \times 10^{21}$ poises (9.6x10^21 Pa s).

Passey (1981) modeled the Bonneville load as two cylinders; one cylinder centered at 41°N and 113°W with a radius of 100 km and a depth of 310 meters and a smaller cylinder centered at 39°20' N and 112°40' W with a radius of 60 km and a depth of 120 meters. With this model the flexural rigidity of the lithosphere was calculated as $1.0 \times 10^{23}$ N m suggesting a lithospheric thickness of 25 km.

Passey also concluded that the response to unloading between the Bonneville and Provo shorelines was at least 91% complete by the time the Provo shoreline formed. Using a time interval of 500 to 3,500 years as the time between these lake levels, the relaxation time can be calculated in the range 208 to 1,450 years; a value much less than that calculated by Crittenden. Using these constraints Passey calculated the viscosity of the substratum as $5.0 \times 10^{21}$ to $3.5 \times 10^{20}$ Pa s, the best estimate being $2.0 \times 10^{20}$ Pa s.

Nakiboglu and Lambeck (1983) constructed a model consisting of an elastic layer overlying a viscoelastic channel bounded below by a high viscosity rigid medium. The Lake Bonneville load was represented by 108 to 192 disks of radius 38.25 km situated at 12 different localities. This model was run using the load histories of Crittenden and Scott. The mechanical parameters affecting the deformation are the rigidities, thicknesses, and Poisson’s ratio of the upper two layers, and the viscosity of the viscoelastic layer. The inverse Laplace transform of the vertical deformation after the loading history is applied determines the surface deformation. Nakiboglu and Lambeck concluded that the thickness of the elastic layer is fixed at 28 to 30 km regardless of whether or not a channel flow exists in the mantle. The range of flexural rigidity of the lithosphere was determined to be $1.8 \times 10^{23}$ N m to $2.1 \times 10^{23}$ N m. The effective viscosity of the upper mantle was in the range $1.5 \times 10^{19}$ Pa s to $3.4 \times 10^{19}$ Pa s depending on the load history.

The model presented in this study is a modification of the Nakiboglu and Lambeck model. Due to differences in load history interpretations, three
load histories were used: Scott et. al. (1983), Currey (1983), and Spencer et. al. (1984). The initial load was computed by subtracting USGS topographic elevations from the lake elevation at any given time. An initial deflection was calculated at each time interval by calculating a fourier transform of the load, multiplying the transform by its respective hankle filter value, and then calculating the inverse fourier transform; a correction of the water load was then computed by adding this deflection value to the initial load.

After computing the initial load correction, the above sequence of calculations was applied to the load history to calculate the amount of isostatic rebound at the Bonneville, Provo, Gilbert, and present lake levels at localities of known shoreline elevations. A north-south profile of the calculated rebound was compared with a north-south profile of presently observed elevations. The residual deflection is calculated as the difference between the calculated and observed values; the residual approaching zero as the calculated elevations more closely approximate the observed elevations. This fit was checked by plotting these residuals with a contour interval of two meters, a blank plot occurring if the residual values are within two meters of the observed values. Using a deep viscoelastic channel the best fit has been obtained using a crustal thickness of 30 km and a viscosity of the substratum at $3.5 \times 10^{20} \text{ Pa s}$. The best fit for a 200 km thick viscoelastic channel has been obtained using a crustal thickness of 30 km and viscosity of $8.1 \times 10^{20} \text{ Pa s}$ for this channel.

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Fluid inclusions resulting from growth irregularities of a solid crystal forming from a fluid medium are almost ubiquitous in geologic samples. The study of fluid inclusions has been applied to many geological problems and areas ranging from the study of ore deposits, to the evaluation of safety of sites for both nuclear reactors and atomic waste repositories (Roedder, 1984). Simply, fluid inclusion study provides an insight to understanding physical and chemical processes which have affected a certain geologic region. The object of this particular study is to determine the properties of volatiles trapped as fluid inclusions in a compositional and chronological range of sedimentary Archean rock samples to help define the conditions under which life arose on Earth.

Analysis of volatiles trapped as primary fluid inclusions, that is, fluid inclusions formed during the initial formation of a host crystal without subsequent metasomatic and/or hydrothermal alteration or fracturing, may reflect the nature of the fluids of formation. These fluids, most of which formed, presumably, at near-surface temperatures and pressures, may represent the composition of the early volatile components present in, or being added to, the Earth's early atmosphere and hydrosphere in the Archean (Gibson, 1982).

As an LPI summer research intern under the direction of Dr. M.A. Sommer at NASA-JSC, my contributions to the Archean project included petrographic determination, description and locations of primary and secondary inclusions (inclusions formed during the healing of a crystal fracture, subsequent to crystal formation) in a suite of sedimentary Archean sample thin sections, photomicrography of selected inclusions, and molar ratio determinations of overall volatile contents within samples using a crushing tube-quadrapole mass spectrometer system.

Under transmitted light, petrographic determination between primary and secondary fluid inclusions (criteria for petrographic determination between primary and secondary inclusions available in Roedder, 1984), as well as descriptions and thin section localities of these inclusions, was conducted on a suite of cherts (3.5-3.3Ga) from the Barberton greenstone belt in
northeastern South Africa, and on carbonates from the Cheshire formation (2.7Ga) of the Belingwe greenstone belt in Zimbabwe (fig.1). Selected inclusions were circled and logged using a calibrated stage for later relocation necessary for selective decrepitation, or "opening-up", of inclusions for volatile analysis using the laser microprobe/gas analysis system.

Determination of primary fluid inclusions was made difficult due to the slight metamorphism (lower greenschist facies) and hydrothermal alteration of the samples. The small, primary inclusions, mainly in chert, average between 2-6 microns and are most convincing when seen in radiating groups that, at times, outline grain boundaries in micro-crystalline quartz. Second phase vapor bubbles, commonly noted in larger inclusions in hydrothermal quartz veins, are absent in the primary inclusions in chert. The absence of vapor bubbles in chert primaries appears to indicate that either the inclusions are too small to contain a vapor phase, or the vapor has homogenized with the fluid at near-surface temperatures and pressures.

Using a 35mm camera attachment on a petrographic microscope, photomicrographs of selected inclusions within the samples were made to aid in relocation of described inclusions and to show a visual sampling of the characteristic features of distinctive primary and secondary inclusion groups within chert, carbonate and hydrothermal quartz veins.

With the induction of the new, highly sensitive Teknivent mass spectrometry data system for data analysis and processing for the Finnigan quadrapole mass spectrometer at NASA-JSC, more reliable semi-quantitative data related to volatile content and molar ratio determinations of volatiles within single, decrepitated fluid inclusions or crushed bulk samples can be quickly generated.

To obtain a rough survey of the volatiles contained in bulk rock samples, a glass crushing tube is used in conjunction with the quadrapole mass spectrometer. After a sample has been crushed, captured gases are pumped into the mass spectrometer where they are analyzed according to mass number and intensity. At the end of a run, mass spectrums can be displayed for each scan (fig.2). Integrations under individual mass peaks, and molar ratios of peak areas of different masses can be calculated almost instantaneously.

Analysis of Archean samples using the Teknivent data system is in the
## Archean Samples Under Study

<table>
<thead>
<tr>
<th>Locality</th>
<th>Age, b.y.</th>
<th>Depositional Environment</th>
<th>Metamorphic Grade</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Frere Formation</td>
<td>1.7</td>
<td>Shallow basin</td>
<td>Unmetamorphosed</td>
<td>Hall and Goode (1978)</td>
</tr>
<tr>
<td>Earhaeedy Group</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Western Australia</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Hamersley Group</td>
<td>2.5</td>
<td>&quot;Quiet&quot; subaqueous basin, Chemical Precipitates</td>
<td>Prehnite-Pumpellyite Facies</td>
<td>Trendall (1976)</td>
</tr>
<tr>
<td>Marra Mamba Iron Fm.</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Western Australia</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Manjeri Formation</td>
<td>2.64</td>
<td>Black chert (Reducing?)</td>
<td>Greenschist</td>
<td>Martin et al. (1980)</td>
</tr>
<tr>
<td>Belingwe Greenstone Belt,</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Zimbabwe, Africa</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Cheshire Formation</td>
<td>2.7</td>
<td>&quot;Quiet&quot; subaqueous area with subsequent evaporation</td>
<td>Lower Greenschist</td>
<td>Abell (1982)</td>
</tr>
<tr>
<td>Belingwe Greenstone Belt,</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Zimbabwe, Africa</td>
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<tr>
<td>Onverwacht Group</td>
<td></td>
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<tr>
<td>Greenstone Belt, South</td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Africa</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Warrawoona Group, Pilbara</td>
<td>3.5</td>
<td>Deposition in a shallow, hypersaline, water body with subsequent evaporation</td>
<td>Lower than Prehnite-Pumpellyte Facies</td>
<td>Eriksson (1980)</td>
</tr>
<tr>
<td>Block Western Australia</td>
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<td></td>
<td></td>
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<tr>
<td>Isua</td>
<td>3.7</td>
<td>Subaqueous, volcanic-exhalative sequence</td>
<td>Amphibolite</td>
<td>Bridgewater et al. (1983)</td>
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<tr>
<td>Isua Supracrustal Belt,</td>
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<td></td>
<td></td>
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<tr>
<td>Greenland</td>
<td></td>
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</tbody>
</table>

*Fig. 1. Larger sampling of Archean sediments under study for volatile inclusion research (from Gibson, 1982).*
The above graph displays the mass spectrum for scan time 144.18. The x-axis displays mass numbers, while the y-axis shows relative ion intensities. Below is a listing of mass numbers detected in the mass spectrometer and their corresponding intensities.
begining stages; however, initial runs of the Barberton cherts have shown high concentrations of H$_2$O with CO$_2$/H$_2$O ratios averaging 0.35%, and CH$_4$/H$_2$O ratios averaging 0.02%. Analysis of closely known CO$_2$/H$_2$O concentrations in synthetically-grown quartz inclusions may be conducted soon to further test the sensitivity and reliability of the new Teknivent$^R$ data analysis system.

In order to fulfill the objectives of the Early Solar System Volatiles project, further work on the these Archean samples must be accomplished in the area of fluid inclusion analysis. This analysis includes microthermometry techniques of heating and freezing inclusions to determine the nature of trapped volatiles, further crushing of samples using the quadrupole mass spectrometer to obtain a semi-quantitative survey of volatiles trapped within samples, and selective decrepitation of primary fluid inclusions using the laser microprobe/gas analysis system to gain insight to the possible composition of the hydrosphere and atmosphere of the pre-biotic Earth.

REFERENCES


Gibson, E.K.(1982), Early solar system volatiles as recorded in Archean samples (proposal summ.), SN7, Geochemistry Branch, NASA-JSC.

