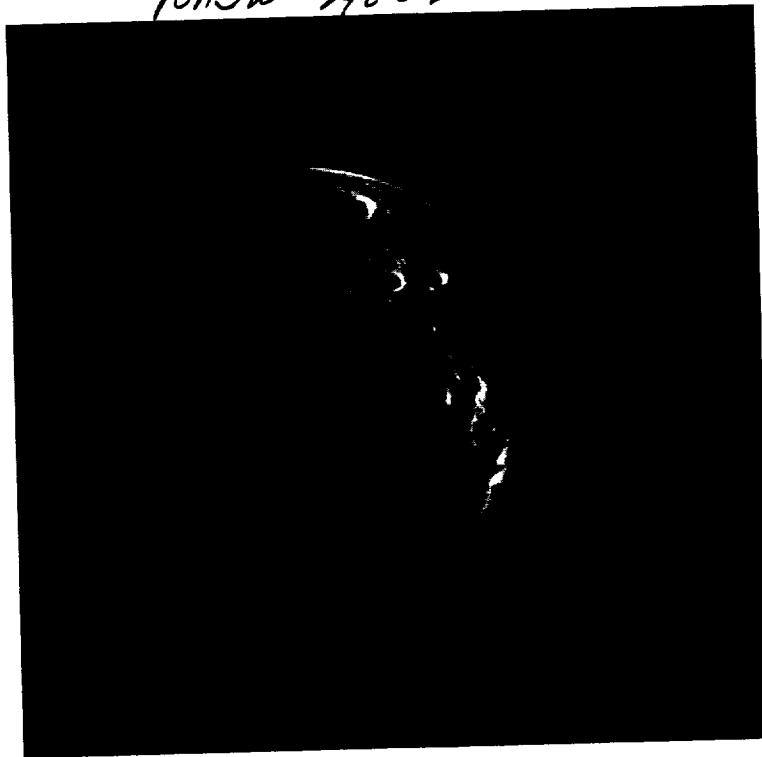


# MEVTV WORKSHOP ON EARLY TECTONIC AND VOLCANIC EVOLUTION OF MARS

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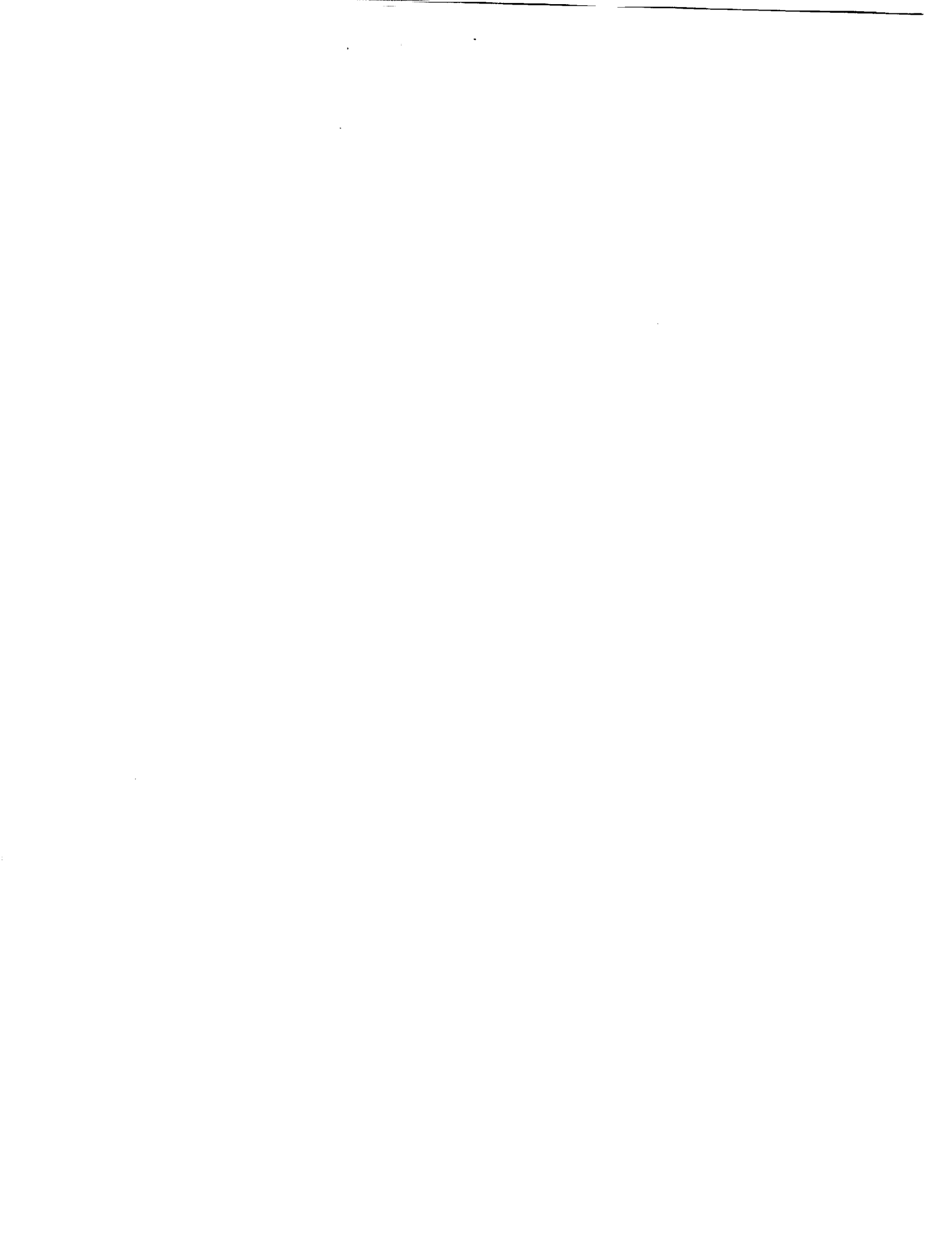
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MEVTV WORKSHOP ON EARLY TECTONIC AND VOLCANIC  
EVOLUTION OF MARS

Edited by  
H. Frey

October 5-7, 1988

Held at  
Easton, Maryland

Sponsored by  
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# Introduction

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Although not ignored, the problems of the early tectonic and volcanic evolution of Mars have generally received less attention than those later in the evolution of the planet. Specifically, much attention has been devoted to the evolution of the Tharsis region of Mars and to the planet itself at the time following the establishment of this major tectonic and volcanic province. By contrast, little attention has been directed at fundamental questions, such as "What were the conditions that led to the development of Tharsis?" and "What is the cause of the basic fundamental dichotomy of the martian crust?" It was to address these and related questions of the earliest evolution of Mars that a workshop was organized under the auspices of the Mars: Evolution of Volcanism, Tectonism, and Volatiles (MEVTV) Program.

The Workshop on Early Tectonic and Volcanic Evolution of Mars was held October 5-7, 1988, on the eastern shore of Maryland at the historic Tidewater Inn. The number of attendees was about 50, and the setting and program provided ideal and welcome opportunity for ample discussion of both contributed papers and more general issues.

A fair question that had to be addressed was what is meant by "early Mars." For a number of reasons I consider early Mars as that part of martian history up through the Early Hesperian in Tanaka's 1986 stratigraphy. At the boundary between the Early and Late Hesperian, martian geologic activity changed from more nearly global to progressively more concentrated at a few localities. Volcanism during the Early Hesperian was widespread; the eruption of the Lunae Planum age ridged plains is recorded throughout Mars. In later times volcanism was more confined to the prominent tectonic centers (Tharsis and Elysium) and to the northern plains. There is also some indication of climatic change at the boundary between the Early and Late Hesperian: Widespread runoff channels give way to the large, prominent, but more localized, outflow channels.

What are some of the major problems that remain to be solved for early Mars? An incomplete and individual list might include:

1. What was the origin and original extent of the crustal dichotomy of Mars? When was this dichotomy established? Was the primary process endogenic or exogenic?
2. Are the cratered highland and lowland plains actually compositionally distinct? Are both basaltic in composition?
3. What was the nature of early geochemical processes? How differentiated is Mars? Were there early surface-atmosphere interactions that were important? Has there been recycling of crustal material?
4. What was the role of large and very large impacts in early crustal evolution? Did these do more than introduce topographic variations? Did they localize later tectonic and volcanic events?
5. Was early mantle convection an important process in crustal evolution? Did convection cause subcrustal erosion? Did it provide for localization of heat sources in the crust, or contribute dynamic support early in the evolution of the major tectonic centers?
6. How and why has martian volcanism changed with time? What was the origin of early highland volcanism and the reason for its apparent early demise? What caused volcanic and tectonic processes to become localized later in martian history?
7. What were the nature and origin of early stresses in the crust? How old is the oldest faulting, and was it only impact related or were other processes equally important?
8. What led to the formation of the Valles Marineris system of large grabens? Why is there only one such system? When did the original faulting begin? What is the relationship of the Valles Marineris to Tharsis?
9. What led to the development of Tharsis and Elysium? Why are there two (and only two) such prominent centers? Why is Elysium so "underdeveloped" in comparison with Tharsis?
10. Were there major early excursions of the martian pole of rotation? Is the evidence for polar change really only climatic change?

The above list is by no means complete, but it does provide some idea of the large number of very fundamental questions about Mars that have their roots in the earliest history of the planet. The workshop did not address all of these issues, but many of these questions were the points around which the four half-day sessions were organized. It is fair to say that no issues were settled as a result of the workshop; but approaches, progress, shortcomings, and new

directions were laid out. Having raised the questions was important; it is clear that there is much we still do not understand about Mars (especially early Mars). It is equally clear that progress on some of these questions can be made, even with existing data and modeling approaches.

Summaries of each of the four sessions are included in this report, thanks to the efforts of the reporters who worked hard to keep up with the many questions and comments that characterized the discussion periods. It was perhaps these periods of discussion as much as the actual presentations that made this workshop so successful.

*Herbert Frey  
Greenbelt, Maryland*



# Program

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Wednesday, October 5, 1988  
8:30 a.m.-12:15 p.m.

## SESSION I: CRUSTAL DICHOTOMY Chairman: K. L. Tanaka

\*The Relevance of Knobby Terrain to the Martian Dichotomy  
*D. E. Wilhelms and R. J. Baldwin*

\*The Martian Crustal Dichotomy  
*G. E. McGill*

Origin of the Martian Crustal Dichotomy  
*H. Frey and R. A. Schultz*

### DISCUSSION

Structural Modification Along the Cratered Terrain Boundary, Eastern Hemisphere, Mars  
*T. A. Maxwell*

Origin of Fluvial Valleys and Early Geological History, Aeolis Quadrangle  
*G. R. Brakenridge*

Ground Ice Along the Northern Highland Scarp, Mars  
*B. K. Lucchitta and M. G. Chapman*

Northern Plains of Mars: A Sedimentary Model  
*R. S. Saunders*

### DISCUSSION

1:30 p.m.-5:00 p.m.

## SESSION II: CRUSTAL DIFFERENTIATION/VOLCANISM Chairman: J. R. Zimbelman

\*Chemical and Physical Properties of Primary Martian Magmas  
*J. R. Holloway and C. M. Bertka*

Sulfide Mineralization Related to Early Crustal Evolution of Mars  
*R. G. Burns and D. S. Fisher*

\*Volatiles, Outgassing, and Water Evolution  
*B. M. Jakosky*

### DISCUSSION

\*Early Volcanism on Mars: An Overview  
*R. Greeley*

The Martian Highland Paterae: Evidence for Explosive Volcanism  
*D. A. Crown and R. Greeley*

Potential Indicators of Pyroclastic Activity near Elysium Mons, Mars  
*K. McBride, J. R. Zimbelman, and S. M. Clifford*

Conditions on Early Mars: Constraints from the Cratering Record  
*N. G. Barlow*

### DISCUSSION

\*Refers to invited papers.

**Thursday, October 6, 1988**  
8:30 a.m.-12:00 noon

**SESSION III: THARSIS, ELYSIUM, VALLES MARINERIS**  
Chairman: H. Frey

- \*Volcanotectonic Provinces of the Tharsis Region of Mars: Identification, Variations, and Implications  
*K. L. Tanaka and J. M. Dohm*
- \*Models for the Origin of Tharsis  
*R. Phillips*
- \*Late Iron Loss from the Martian Lithosphere: Not a Likely Cause of Vertical Tectonics  
*N. H. Sleep and K. J. Zahnle*

DISCUSSION

The Role of Mantle Convection in the Origin of the Tharsis and Elysium Provinces of Mars  
*W. S. Kiefer and B. H. Hager*

Heterogeneities in the Thickness of the Elastic Lithosphere of Mars: Constraints on Thermal Gradients, Crustal Thickness, and Internal Dynamics  
*S. C. Solomon and J. W. Head*

An Ancient Valles Marineris?  
*R. W. Wichman and P. H. Schultz*

DISCUSSION

**Friday, October 7, 1988**  
8:30 a.m.-12:00 noon

**SESSION IV: RIDGE AND FAULT TECTONICS**  
Chairman: S. Solomon

Fault and Ridge Systems: Historical Development in Western Region of Mars  
*D. H. Scott and J. Dohm*

Noachian Faulting in the Memnonia Region of Mars  
*J. R. Zimbelman*

Origin and Age of Grooved Features in the Memnonia Quadrangle (MC-16), Mars  
*R. A. Craddock, R. Greeley, and P. R. Christensen*

Early Resurfacing Events on Mars  
*H. Frey, J. Semeniuk, and T. Grant*

DISCUSSION

Towards a Chronology of Compressive Tectonics on Mars  
*A. F. Chicarro*

Origin of Wrinkle Ridges  
*J. Plescia and M. Golombek*

Periodically Spaced Wrinkle Ridges in Ridged Plains Units on Mars  
*T. R. Watters*

The Shallow Structure of the Lithosphere in the Coprates and Lunae Planum Regions of Mars from the Geometries of Volcanic Plains Ridges  
*M. T. Zuber and L. L. Aist*

DISCUSSION

# Workshop Summaries

## SESSION I: CRUSTAL DICHOTOMY

R. A. Schultz

Don Wilhelms (coauthor, R. J. Baldwin) described the relationship between the age of knobby terrain and the existence of the Borealis basin. Ages previously suggested for knobby terrain (based on crater counts) within the northern lowlands were not contested. Wilhelms and Baldwin view the Borealis basin as having occurred so far back in martian history that it predates all observed geology. Unfortunately, this assumption also means there will be little evidence of the impact remaining for testable study. The authors also like the idea of a large (not giant) impact basin in southwestern Tharsis in Daedalia, as also suggested by Herb Frey and Richard Schultz. No explanation as to why Tharsis is located outside or astride the proposed Borealis rim was offered. The inference is that in their model the origin of Tharsis is not related to the origin of the crustal dichotomy (as produced by the giant impact that formed the Borealis basin. During the discussion that followed, George McGill questioned the need for a giant impact to explain the lowland topography, since his proposed Utopia basin could account for many of the effects suggested for Borealis (at least for eastern Mars). Frey pointed out that crater retention ages for the knobby terrain near the center of the proposed Borealis basin were as great as those in the southern highlands. He asked why old knobby terrain of such age would exist at high elevations within the Borealis basin if a giant impact had indeed occurred prior to all martian geologic history and produced a lowlying basin. Various suggestions were offered about interior rings, superimposed impact basins and the like. Baerbel Lucchitta pointed out that Orientale may not be a good model for basin topography and rebound when scaled up to the size of the proposed Borealis basin. Bill Hartmann raised again the issue of whether the adjacent location of Tharsis implied that there was no relation between Tharsis and the Borealis basin. In response to a question by Sean Solomon about the size of the transient cavity of his proposed basin, Wilhelms suggested the full diameter of 7700 km.

McGill summarized his concerns about the Borealis basin hypothesis and that of overlapping large impacts by Frey and Schultz (not yet presented) as the way to explain the origin of the crustal dichotomy. Arguments raised included the poor fit of Borealis to the cratered terrain/lowland plains boundary, the nonbasin-like interior topography of the northern lowlands, and an incorrectly assumed (by

McGill) need for clustering if the dichotomy is to be made by multiple impacts. Scaling questions were again raised during the discussion period. Despite his objections to the impact (single or multiple) hypotheses for the origin of the crustal dichotomy on Mars, McGill offered no alternative model.

Frey (coauthor, R. A. Schultz) presented a model for forming the dichotomy by overlap of large (but not giant) impacts, pointing out that the variable lowland topography and terrain boundary characteristics are more easily explained if several large impacts had contributed. They presented suggestions of additional large impact basins, especially in the Tharsis area and suggested that one of the effects of overlapping large impacts might be to concentrate long-lived thermal and volcanic effects, perhaps leading to the growth of Elysium and Tharsis. During the extended discussion period that followed, Wilhelms raised the issue of what was the appropriate production function for very large impacts early in martian history. Hartmann asked how the  $D^{-2}$  function assumed by Frey and Schultz matched up with that for large impact craters smaller than 100 km. Hartmann also later pointed out that the discussion about the origin of the dichotomy was concentrating on an impact-related model, although there was considerable disagreement about whether one or several impacts were the cause. Steve Squyres argued that the probability of having many large impacts within the small area of the northern lowlands was small and that it was dangerous to extrapolate the expected effects of large or giant impacts from the Orientale model. For eastern Mars, though, at least two large, overlapping impacts have been proposed: Elysium by Peter Schultz and Utopia by George McGill. Frey and Schultz think these dominate the origin of the dichotomy in eastern Mars. It is not clear how relevant such probability arguments are, nor is the significance of the high-standing old relic terrain within the northern lowlands fully understood. Better topography should help resolve some of the questions raised about the number and size of impacts and their contribution to the origin of the northern lowlands of Mars.

Ted Maxwell described his model for erosional landform development along the highland boundary in one portion of eastern Mars. During the discussion, Baerbel Lucchitta noted the similarity of fretted channels to the channeled smooth deposits studied by Maxwell. Ron Greeley emphasized that the origin of knobs is still not clear and that knobs at high elevation may be different from those in lowlying regions.

George Brackenridge described relative ages of wrinkle ridges and channels in the Aeolis portion of eastern Mars. Landform orientation as shown by rose diagrams were an important basis of his conclusions, since no clear embayment relations were found commonly throughout the region. Squyres suggested that if the statistical alignments were real, preexisting structures might be controlling the orientations of both the ridges and the channels. He cited Dave Pieri's work that showed that digitate channels in some areas can predate the ridged plains.

Lucchitta (coauthor, M. G. Chapman) discussed evidence for ground ice along the highland boundary in Ismenius Lacus. It appears that some hills and mesas once had lobate debris aprons that were embayed by plains and then had the debris aprons removed. The mechanism for removing the aprons to form the depressions around the hills remains a problem.

Steve Saunders summarized observational evidence and calculated estimates of the martian water inventory, then discussed locations within the northern lowlands where lacustrine deposits appear to demonstrate that a Borealis "ocean" may once have existed. The session ended with an artistic rendition of such a Mars, on which point the meeting was adjourned.

## SESSION II: CRUSTAL DIFFERENTIATION AND VOLCANISM

C. M. Bertka and M. W. Schaefer

John Holloway (coauthor, C. M. Bertka) opened this session with an invited paper discussing the chemical and physical properties of primary martian magmas. The first (1-2%) partial melt from a composition consistent with both phase equilibria and the estimated density of the martian mantle will be relatively rich in iron and poor in magnesium, compared to a similar melt on the Earth. Fractionation of the magma will increase the relative iron enhancement. Because of the higher iron content of the martian mantle, the density cross-over, which determines whether a melt will sink or rise, occurs at a lower pressure on Mars (at 35-40 kbar) than on the Earth (50 kbar). Based on experimental results the viscosity of the melt was very fluid, about equivalent to that of Prell shampoo. During the discussion Sean Solomon pointed out that the moment of inertia of Mars is not well known, so the density of the martian mantle is also too poorly known to estimate the iron content reliably. He also questioned why the phase equilibria studies were done at a pressure of 23 kbar rather than at some lower pressure perhaps more reasonable for early martian magma source regions. Holloway replied that these results were part of an ongoing study of terrestrial melting processes and that data at lower pressures would soon be available. He indicated that the general results

were probably applicable over a range of pressures from 15-300 kbar. Solomon then asked how such dense magmas could reach the martian surface. Holloway pointed out that the addition of as little as 1% water or carbon dioxide in the magma would drastically reduce the density of the melt. If water were the volatile, silica-rich magmas would result; carbon dioxide would produce silica-poor, low viscosity magmas. Estimates of flow length for martian volcanic flows (the few that exist) favor the low viscosity case.

Roger Burns (coauthor, D. S. Fisher) described how sulfide mineralization could be related to early crustal evolution of Mars. Komatiites have been proposed as analogs of early martian lavas, based on compositional similarities to SNC meteorites and martian fines (from Viking XRF data) and on their presence on the early Earth. On the Earth komatiites are often associated with massive sulfide deposits. Because sulfur is very soluble in iron-rich basaltic melts, there could be significant sulfur in the martian mantle and in martian lava flows. This might provide the source for the observed sulfur in the martian fines. During the discussion Herb Frey asked whether most of the sulfur would have been depleted early on Mars and whether later lava flows might be sulfur poor. Are not the fines some long-time average of the entire martian crust? George McGill asked whether the source region would have to be deep to account for the high temperatures normally associated with komatiitic lavas (on the Earth). Burns replied that early geothermal gradients on Mars were probably high, a point echoed by Norm Sleep. Solomon asked about chemical equilibrium between the mantle and an assumed iron-rich core, wondering if the martian mantle could indeed be sulfur rich. Burns replied that with a chondritic abundance of sulfur on Mars, equilibrium between the core and mantle still leaves appreciable amounts of sulfur in the martian mantle, so the komatiitic model has merit.

Bruce Jakosky (coauthor, T. A. Scambos) discussed the volatile inventory, outgassing history and evolution of water on Mars. Based on a variety of observational and theoretical estimates, fifty to several hundred meters (surface equivalent) of water has outgassed from Mars and several hundred meters or more still remain within the planet. Most of the outgassed water has escaped to space, especially during times of maximum insolation stemming from obliquity variations. Most of the remainder is currently locked up in the polar caps. Questions during the discussion period focused on isotopic comparisons between the Earth and Mars. The D/H ratios for Venus and Mars are higher than for the Earth. This is taken to indicate loss of water to space. This means the model for loss is highly dependent on the nitrogen isotopic ratio values. Solomon expressed concern about the Greeley calculations of outgassed water



over martian history, because these are critically dependent on the assumed amount of water in martian lavas. Jakosky replied that the best estimates for minimum outgassed water on Mars came from Carr's calculation of the amount required to carve the outflow channels. These agree with Greeley's estimates. Solomon asked whether this might not be water accreted as a veneer late in the formation of Mars. Holloway pointed out that the presence of hydrated minerals in one SNC meteorite suggested water in the martian interior. Sleep said it is possible to put almost arbitrary amounts of water in sediments and hydrated silicates, where it would remain isolated from the atmospheric D and H budget.

Ron Greeley presented an overview of early volcanism on Mars. The emphasis was on photogeologic studies of flood-type, apparently low-viscosity lavas and on the highland patera, which may have formed as phreatomagmatic eruptions of lavas through a water-saturated regolith. He suggested that komatiites were a good analog for the lavas and that high discharge rates were likely to have produced turbulent flow and significant mechanical erosion such as that caused by catastrophic water flow on the Earth. The highland patera eruptions may have produced large volumes of ash deposits. Jakosky asked whether the early atmospheric pressure on Mars could suppress explosive volcanism, but the answer appears to be no.

Solomon wanted to know how long a single lava flow on Mars could be. Jim Zimbleman and Greeley replied that some could be up to 500 km long. Ben Chao asked about the importance of gravity on volcano morphology. Greeley replied that while gravity is important, the high diffusion rate is a more dominant factor. Frey again asked if the main reason for the komatiitic analog model was the XRF analysis of martian fines, and if so, were these not a time-average of all martian volcanism, not confined to just early martian time? Greeley admitted this could be the case but suggested that the lavas were from a single basin. Greeley also replied to a question by Frey as to whether komatiitic lavas were reasonable analogs for younger volcanic events on Mars by stating emphatically that the morphology of the large, young shield volcanoes is not consistent with komatiites.

Dave Crown (coauthor, R. Greeley) then discussed evidence for explosive volcanism on the martian highland patera. Two cases for such eruptions exist: lava can rise through a water-saturated regolith, or the lavas themselves can be water rich. Energetics for either possibility appear reasonable to produce the observed morphologies seen on Mars. Pyroclastic flow is more likely than air-fall ash deposits because of the large inferred volumes of the patera. Crown favors the hydromagmatic origin rather than water-rich lavas, suggesting that the decrease in explosive activity over

time could be due to progressive loss of water from the regolith. In discussion Zimbleman asked whether there was evidence for multiple explosions at the same site, but Crown said the resolution of the available imagery was not good enough to answer the question. Crown also pointed out his model for flow was conservative and not likely to be sensitive to choice of coefficient of friction, which Walter Kiefer had suggested was incorrect. Baerbel Lucchitta asked if the explosive activity would put large amounts of water into the early atmosphere. Ken Tanaka suggested that the martian surface was even more permeable than previous estimates, making Crown's model even more plausible.

Kathleen McBride (coauthors, J. R. Zimbleman and S. M. Clifford) described potential indicators of pyroclastic activity near Elysium Mons. The eastern flank of the volcano appears relatively smooth with no distinct lava flows. Impact craters are subdued as though mantled. Cinder cones and other small domes are widespread; volcano-ground ice interactions seem possible. The slope of Elysium Mons steepens toward the summit to about 18°, which may indicate pyroclastic activity. In the discussion that followed there was some debate about the relevance of steep upper slopes. Frey asked why there were so few indicators of pyroclastic deposits observed on Mars, given the abundant evidence for both extensive volcanism and water. Greeley noted that ash flows are not easily distinguished from aeolian deposits and also are easily degraded. Lucchitta recalled Dave Scott's idea that there are extensive ash flow deposits on Mars. Sleep said that many pyroclastic constructs are quite small and may be below the resolution limit of current imagery. Lucchitta suggested that layered deposits in the interior of Valles Marineris may be ash deposits.

Nadine Barlow presented the final talk of the session, discussing constraints on the conditions of early Mars provided by the cratering record. Her results of crater counts suggest that small volcanoes on Elysium and Tharsis formed at the end of the heavy bombardment during the major episode of intercrater plains formation in the southern highlands. Fretted terrain along the highland boundary has a much younger age, with knobby terrain in the northern lowlands having a similar age. The discussion that followed was lively. Don Wilhelms questioned exactly which craters were used for counts in the knobby terrain. Barlow replied only whole, clearly visible (no partial or ghost) craters were counted. The counts include both superimposed and protruding craters, but separation on the basis of geologic unit had been done. Lucchitta and Frey spoke about the alternative technique, used by Neukum and Hiller, in which all craters are counted and geologic events are deduced from changes in slope of the curves. Frey reported that his studies of many of the same areas described by Barlow produced similar results for the sequence of events.

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**SESSION III:  
THARSIS, ELYSIUM, VALLES MARINERIS**  
W. B. Banerdt

Ken Tanaka (coauthor, J. M. Dohm) opened the session with a description of volcano-tectonic provinces in western Mars. Much greater detail is now available in the stratigraphic record, and the authors have tentatively identified seven volcano-tectonic provinces surrounding the Tharsis area, each with locally distinctive features. Faults and ridges have been grouped by geologic age, based on the stratigraphic relationships augmented by crater counts. Questions were raised about the dating of linear features by crater counts vs. geologic relationships. A general consensus emerged that geologic relationships provide the best constraints in active areas such as Tharsis. In more homogeneous regions, linear crater counts become the method of choice.

Roger Phillips summarized much of the theoretical work done in the last 10 years on the evolution of Tharsis, then described his recent work on isostatic open-system magmatic differentiation models. He concludes that the source depth for Tharsis magmas must have been less than 250 km and that the crust has been thickened and its density increased by intrusion. There was no proposal for what led to the origin of Tharsis. In the discussion that followed, Bruce Jakosky asked about the effect of thermal contraction on the stress models. Phillips replied that this would provide a late but comparatively small load on the lithosphere and that other loads were probably more important in Tharsis tectonics. Sean Solomon pointed out that in a convecting system, the residual products of partial melting might be recycled into the mantle and replaced by pristine material. Phillips said his model assumes that the source region lies above the convecting portion of the mantle, interacting only in a thermal sense. Herb Frey noted that while the Tharsis stress models adequately explain the orientation of the Valles Marineris, these models do not offer any explanation as to how the huge canyons formed or why they are so deep. Tanaka added that much of the local faulting, although consistent with regional trends, appears to have been triggered by local activity. Norm Sleep suggested that some insight into the past state of stress could be derived from reconstructed "paleo-topography" based on old drainage patterns.

Sleep (coauthor, K. J. Zahnle) then discussed the theory that late iron loss from the lithosphere could provide a mechanism for the formation of the crustal dichotomy and/or Tharsis. He concluded that the mechanism is unlikely to be important, based on constraints from Pb isotope ratios in shergottites and the lack of any evidence for late energetic impacts. Responding to a question by Jakosky, Sleep stated that the time of Fe-loss could be

determined from Pb ratios to an accuracy of 0.1 to 0.5 b.y. Regarding the question of a late impact flux, Solomon asked whether anyone had ever added up the material involved in making the observed impact basins to see whether these represented a significant addition to the crust. Sleep replied that his estimate was that 30 Orientale-sized impacts would be needed to have any major effect, but that scaling uncertainties for large diameter basins made such an estimate difficult. Furthermore, there is little evidence for a special class of late impacting large objects, which the mechanism would require.

Walter Kiefer (coauthor, B. H. Hager) presented a convective model for support of Tharsis and Elysium. He suggested that the broad swells of Tharsis and Elysium implied internally heated convection. Bottom-heated convection would produce narrow plumes that would produce more localized volcanic structures at the surface. Using a finite-element convection model with a Rayleigh number of  $10^6$ , the computed amplitudes of the long-wavelength topography and gravity were similar to those seen for Tharsis and Elysium. Solomon asked what fraction of the planetary heat flow would be contributed by the postulated two major plumes (one each for Tharsis and Elysium) and why convective effects on Mars should be so much greater than those on Earth. Kiefer replied that he used the average planetary heat flux in his model (which underestimates the actual heat flow contribution) and that on the Earth the higher Rayleigh numbers appropriate for the mantle would produce smaller surface effects. Don Wilhelms asked whether the plumes "care" what the surface is like (are plumes affected by near-surface structure?). Kiefer replied that small plumes responsible for discrete volcanic constructs will "prefer" to form in the plateau regions associated with general upwelling. In response to a question by Frey, Kiefer said that the style of convection he described (one or two major upwellings with a number of smaller plumes) might not be typical of all martian history. Phillips suggested that as the planet evolves thermally and the Rayleigh number changes, an increase could be expected in the amount of uplift accompanying convection, for which there is no evidence. Kiefer said this effect could be counterbalanced by an increase in lithospheric thickness with time.

Solomon (coauthor, J. W. Head) presented a paper in which he used inferred variations in the elastic lithosphere thickness to constrain near-surface thermal gradients and the crustal thickness. Temperature differences appear to be too large to be the remnants of impact heating, so he inferred that lithospheric thickness variations were due primarily to mantle dynamic processes. These may be similar to lithospheric reheating that occurs beneath hot spot volcanic centers on the Earth. Bill Hartmann asked whether impacts could cause major variations in the lithospheric

thickness. Solomon replied that this is a second order effect compared to mantle dynamic processes. Frey noted that multiple or overlapping impacts in the same region would prolong the local heating. Phillips pointed out that the temperature cannot continue to rise indefinitely over a static hot spot, but Solomon said he could not say anything quantitatively about the nature of the plume. He noted that even at Hawaii it is not clear whether the lithosphere is in thermal equilibrium with the plume.

Richard Wichman (coauthor, P. H. Schultz) gave the final paper of the session in which he discussed a shallow trough system near Hellas that he interprets as the remains of an ancient canyon system comparable to Valles Marineris. He suggested the two canyon systems formed along impact-related fracture zones concentric to the Hellas and Chryse basins, respectively. Tom Watters asked whether the observed much shallower depths of the Hellas canyons might be close to the original depths. Wichman noted the age of the Hellas Canyons is roughly the same as Hellas itself, and they are therefore much older than the Valles Marineris. As such they may have suffered greater erosional modification. Frey pointed out that there is no indication of scarp retreat in the Hellas canyons such as is common in the Valles Marineris. Tanaka noted that there is no evidence that the Valles Marineris follow older structures, such as impact-related fracture zones. Wichman replied that the older structures may have been capped by flood volcanics. Ted Maxwell suggested the extent of the Hellas canyons was too limited to truly describe a concentric pattern and wondered if there were other examples elsewhere around Hellas. Wichman stated that this work was still in progress.

#### SESSION IV: RIDGE AND FAULT TECTONICS

R. A. Schultz

The paper by Scott and Dohm was not presented due to illness but had been summarized in the previous session by Ken Tanaka.

Jim Zimbelman described his results of mapping large ridge-like structures in NE Memnonia. These do not appear to be normal wrinkle ridges, having great relief (2-3 km) and are possibly of Noachian age. He suggested an origin by normal faulting rather than by contraction. During the discussion he also pointed out that the orientations of these features were inconsistent with Tharsis-centered stresses, tidal despinning models, or polar wander-induced stresses. Baerbel Lucchitta asked why these ridges look so different from younger ones and suggested that material differences were a possibility. Sean Solomon asked about the origin of the 3 km fault relief. There followed general discussion about timing relationships between cross-cutting Memnonia

Fossae grabens and the high-relief ridge, and why the ridge morphology and relief changes across the apparently younger graben remain unknown.

Bob Craddock (coauthors, R. Greeley and P. R. Christensen) discussed grooved landforms on older highland units and their relationship to basin ejecta. They infer the possible existence of a large impact basin in Daedalia, as suggested earlier in the meeting by Frey and Schultz. George McGill raised questions about sampling bias. Zimbelman asked about parallel subadjacent ridges in Memnonia. Buck Sharpton suggested the grooves may be too far from the suggested source impact basin to be related to it.

Herb Frey (coauthors, J. Semeniuk and T. Grant) discussed estimates of the thickness of resurfacing materials using the Neukum and Hiller crater counting technique. Ridged plains may be as thin as 250 m, and probably do not exceed 500-600 m in Lunae Planum and Coprates. He also suggested that ridge plains in Memnonia thought to be of Noachian age based on total crater counts may in fact be relatively thin flows of Hesperian age through which an older surface still shows. The discussion that followed was lively. Don Wilhelms noted that his approach to crater counts (selecting fresh vs. degraded craters) gave equivalent results to those of Frey et al. (who use bendaways in the cumulative frequency curves to define resurfacing ages). McGill noted that an assumption of the technique was that timescales for resurfacing must be shorter than those for repopulation by new impact craters. The comparison between Wilhelms and Frey et al. suggests this assumption is good for the areas studied. Tom Watters asked about the role of topography in estimates of thickness of resurfacing units. Frey replied that this technique only provided an average over the entire area in which craters were counted. Zimbelman suggested that very high resolution images could be used to determine whether the statistically defined resurfacing coincides with the observed overlap of plains units at the indicated diameter. In response to a question by Jim Garvin, Frey emphasized that the resurfacing ages derived refer to the end of the resurfacing event and not to its duration.

Augustine Chicarro presented some generalizations derived from his extensive data set of 16,000 ridges on Mars, classified by several useful parameters. He described a possible correlation between well-formed ridges and craters having fluidized ejecta blankets. He discussed ridge orientations for different ridge morphologies and locations. During the discussion Wilhelms pointed out the importance of separating age of ridges from ages of underlying terrain. Lucchitta asked about the significance of high-relief vs. low-relief ridges. Sharpton stated that a volatile-rich substrate was not required for the formation of lunar wrinkle ridges. He suggested mechanical decoupling of competent and less competent underlying materials. Roger Phillips

asked whether different models for Tharsis-centered compressive stresses could be discriminated by Chicarro's data set.

Jeff Plescia (coauthor, M. P. Golombek) presented their model for production of ridges by thrust faults. Ridge morphology would be due to a combination of the main thrust and additional near-surface faults. One problem with the buckling-instability model for ridge spacing control is that shallow dipping limbs implied in their ridge topography models (about 5°) do not allow faulting to begin along the fold limbs (need about 25°). In the discussion period Jayne Aubele asked how they chose to partition strain between faulting and folding. With their choice, thrust faults fail to reach the surface, which is contrary to their preferred model. Tom Watters noted apparent reversals in the sense of ridge asymmetry and suggested this surface phenomenon argued against thrust faults dipping in opposite directions. Ted Maxwell noted the Plescia-Golombek model required secondary, near-surface splay faults to control most of the wrinkle ridge topography, rather than the main thrust fault. Lucchitta reported no evidence for splay faults in a terrestrial analog, the Meckering thrust in Australia.

Watters presented an analysis of wrinkle ridge spacing as controlled by buckling instability. He invoked freely slipping multilayers to reduce critical stress for buckling to reasonable levels. In this way he could explain ridge spacing with a combination of thick competent layers above much weaker material, assuming the underlying lithosphere

did not participate in the folding. In the discussion Richard Schultz noted that regular spacing of wrinkle ridges must be explained by a successful folding or faulting model. Based on fault mechanics considerations, he pointed out that thrust faults must be at least twice as deep as ridge spacing for the faulting model to be applicable. Sharpton asked about the role of initial topography (or topographic perturbations) in localizing ridge location. Norm Sleep raised the spectre of thick-skinned vs. thin-skinned martian tectonics. He speculated that creep at depth along deep faults may contribute to ridge spacing, thereby implying a thick-skinned model was appropriate.

Maria Zuber (coauthor, L. L. Aist) also investigated buckling instability as controlling ridge spacing and summarized results from a suite of models, including both thick- and thin-skinned folding. During the discussion Watters pointed out that the thickness of the underlying lithosphere was unconstrained in their model and that very long wavelength gentle folding would be implied. McGill spoke about the use of viscous vs. elastic formulations and mentioned cases in which they could be interchangeable. Lucchitta asked the dreaded but relevant question: Where did the compressive stress for folding or faulting come from (a point ignored in all the presentations)? To date none of the analyses have been able to link stress magnitudes required for control of ridge spacing with those associated with Tharsis evolution.

# ABSTRACTS



CONDITIONS ON EARLY MARS: CONSTRAINTS FROM THE  
CRATERING RECORD. N. G. Barlow, Lunar and Planetary Institute,  
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Crater size-frequency distribution data provide important information about the relative ages of different geologic units. The shapes and vertical positions of size-frequency distribution curves give identifying information about the population of impactors responsible for the crater record retained within a particular unit. The relative plotting technique is a normalization method which clearly shows the slope variations associated with differences among the impacting populations (1). Within the inner solar system, two populations of impactors have been determined from the shapes and crater densities of the distribution curves: an old population displaying a multi-sloped distribution curve emplaced during the period of heavy bombardment and seen in the heavily cratered regions of the moon, Mercury, and Mars, and a younger population whose distribution curve follows a power law function and which has dominated the cratering record since the end of heavy bombardment. This latter population is primarily recorded within the lightly cratered regions of the Moon and Mars (2). On Mars approximately 60% of the terrain units display distribution curves indicative of formation during the period of heavy bombardment. These units consist of the heavily cratered southern highlands, the equatorial ridged plains, and many of the small volcanic constructs within the northern hemisphere (3).

Size-frequency distribution curves of craters superposed on small volcanoes in the Tharsis and Elysium regions reveal that most of these constructs formed during the period of heavy bombardment. Crater densities indicate that most formed contemporaneously with the major episode of intercrater plains formation in the southern highlands, one of the oldest terrain units seen on Mars (Fig. 1). The process(es) which formed the hemispheric dichotomy therefore must have occurred extremely early in martian history, prior to formation of most if not all terrain units presently existing on the planet's surface. In particular, the fretted and dissected terrain located along the plain-highlands boundary scarp cannot date from the dichotomy-forming event since this region exhibits a relative age younger than that of the small volcanoes north of the boundary. In addition, the relative age of knobby terrain within the northern plains (regions interpreted as buried cratered upland material (4)) is statistically identical to that of the boundary scarp material (Fig. 2), further supporting the assertion that the process causing the global dichotomy, whether internal (5) or external (6, 7), predated the formation of many younger or intermediate heavy bombardment-aged units.

The heavily fractured terrain surrounding Tharsis has been linked to the formation of the bulge itself. Multiple episodes of fracturing are known to exist throughout this region, but among the oldest fractures are those located south of the bulge

CONDITIONS ON EARLY MARS  
Barlow, N.G.

within the heavily cratered southern uplands (5). Crater size-frequency distribution plots of the terrain indicate that this region has an age similar to that of the fretted terrain along the boundary scarp, dating from near the latter part of heavy bombardment. The fractures must post-date the terrain which they cut and they must be younger than the old post heavy bombardment-aged flows in Solis Planum which cover them. Superposition and cross-cutting relationships, together with crater statistical-derived age data for this area, thus suggest that the Tharsis Bulge began leaving a tectonic signature on its surroundings near the end of the heavy bombardment period.

Tectonism and volcanism as well as cratering were important processes shaping the martian surface during the period of heavy bombardment. The cratering record provides important information about the ages of martian terrain units which in turn can be used to constrain the timing of events such as formation of the global dichotomy and fracturing caused by the uplift of the Tharsis Bulge.

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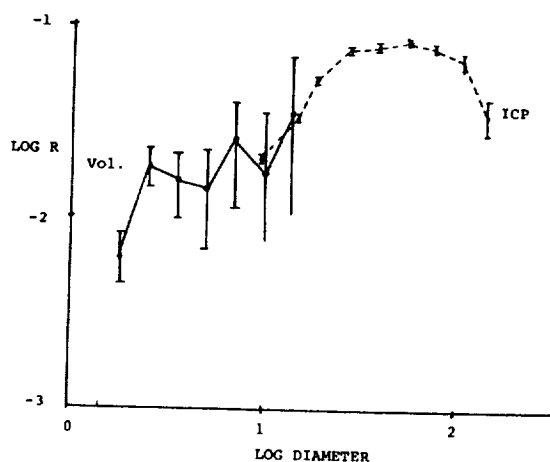


FIGURE 1

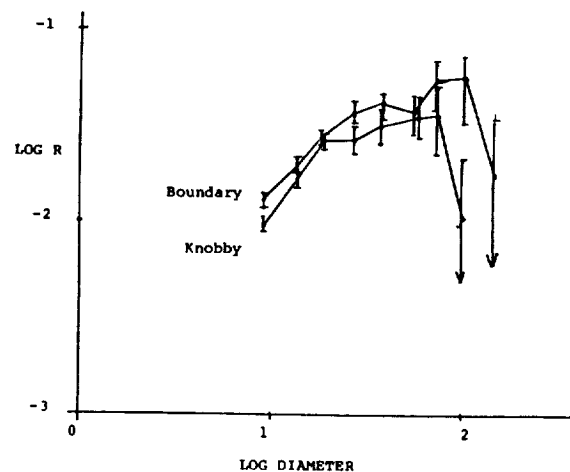


FIGURE 2



ORIGIN OF FLUVIAL VALLEYS AND EARLY GEOLOGICAL HISTORY, AEOLIS

QUADRANGLE

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03755

In southern Aeolis Quadrangle in eastern Mars, parallel slope valleys, flat-floored branching valleys, v-shaped branching valleys, and flat-floored straight canyons dissect the ancient (pre-3.2 Ga) cratered landscape (Fig. A). Associated knife-like ridges are interpreted as fissure eruptions, conical hills as cinder cones, and thin, dark, stratiform outcrops as exhumed igneous sills or lava flows. Extensive ridged lava plains are also common, but are not modified by fluvial processes. I mapped 56 thrust faults in Aeolis (Fig. B). These faults exhibit an orientation vector mean of  $N63W \pm 11$  (95% confidence interval), and they transect the lava plains, the older units, and some of the valleys. By comparison, the vector mean for the 264 valleys mapped is  $N48W \pm 12$ , with a larger dispersion about the mean. The similar orientations displayed by thrust faults and valleys suggests that many valley locations were controlled by pre-existing thrust faults. Other fault or fracture orientations have modes at ca.  $N10W$  and  $N65E$ ; very few valleys, faults, or fractures exhibit orientations between  $N10E$  and  $N50E$ . Detailed mapping also corroborates structural control over some valley locations, and it indicates that development of valley drainage systems occurred over relatively long periods of time, while faulting was still underway (e.g., Fig. C).

The dissected deposits in Aeolis may include interstitial or interbedded water ice, and their Late Heavy Bombardment age agrees with atmospheric models showing stability of ice accumulations in non-polar latitudes at this time. Freshly outgassed and crystallized water became entombed as frost or snow within the cratered terrains during accumulation of impact ejecta, volcanic ash, and recycled eolian debris. Volcanism then caused geothermally heated groundwater to be generated within overlying or adjacent ice-rich strata, and faults and fractures provided thin zones of increased permeability for water transport to the surface. Hot springs were common on Late Heavy Bombardment Mars, and they facilitated valley development by headward sapping and by the generation of episodic, topography-following fluid flows.

DATA SUMMARY FOR TECTONIC AND FLUVIAL LANDFORM ORIENTATIONS

Landforms	n	Vector Mean <sup>1</sup>	Strength of Vector Mean <sup>2</sup>	Standard Error	Raleigh Test For Uniformity <sup>3</sup>
Small Valleys	264	$N48W \pm 12.0$	0.39	6.1	0.00
Undifferentiated Faults and Fractures	83	$N49E$	0.19	23.1	0.05
Thrust Faults	56	$N63W \pm 11.2$	0.75	5.7	0.00

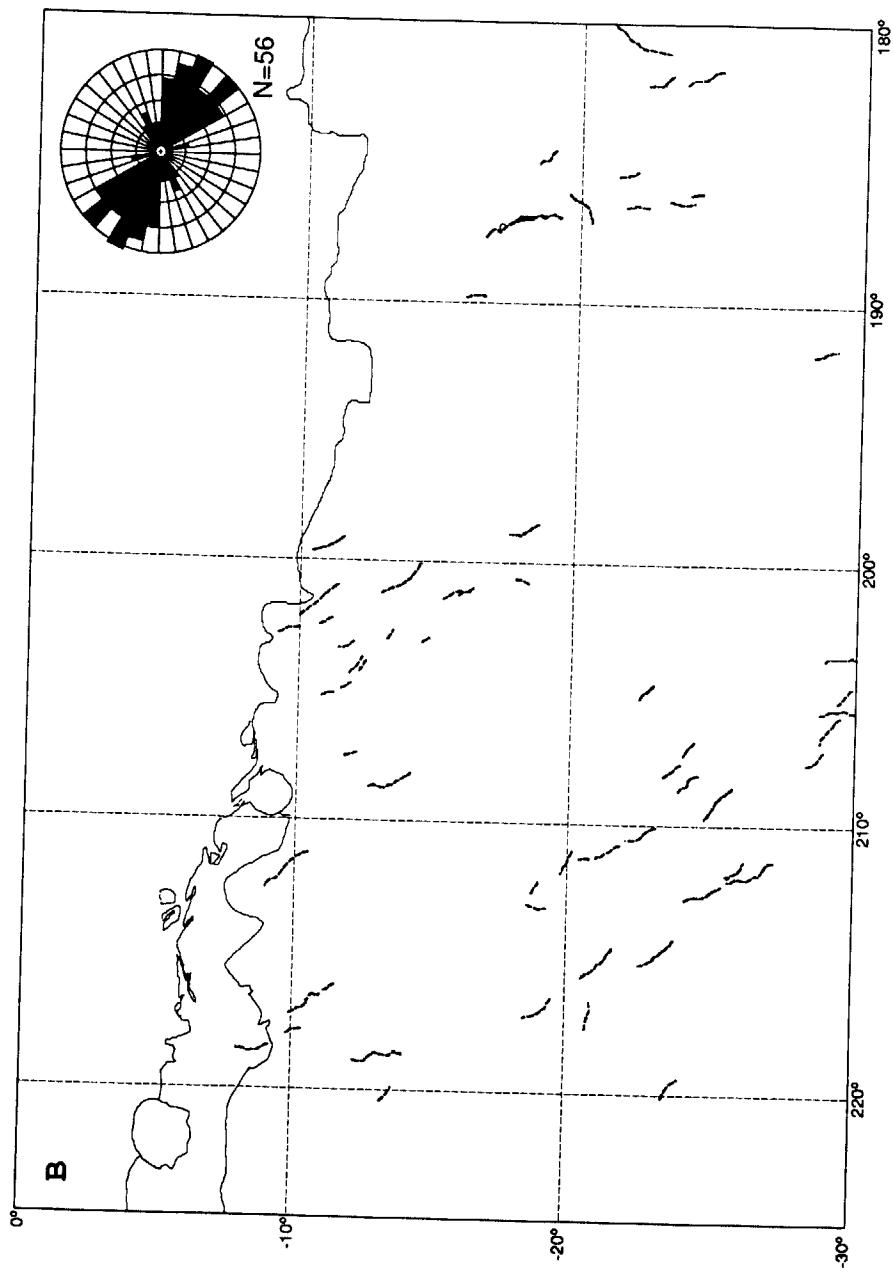
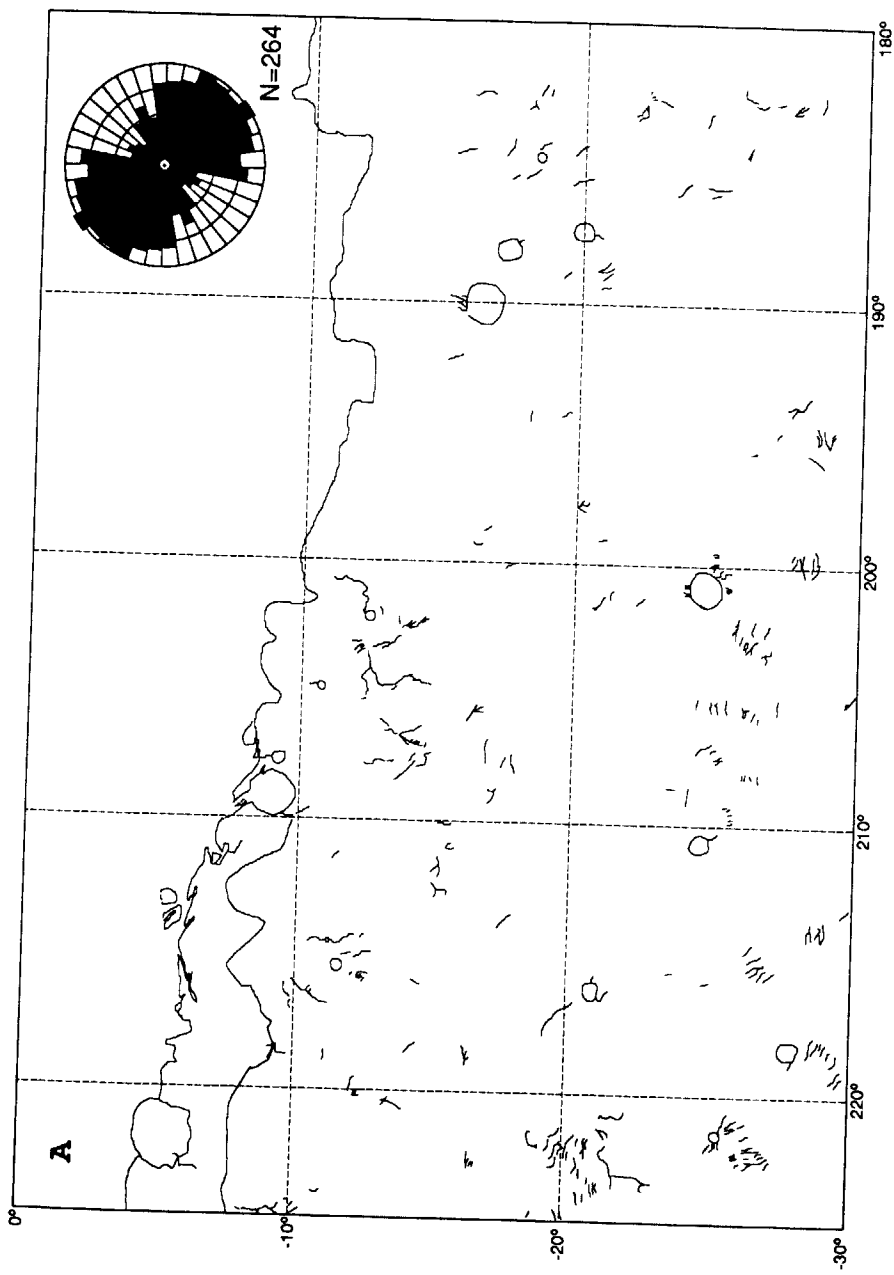
<sup>1</sup> Vector mean is  $\arctan(X/Y)$ ;  $X = \sum \cos \theta_i$ ;  $Y = \sum \sin \theta_i$ ; shown with 95% confidence intervals

<sup>2</sup> Strength of vector mean is  $R/n$ , where  $R = (X^2 + Y^2)^{1/2}$

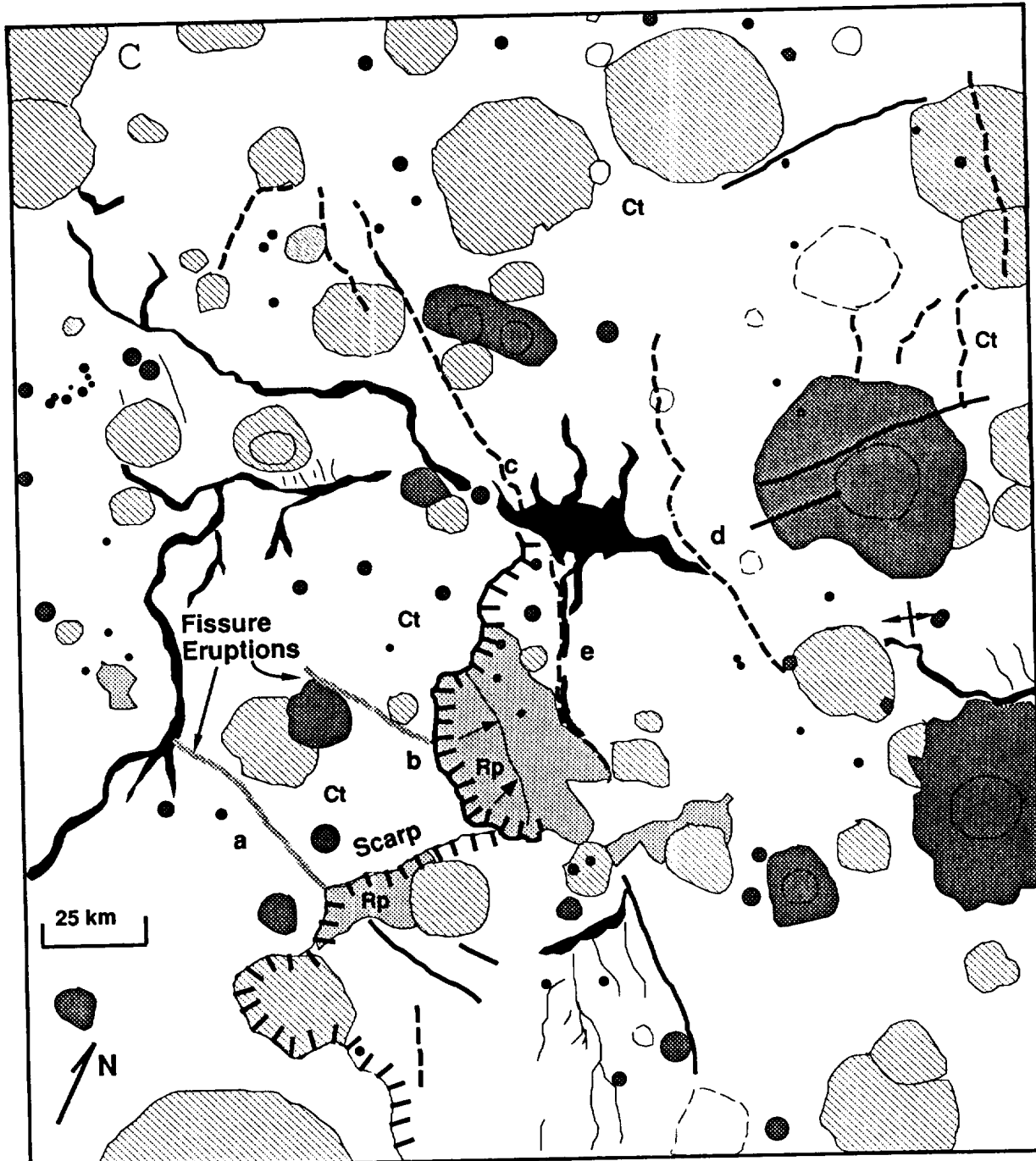
<sup>3</sup> Raleigh statistic is  $\exp -(R^2/n)$ ; for Raleigh values  $< 0.05$ , the uniform vector distribution hypothesis is rejected at the 0.05 level of significance

ORIGIN OF FLUVIAL VALLEYS

Brakenridge, G. R.



Brakenridge, G. R.



GEOLOGICAL MAP OF A PORTION OF VIKING FRAME 596A26

At locations "a" and "b" within the Cratered Terrain (Ct) in Aeolis, knife-like ridges are interpreted as fissure eruptions; their low-albedo cross sections are exposed in the wall of a prominent scarp. The adjacent lava plain ("Rp") is apparently an exhumed sill and may be continuous with at least one of the ridges. NW-trending thrust faults (dashed lines) are common, and, at "c", such a fault transects a flat-floored branching valley. There, thrust faulting is younger than the valley, but, at "e", a major NNW-oriented tributary valley modifies an older thrust fault. Thus, some links of this valley system are younger, and some are older, than individual thrust faults. Upstream (northwest) from the thrust fault and topographic ridge at "d", fluid impoundment is suggested by the broad valley floor and the wide tributary valley mouths: an ice-covered lake may have existed at this site.

## SULFIDE MINERALIZATION RELATED TO EARLY CRUSTAL EVOLUTION OF MARS

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**Summary.** Komatiitic magmatism, which occurred on Earth during stabilization of the crust in the Archean and was associated with ultramafic Fe-Ni sulfide ores and massive volcanogenic Fe-Cu-Zn sulfide deposits, lasted longer on Mars. As a result, ore deposits evolved little on Mars from these pyrrhotite-pentlandite and pyrite-chalcopyrite-sphalerite assemblages due to minimal interactions of martian mantle with the crust, hydrosphere and atmosphere.

**Background.** The age and formation of continental crust on Mars and the existence of highland terrains in the southern hemisphere remain fundamental problems in the evolution of that planet. On Earth, stabilization of the crust and the development of proto-continentals took place during the Archean, >3.8 to 2.5 billion years (b.y.) ago, and were associated with rock-types characteristic of that era, including komatiitic mafic and ultramafic igneous rocks in greenstone belts. Distinctive ore deposits are associated with these komatiites and other Archean rock-types [1], including banded iron formations, gold-uraninite conglomerates, and disseminated and massive sulfide mineralizations hosted by volcanic and plutonic rocks [2,3]. However, terrestrial ore-forming processes have evolved over geologic time as a result of plate tectonic activity along subduction zones, as well as by interactions of the crust with the hydrosphere, developing atmosphere and evolving biosphere [2-4]. The inference that sulfide ore deposits may exist on Mars [5] is based on compositional and petrographic similarities noted between komatiites (the host rock for terrestrial massive Fe-Ni sulfides), SNC meteorites and the silicate portion of Martian regolith fines [6-8]. Paragenetic relationships that might exist between these mafic and ultramafic igneous rocks on Mars raise questions about the localities of possible Archean-type sulfide and related ore deposits on Mars, and whether or not they have undergone tectonically-induced temporal variations on that planet, too. Did ore deposition take place on Mars during the formation of its crust, what were the emplacement mechanisms, and how have such ore deposits influenced the geochemical evolution of the surface of that planet? These problems are central to our investigations of the evolution of Fe-S minerals on Mars [5,9], clues to which may be deduced from associations of terrestrial ore deposits with Earth's earliest crustal rocks.

**Temporal Patterns of Ore Distribution on Earth.** The types and stratigraphic settings of ore deposits have changed through geologic time in response to the tectonic and chemical evolution of the Earth [1-4]. They encompass chromite deposits found in the oldest preserved crustal rocks (e.g. the pre-3.8 b.y. old Isua supracrustal belt in West Greenland) to present-day sulfide "chimneys" and ferromanganese oxide crusts forming at oceanic spreading centers (e.g. at 21°N along the East Pacific Rise). The ore deposits include mineralizations occurring in different types of igneous and sedimentary environments. Types of mineral deposit are unequally distributed in time due to the interplay of processes related to the composition, heat flow and convective motions of the Earth's mantle, and the evolution of Earth's oxygen budget, the atmosphere, and life. The temporal patterns demonstrate that the proportion of ore deposits associated with volcanic activity was greater in the Archean than in more recent geologic eras. Sediment-hosted and granitic crust-controlled mineralizations became increasingly important during the second half of Earth's history as crustal (tectonic), biological, and atmospheric (oxidation) interactions took effect. Thus, volcanic-hosted porphyry copper, Cyprus-type (ophiolites), and Kuroko-type massive sulfides, as well as sediment-hosted Pb-Zn sulfide ores, are comparatively recent deposits and probably did not evolve on Mars. Although sedimentary ore-types such as banded iron formations and placer gold-pyrite-uraninite conglomerates were prevalent during the Archean [2,3], it is the igneous-hosted sulfide ore deposits that are diagnostic of that era of early crustal development on Earth [1] and may be more relevant to Mars. Since the focus is on sulfide mineralization, cumulate chromite deposits which are found in mafic igneous rocks spanning the whole range of geologic time (and may also exist on Mars because they are observed in some SNC meteorites [10,11]) are not considered further.

# CRUSTAL SULFIDE EVOLUTION

Burns, R. G. and Fisher, D. S.

**Sulfide Mineralization During the Archean.** Sulfide ores associated with mafic and ultramafic igneous rock-types [12] that were deposited in terrestrial rocks during the Archean include: (1) komatiitic (Kambalda-type) and tholeiitic (Noranda- or Abitibi-type) flows and intrusions in greenstone belts; (2) gabbroic intrusions either feeding flood basalt activity associated with intracontinental rift zones (e.g. Duluth) or occurring as layered complexes (e.g. Stillwater); and (3) norites intruding an astrobleme (Sudbury-type). The komatiitic suites, alone, are almost totally confined to the Archean era. Although they are associated in space and time with Noranda-type calc-alkaline suites, the latter volcanogenic massive sulfide deposits in a modified form are also found in younger igneous rocks [2,3]. So, too, are sulfides in layered intrusions (e.g. Bushveld), while the impact event producing sulfide ores at Sudbury 1.84 b.y. ago is also relatively young.

**Komatiitic Sulfide Deposits.** In Archean greenstone belts, older broad basal platforms of mafic and ultramafic komatiitic flows are overlaid by scattered subshields and stratovolcanic calc-alkaline volcanic rocks consisting of tholeiitic basalt and andesitic flows and pyroclastic deposits which are capped by dacitic to rhyolitic tuffs [13]. Repeated cycles built thick (~10 km) sequences of these volcanic deposits which were extruded in a continuously subsiding submarine environment. The komatiitic suites host sulfide mineralization consisting of pyrrhotite-pentlandite ( $\pm$  chalcopyrite, pyrite) deposits which occur as stratabound lenses or sheets a few meters thick at the base of magnesian ultramafic lava flows. These Fe-Ni sulfides appear to have formed by gravitational separation of an immiscible sulfide liquid from the mafic or ultramafic magma. Petrogenic models based on experimental studies [14] indicate that komatiitic magmas may represent large-percentage partial melts of the mantle, requiring very high extrusion temperatures (1425-1650 C, [12,14]) and suggesting diapiric emplacement from depths in excess of 200-400 km [14,15]. Such magma erupting to the Earth's surface could transport high concentrations of sulfur, either dissolved in the magma, or as immiscible FeS melts. Experimental measurements [16,17] have demonstrated that the solubility of sulfur increases with rising temperature, pressure, FeO content, and sulfur fugacity (but decreasing oxygen fugacity) of silicate melts. Temperature, in particular, magnifies the capacity of a melt to dissolve sulfur by a factor of 5 to 7 times per 100 degree increment [16]. Thus, an experimental komatiitic melt accomodating 0.1 wt % S at 1200 C indicates that several percent of dissolved sulfide could be transported to the surface at extrusion temperatures believed to be in the range 1400-1600 C, particularly during the Archean when geothermal gradients were significantly higher than they are now.

The Noranda-type massive sulfides occurring in overlying volcanogenic deposits associated with komatiites are stratabound lenticular bodies of pyrite mineralization containing variable amounts of chalcopyrite and sphalerite [1]. These Fe-Cu-Zn sulfides were deposited on and below the seafloor by hydrothermal solutions quenched by seawater. They differ from more recent Cyprus-type and present-day spreading-center ore bodies in their tectonic setting and by containing only minor Pb as galena.

**Archean Crustal Evolution.** Two mechanisms have been proposed for triggering the komatiitic peridotite-basaltic volcanism during the Archean [1,18]: first, sustained bombardment of the surface by meteorites which enabled the thinly crust-covered mantle to be tapped producing lunar maria-type volcanism [18]; and second, magmatism resulted from the higher heat flow then, inferred from geochemical and radioactive decay data. Turbulent processes in the mantle probably resulted in smaller, more abundant convective cells than in later times. Initial volcanic activity characterized by ultramafic flows and Fe-Ni sulfide mineralization may have been initiated by breakage of the thin Archean crust, which established zones of high heat flow and set up partial melting of the upper mantle to provide basaltic magma. As foundering continued, the basaltic crust itself was recycled by partial melting, and andesite to rhyolite magmas were generated, along with Fe-Cu-Zn-S components for the ultimate massive sulfide ore-bodies. Thus, vertical tectonics involving the sinking of simatic lithospheric slabs accompanied by diapiric upwelling led to a succession of divergent magma types and ore deposits dominated by pyrrhotite-pentlandite and pyrite-chalcopyrite-sphalerite assemblages found in Archean greenstone belts.

## CRUSTAL SULFIDE EVOLUTION

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**Applications to Mars.** Direct evidence for sulfide mineralization on Mars stems from the Viking XRF experiment, which analysed ~3 wt % sulfur in the regolith [19]. Although some of this sulfur may have originated from volcanic exhalates, a significant portion was probably derived from oxidative weathering of iron sulfide minerals [5] associated with basaltic rocks on Mars. Indirect evidence for the presence of sulfides in near-surface igneous rocks on Mars comes from petrographic, geochemical and textural studies of SNC meteorites, which have established strong analogies to terrestrial Archean magmas, particularly komatiites [7,8] and hence pyrrhotite-pentlandite mineralization [5,9]. Assuming that SNC meteorites did originate from Mars, the crystal cumulates in them indicate that most of the martian igneous rocks underwent magmatic differentiation, particularly the ultramafic units, products of which may include Fe-rich gabbro, pyroxenite, peridotite and massive sulfide deposits [7]. Since the Fe/(Fe + Mg) ratio of Mars' mantle is probably higher than that of the Earth's mantle [20,21], partial melting may have produced iron-rich komatiitic magmas which extruded to the surface of Mars at lower temperatures than terrestrial Archean komatiites. Although increased sulfur solubility in such Fe-rich magmas is compensated somewhat by the lower liquidus temperatures, significant amounts of sulfur could have been transported to the martian surface in komatiitic melts to become segregated later as massive sulfide deposits, accounting for the paradox of high sulfur concentrations in martian fines [22]. Furthermore, although the modal mineralogy of SNC meteorites arriving on Earth represents <1 % Fe-Ni sulfides as pyrrhotite, Ni-bearing troilite, pentlandite, chalcopyrite, marcasite and pyrite [10,11,23-25], associated lenses of massive sulfides may exist on Mars and not have been sampled by the impact event(s) that produced those meteorites from the Tharsis region ~165-187 m.y. ago [26]. Furthermore, the young ages of SNC meteorites, crystallizing on Mars  $\leq 1.3$  b.y. ago [26], suggest that komatiitic magmatism persisted there beyond the Archean era on Earth, and that associated Fe-Ni sulfide mineralization although possibly more voluminous on Mars, has not evolved there to the extent that it has on Earth. The prolonged komatiitic volcanism may also have contributed to the Tharsis and Elysium domes on Mars. Erosion may not yet have exposed many of the pyrrhotite-pentlandite deposits there, particularly if gravitation settling in magma chambers segregated the ores well below the surface of Mars [27].

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## TOWARDS A CHRONOLOGY OF COMPRESSIVE TECTONICS ON MARS. A. F. Chicarro, Space Science Department, ESA, ESTEC, 2200 AG Noordwijk, The Netherlands.

Tectonic features on Mars have been extensively studied in recent years, in order to understand the geological evolution of the planet's surface and interior. Although early studies were devoted primarily to extensional features associated with the Tharsis (1) and Elysium (2) bulges, the planet-wide distribution of ridges (3) shows a much more complex picture. Deciphering the chronology, extent, and intensity of ridge formation episodes is considered of the utmost importance in understanding the global geological evolution of Mars. This paper presents some preliminary interpretations.

**Distribution of Martian ridges:** Ridges on Mars are most easily recognizable on the smooth plains of presumed volcanic origin, thus named 'ridged plains', such as Lunae Planum, Hesperia Planum, Syrtis Major Planitia, and on the 'old volcanic material' West of the Hellas basin (4). However, the vast majority of the many thousands of mapped and digitized ridges are located in the old terrains ('cratered plateau', 'hilly and cratered material', and 'intercrater smooth plains') of the Southern Hemisphere (5), where they exhibit a less pronounced morphology. Also, Martian ridges are found, although in lesser numbers, on the younger plains ('cratered plains' and 'rolling plains') of the Northern Hemisphere.

Selective retrieval of given data subsets from the overall computer-based data set, according to ridge morphology and/or geological environment, allows further analysis of ridge distribution. A cross-comparison of both ridged plains features (Fig.1) and cratered highlands structures (Fig.2) with high relief (over 500 m high) ridges (Fig.3) and low relief (under 500 m) ridges (Fig.4), reveals that while most ridged plains trends belong to the high relief group, there is a significant density of high relief structures both in the old terrains (such as in Phaethontis and Eridania quadrangles) and in the younger plains (such as in Cebrenia). Therefore, major ridge formation episodes have affected three broad types of materials: ridged plains, cratered highlands, and northern younger plains. Conversely, low relief structures can be found over the entire planet, with the noteworthy exceptions of the most recent volcanic units of Tharsis, the polar terrains, and the lightly cratered northern plains. Thus, ridge formation activity ceased before the formation of these most recent units.

**Geologic history:** Recent stratigraphic studies on the surface history of Mars (6) reveal that tectonism and volcanism were widespread during Noachian and Lower Hesperian times, while they became more localized during Upper Hesperian and Amazonian epochs. The oldest observed Martian ridges are found on Middle and Upper Noachian terrains, and were formed after the final heavy bombardment, when large mass anomalies (basins) were being compensated isostatically due to the elasticity of the lithosphere (7). Major ridge trends (over 1 km high) are located in the Copernicus-Newton area, extending southward for several thousand km. While some uncertainties remain concerning the beginning of this ridge formation period, structural directions of ridges located on Middle and Upper Noachian terrains are similar. Therefore, the Upper Noachian was a major ridge formation episode, but probably not all ridges survived the intense degradation that preceded the formation of the ridged plains in the Lower Hesperian (6). Crater counting ages indicate (8) that ridges in Lunae Planum, as in most other ridge plains regions, are of about the same age, or slightly younger, than the plains materials where they are found. Most ridged plains tectonic structures are of Lower Hesperian age, when fluvial activity was significant and the atmosphere was thicker and warmer. Volcanic flows from localized emission centers overlaid Lower Hesperian ridge plains (9). The Upper Hesperian is a period when basically no ridge formation has been recorded. Outflow channel activity, as well as the formation of chaotic terrain by collapse due to the withdrawal of subsurface volatiles, has occurred in the Upper Hesperian (6). Again, in the Lower Amazonian, moderate ridge formation has been observed, in particular in the smooth cratered plains of Amazonis, Arcadia and Acidalia. This is a time dominated by eolian processes and subfreezing temperatures at the surface (10). Gravity anomalies exist over younger masses (e.g. Tharsis), due to a thickness and rigidity increase of the crust (11).

**Mechanisms of ridge formation:** Like on the Moon and Mercury, ridges on Mars are considered structures of tectonic origin, essentially caused by compressive stresses (12). Recent analysis of lunar ridges (13) suggests that the compressive stress fields responsible for ridge formation at or near the surface, have in fact been generated by faulting at depth and do not exceed the shear strength of the surface material. Golombek et al. (14) have shown that regional elevation changes at the surface reflect shortening at depth, and the average shortening by low-angle thrust faults at depth exceeds the shortening by folding of overlying surface units by a factor of 7. Terrestrial analog studies tend to confirm these findings (15). Also, the scarcity of deformed craters on Mars indicates that small compressive stresses may be sufficient for ridge formation. However, Martian ridge distribution and orientations have been influenced by the stresses related to the formation of impact basins (16), in addition to regional tectonic control like Tharsis (17). The dual expression of tectonic compressive stresses on Mars (thrust faults at depth and folding of anisotropic layers at the surface) indicates different rheological properties for surface and depth materials. Certain materials, when compressed, may fold easier (e.g. volatile-rich upper layers) than others (e.g. dry bedrock), which might fault preferably. A comparison of ridge and fluidized crater distribution (18) based on Viking images, clearly shows that all areas with a high density of fluidized craters display a significant number of ridges. Fluidized crater ejecta morphology seems to indicate that the target material was rich in volatiles at the time of the impact (19). However, the relative importance of water-ice (20), liquid water and also atmospheric effects (21) is still debatable. Therefore, the presence of subsurface volatiles contributed to form a relatively low viscosity material, where compressive stresses are best expressed as ridges. Ridge length, such as in Sinai Planum or Hesperia Planum, seems to be related to the duration of the low viscosity state (22), which also controls the density of fluidized craters. We propose that the smooth ridge plains target material was not in a permanent frozen state at the time of ridge formation, prior to fluidized crater impacting, and that the amount of subsurface liquid water was significant, possibly a dominant trend, due to Lower Hesperian climatic and atmospheric conditions (6). Therefore, compressive tectonic effects have been 'amplified' at a time and in areas of low viscosity and high volatile content (possibly in a liquid phase), such as in the Lower Hesperian, when most high relief ridges, but by no means all, were formed.

**Analysis of compressive tectonic episodes:** A detailed structural analysis of Martian ridge patterns (3) allows us to draw some conclusions, based on rose diagrams by 10° bins for the entire ridge data set and for some specific subsets, like: high/low relief ridges and ridge plains/cratered highlands structures. Major ridges

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in the Copernicus-Newton area display trends that cross different ancient units. However, ridge systems throughout the cratered highlands exhibit complex patterns and morphologies that reflect regional trends superimposed on locally modified stress fields, including the influence of impact basins (23). High relief ridges tend to have preferred orientations, while low relief ridges tend to have a broader fan-like pattern. In particular, high relief trends are quite different from low relief ones in Amethes, Elysium and Cebrenia, suggesting that ridges located in Upper Amazonian northern cratered and rolling plains were formed at a different age than the global low relief ridge network. In Lunae Planum and other ridge plains regions, both high and low relief ridges have similar trends, and thus the direction of compression was the same during the formation of both types of ridges, many of which may be contemporaneous indeed. Also, it could be argued that high relief ridges were formed in volatile-rich terrains because these units did 'amplify' the effects of compressive stresses. The old cratered highlands typically display a NW trend (for both high and low relief ridges), primarily in the equatorial region, while the ridged plains regions show additional preferred trends. These differences suggest changes in compressive stress directions with time, namely between the Upper Noachian and the Lower Hesperian.

**Chronology of Martian ridge formation:** Although a definite assessment of Martian compressive tectonics chronology proves difficult, the present study allows us to propose some working hypothesis based on the most recent data and analysis. We infer 3 periods of ridge formation: Upper Noachian (UN), Lower Hesperian (LH) and Lower Amazonian (IA).

- \* UN: probably planet-wide, dominant NW trend, numerous low relief ridges, some very high relief ridges, duration of compression uncertain.
- \* LH: localized ridge plains, directions influenced by old impact basins and regional uplifts, numerous high relief ridges, relatively long ridges, long duration of compression, stresses 'amplified' by the presence of subsurface volatiles close to freezing temperature at the end of this period.
- \* IA: localized on rolling and cratered plains, dominant N trend, both high and low relief ridges, duration of compression uncertain, tiny ridges found to crosscut young grabens, numerous ridges might have been buried underneath the youngest northern plains.

Based on Hartmann's model (24), we assume the first ridge formation period starting at about -3.85 billion years, and the last period ending at about -1 billion years. Therefore, ridge formation on Mars occurred after the final heavy bombardment and most of basin readjustments, and before or during Tharsis influence. Mars compressive tectonics contrast in style, distribution, timing, basin control and rheological properties of the deformed material, with the Moon and Mercury. Further studies should help to determine the relative roles of global shrinking, despinning (25) and polar wandering (26), during Martian tectonic history.

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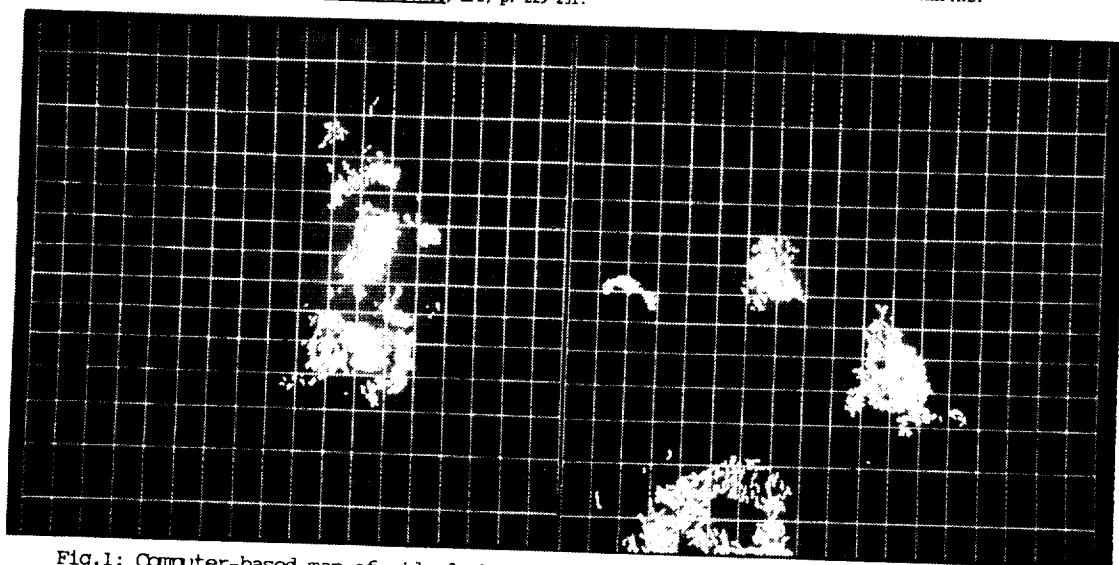


Fig.1: Computer-based map of ridged plains structures on Mars, between latitudes of  $\pm 65^\circ$ .



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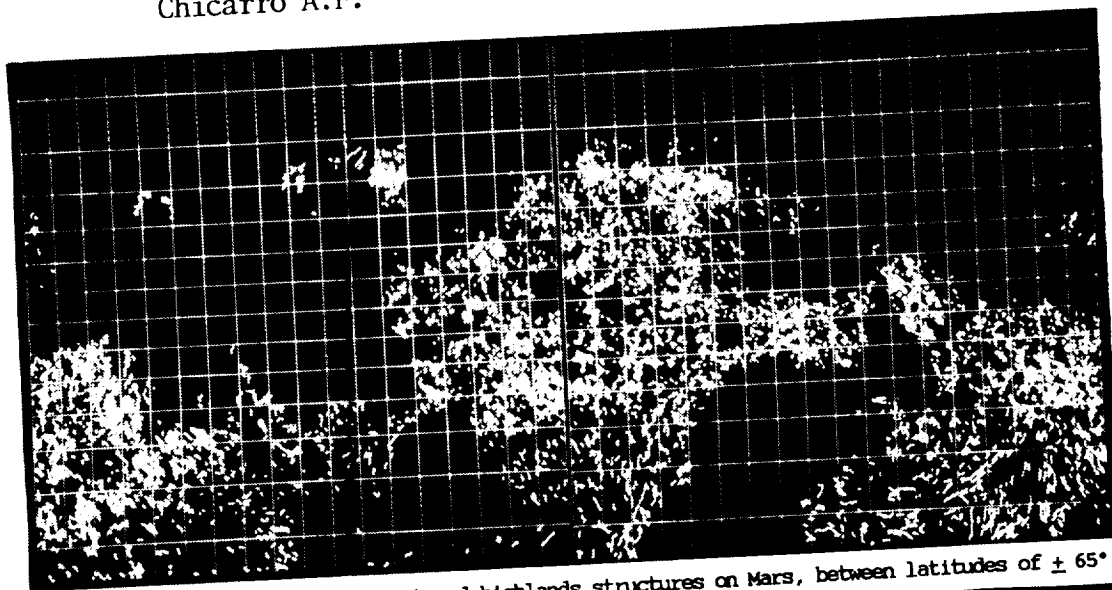


Fig.2: Computer-based map of cratered highlands structures on Mars, between latitudes of  $\pm 65^\circ$ .

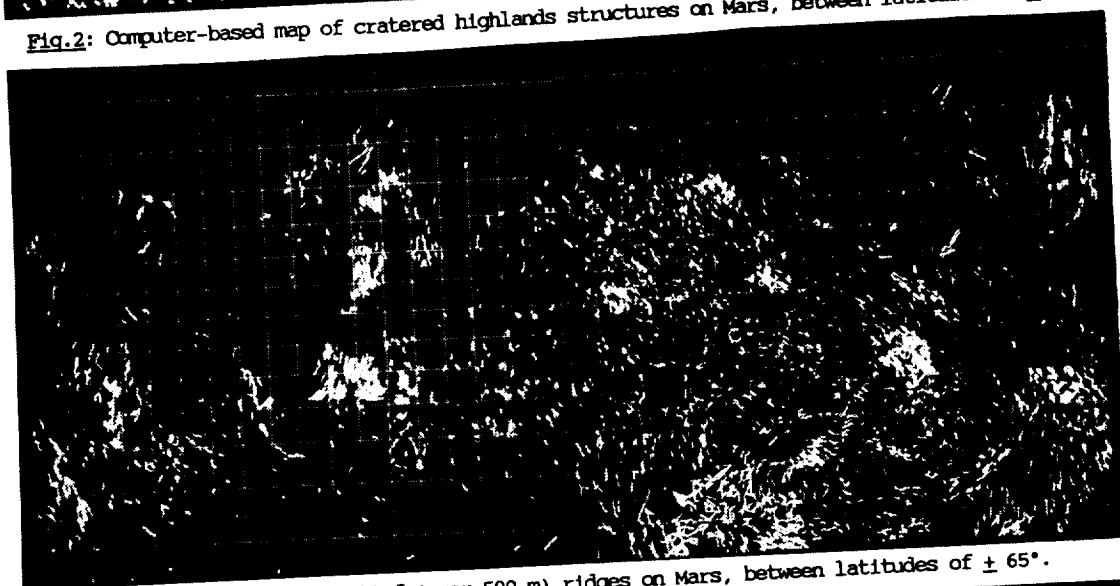


Fig.3: Map of high relief (over 500 m) ridges on Mars, between latitudes of  $\pm 65^\circ$ .

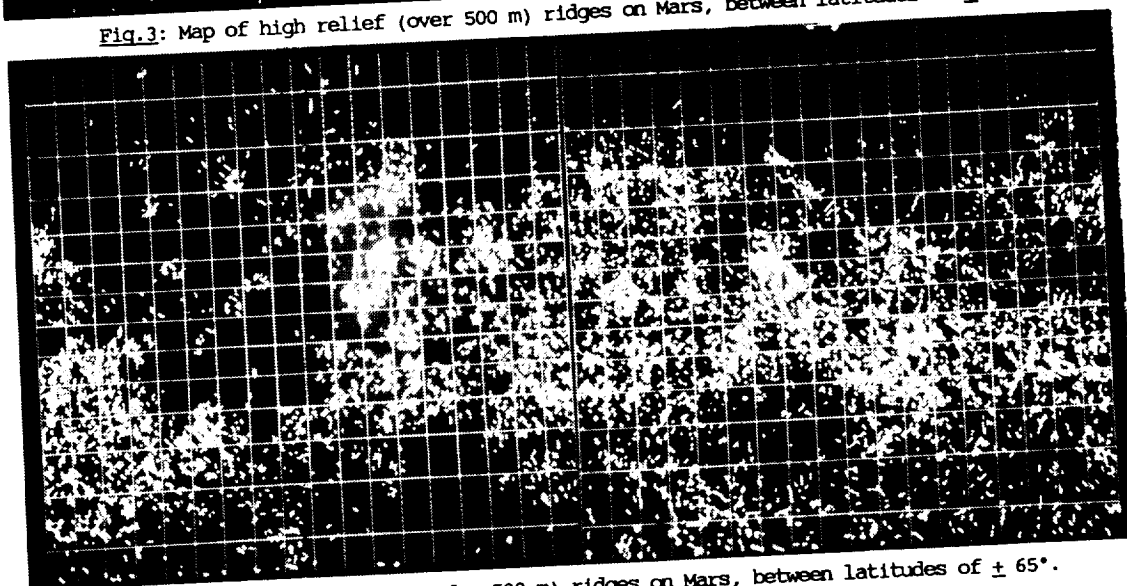


Fig.4 Map of low relief (under 500 m) ridges on Mars, between latitudes of  $\pm 65^\circ$ .

ORIGIN AND AGE OF GROOVED FEATURES IN THE MEMNONIA QUADRANGLE (MC-16), MARS. Robert A. Craddock\*, Ronald Greeley, and Philip R. Christensen, Department of Geology, Arizona State University, Tempe, Arizona 85287. \*Current address: Center for Earth and Planetary Studies, National Air and Space Museum, Smithsonian Institution, Washington, D.C. 20560

One of the most interesting features occurring in the martian highlands of the Memnonia quadrangle (MC-16) are parallel grooves, or grooved features. These features typically occur in clustered sets of a dozen or more, are spaced 1 to 10 km apart, are >5-km-long, and are commonly oriented west to northwest (oblique to Memnonia Fossae). As part of the Mars Geologic Mapping Program and investigations into the southern Mangala Valles area, the objective of this project was to assess possible origins and ages for these features. The approach taken was to first classify the grooved features and separate them from other similar martian features based on their morphology. Secondly their distribution and orientations within the Memnonia quadrangle was mapped, and then these results were investigated.

Investigations into the morphology of the grooved features show that they are similar to lunar Imbrium Sculpture materials, which occur radial to the Imbrium Basin on the moon and probably represent scouring by low-angle projectiles or are bulk materials emplaced ballistically or through base-surges (1). The martian grooved features commonly occur as long narrow troughs (e.g., Viking orbiter photograph 637A84), but infrequently they also occur as a series of sharp crested ridges (e.g., Viking orbiter photograph 457A33). Both morphologies are similar to those typical of Imbrium Sculpture (e.g., Lunar orbiter photograph IV-090-H2). Grooved features are distinguishable from martian yardangs, mass-wasting features, tectonic features, and ancient valley networks, all of which similarly occur as linear features: yardangs in Amazonis Planitia occur in two perpendicular directions (2), unlike the range of orientations observed for the grooved features in this study. Yardangs are also typically streamlined in shape as opposed to the straight form of the grooved features. Mass-wasting features are highly dependent upon slope (3), whereas grooved features tend to cut across topographic boundaries and also occur on flat surfaces. Tectonic features, common in the Memnonia quadrangle, are also typically much longer than any of the observed grooved features. Extensionally derived Memnonia faults (e.g., 4) also have characteristic grabens, unlike the grooved features. Ancient valley networks consist of anastomosing "channels" which converge to form dendritic patterns (5), unlike the linear structure of the grooved features. The assumption made, therefore, is that the grooved features are materials similar to Imbrium Sculpture and relate to some impact event.

Mapping results show that the grooved features are widespread in the Memnonia quadrangle (Table 1). They are also found only on the older Hesperian/Noachian highland materials, and in places are embayed by Amazonian age Tharsis volcanics. Because a wider range in orientation of ejecta would occur closer to the impact site as opposed to further away, the wide range of grooved feature orientations indicates a local source. Also, lunar mapping of the Imbrium Sculpture (6) shows that this material occurs up to ~1,700 km away from the 670 km diameter Imbrium Basin (the Memnonia quadrangle is ~3,000 km across by comparison). Because of the higher martian gravity and atmospheric drag induced on ejecta, material

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from an Imbrium-size impact on Mars should be closer to the crater (7). The combination of these data suggests that the grooved feature source area should be relatively close to the Memnonia quadrangle.

Determination of potential source areas was accomplished by plotting the orientation of the grooved features on a stereo net (i.e., Wulff net). Observations made during the geologic mapping suggested that not all of the grooved features had ambiguous source areas. Memnonia craters Re, Sn, and Yn each had associated local grooved features (Table 1), and stereo net plots confirmed to within 1.0° that these craters were the source areas. Plotting the orientations of the remaining grooved features on a stereo net shows two areas where the lines converge or intersect (Fig. 1). The first such area occurs at ~10.0°, 170.0° in southern Amazonis Planitia ("A" in Fig. 1), and it may be that the circular occurrence of knobby materials in this area (e.g., apparent in 8) represent the rim of an ancient impact structure that has been buried by Elysium volcanics. The determination of the second potential source area was aided by additional data from the Phaethontis quadrangle (MC-24), suggesting a possible basin at -30.0°, 125.0° in Daedalia Planum ("B" in Fig. 1) which may have been buried by Tharsis volcanics. The circular nature of Daedalia Planum (e.g., apparent in 9) may define this basin.

Because Elysium and Tharsis volcanics are both Amazonian in age (e.g., 10), and because the mapping results showed that the grooved features occur on Hesperian/Noachian age materials, the grooved features and their associated basins must be at least Hesperian in age.

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TABLE 1.

Lat.	Long.	Orientation	Lat.	Long.	Orientation
-21.5	145.0	N65W*	-17.5	171.0	N106W
-20.5	147.6	N66W	-14.8	177.0	N42W
-40.0	122.4	N19E	-19.5	170.5	N56W
-13.0	147.0	N56W*	-19.5	171.4	N56W
-14.0	147.9	N56W*	-21.4	175.8	N29E*
-13.5	145.0	N56W*	-26.8	179.0	N20E
-40.0	123.0	N19E*	-14.0	174.8	N41W
-28.6	137.5	N69W*	-16.4	161.6	N73W*
-28.0	156.5	N56W*	-26.4	160.2	N70W (Re)
-22.5	141.0	N66W*	-24.5	159.0	N69W
-28.4	139.5	N75W	-13.0	164.8	N110W (Sn)
-20.5	152.8	N64W*	-13.7	162.6	N108W (Sn)
-21.0	151.2	N63W	-15.4	161.4	N90W (Sn)
-22.0	152.3	N72W*	-9.0	162.0	N95W
-28.5	153.0	N85W	-28.1	139.9	N25E
-9.5	156.0	N72W	-25.5	166.5	N82W
-11.5	155.4	N47W	-24.0	165.8	N75W
-8.8	146.5	N40W	-22.5	164.0	N76W
-7.0	149.0	N47W	-5.5	156.0	N41W
-7.0	147.0	N62W	-22.5	170.9	N100W (Re)
-14.6	177.2	N101W (Yn)	-14.8	176.2	N30W (Re)

(Letters in parentheses indicate associated Memnonia craters.)

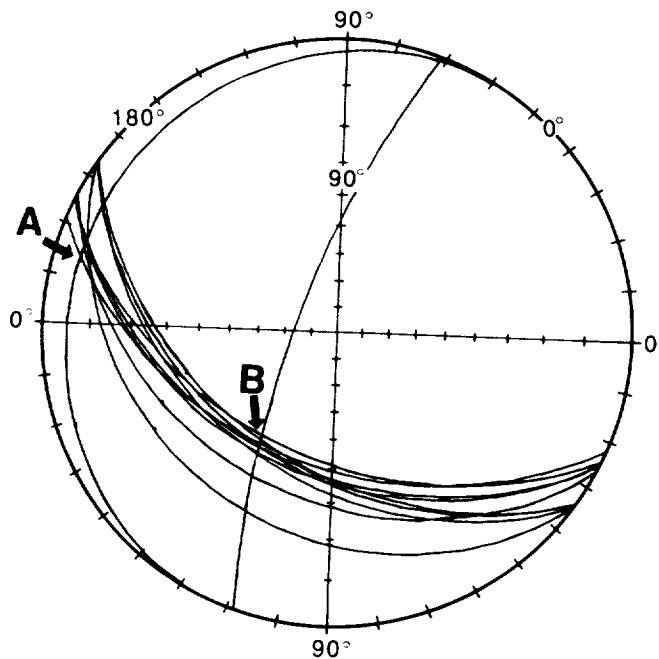


Figure 1. Stereo net (Wulff net) projection of Mars centered at  $0.0^\circ$ ,  $90.0^\circ$ . Lines show trend of grooved features denoted by an asterisk in Table 1. "A" represents possible source area in Amazonis Planitia; "B" represents possible source area in southern Daedalia Planum. The Memnonia quadrangle is to the lower left of the equator ( $0.0^\circ$  to  $-30.0^\circ$  Lat.,  $135.0^\circ$  to  $180.0^\circ$  Long.).

**THE MARTIAN HIGHLAND PATERAE: EVIDENCE FOR EXPLOSIVE VOLCANISM ON MARS; David A. Crown and Ronald Greeley, Department of Geology, Arizona State University, Tempe, Arizona 85287**

The martian surface exhibits numerous volcanic landforms displaying great diversity in size, age, and morphology [e.g. 1]. Most research regarding martian volcanology has centered around effusive basaltic volcanism, including analyses of individual lava flows, extensive lava plains, and large shield volcanoes. Various eruption mechanisms have been considered theoretically under martian conditions [2,3], and recently a number of investigations have focused on the existence of explosive volcanism on Mars [3-7]. These studies have been hindered by a lack of definitive morphologic criteria for the remote identification of ash deposits. Knowledge of the abundances, ages, and geologic settings of explosive volcanic deposits on Mars is essential to a comprehensive understanding of the evolution of the martian surface, with implications for the evolution of the lithosphere and atmosphere as well as the histories of specific volcanic centers and provinces.

The presence of small-scale or localized explosive volcanic activity is suggested by a variety of features. Cinder cones [8,9] and "pseudo craters" resembling those in Iceland [10] have been identified in the northern plains and in the Tharsis and Elysium provinces on Mars. A smooth, mantled region on Hecates Tholus has been attributed to an airfall deposit resulting from a plinian eruption [3], and volcanic density currents have been proposed to account for channels associated with volcanoes in the Tharsis and Elysium regions [11]. A geomorphic comparison of the Elysium province on Mars with the Tibesti region on Earth reveals similarities between Elysium Mons and terrestrial composite volcanoes [12].

The existence of large-scale explosive volcanic deposits on Mars is controversial. Although the basal scarp [13] and aureole deposits [14] of Olympus Mons and extensive deposits in the Amazonis, Memnonia, and Aeolis regions have been attributed to pyroclastic origins, other hypotheses are implied by the observed morphologies [15-16]. Recently, morphologic evidence has been employed to suggest that pyroclastic deposits are associated with Alba Patera and most likely were emplaced as pyroclastic flows fed by ash fountain eruptions [4,5].

The martian highland paterae (Apollinaris Patera, Amphitrites Patera, Hadriaca Patera, and Tyrrhena Patera) are areally extensive, low-relief features with central caldera complexes and radial channels separating plateau-like erosional remnants [1,17]. The highland paterae, which formed in Upper Noachian to Lower Hesperian time, were initially proposed to be the consequence of eruptions of extremely fluid lavas [18]; however, currently they are believed to be composed predominantly of ash deposits on the basis of morphologic similarities to terrestrial ash sheets [19] and their apparently easily erodible nature [1]. Greeley and Spudis [1], from an analysis of Tyrrhena Patera, proposed an evolutionary sequence for the paterae beginning with extensive pyroclastic eruptions due to the contact of rising magma with the water- or ice-saturated megaregolith followed by erosion of the ash deposits and late stage effusive eruptions with deposits concentrated near the summit.

Hydromagmatic origins for Hadriaca Patera and Tyrrhena Patera have been evaluated quantitatively [6]. Hadriaca Patera is a large, asymmetric volcano with channeled flanks and a central caldera complex filled with smooth plains displaying wrinkle ridges (Figure 1) [20]. Tyrrhena Patera is irregular in plan view and is composed of a lower shield unit of dissected material and a smooth, upper undissected unit containing wrinkle ridges (Figure 2) [1]. Also associated with the late stage summit activity at Tyrrhena Patera are several sinuous rille-like features. The volumes of Hadriaca Patera and Tyrrhena Patera have been estimated from the mapped boundaries of the volcanoes and topographic data from the *1:15M Topographic Map of Mars Eastern Region*. Assuming a density of  $1500 \text{ kg/m}^3$  in the deposits, masses for the two volcanoes can be calculated and used to estimate the initial thermal energy of the magmas (Table 1).

If it is assumed that the eruptions producing Hadriaca Patera and Tyrrhena Patera occurred near the present summit regions (and there is no evidence to the contrary), the dimensions of the volcanoes can be used to constrain possible eruption mechanisms. As is the case for Alba Patera, an air-fall origin for most of the deposits can be dismissed because eruption clouds with heights comparable to the total widths of the volcanoes are required [5]. Models for the emplacement of a

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gravity-driven flow resisted by a frictional force are used to determine the lengths of terrestrial pyroclastic flows for a given pre-flow topography [21]. The relationship between the flow length and the initial velocity of pyroclastic flows potentially associated with Hadriaca Patera and Tyrrhena Patera for a slope of  $0.25^\circ$  (comparable to the present surfaces of the volcanoes) and for values of the apparent coefficient of sliding friction,  $\mu$ , equal to 0.05 and 0.10 are shown in Figure 3. Large volume terrestrial pyroclastic flows have values of  $\mu = 0.06 - 0.20$  [22]. On Mars, comparable coefficients of friction should be less due to the lower gravity and less dense atmosphere. Figure 3 illustrates that the entire range of paterae flank widths can be produced for both values of  $\mu$  if only slightly greater than 10% of the initial thermal energy of the magma is converted into the kinetic energy of the pyroclastic flows. This is in agreement with experimental results which indicate a 1 - 10 % energy conversion in hydromagmatic eruptions [23].

Terrestrial magmatic eruptions feed pyroclastic flows either by ash fountaining or by eruption column collapse. To produce flows with lengths comparable to the entire range of flank widths of the paterae, a minimum initial velocity of  $\sim 350$  m/sec is necessary (for  $\mu = 0.05$ ) (Figure 3). For an eruption driven by magmatic water, this implies an exsolved magma volatile content of  $> 1\%$ , a mass eruption rate of  $> 10^7$  kg/sec, and eruption cloud height of nearly 70 km [Fig. 8 in 3]. The eruption rate and column height indicated are similar to those derived for the eruption at Hecates Tholus thought to have produced the mantling deposit seen near the summit.

From an energy perspective origins of Hadriaca Patera and Tyrrhena Patera can be explained by the emplacement of pyroclastic flows fed by eruptions driven by either magmatic or external water. Hydrovolcanic explosions would be favored in the near-surface environment on Mars as water in the megaregolith could come into contact with a rising magma body. Although the magmatic and hydromagmatic models both merit further attention, the formation of the paterae by hydromagmatic eruptions in only an early period of martian history is consistent with suggested global changes on Mars and could explain why this style of volcanism is not evident in later eras.

Table 1. Characteristics of Hadriaca Patera and Tyrrhena Patera

Volcano	Dimensions (km)	Edifice Height (km)	Flank Slopes	Caldera Diameter (km)	Volume ( $m^3$ ) $\times 10^{14}$	Mass <sup>#</sup> (kg) $\times 10^{17}$	Thermal* Energy ( $kg\ m^2/sec^2$ ) $\times 10^{23}$
Hadriaca	288 x 570	~3	0.05 - 0.60°	63 x 78	1.867	2.801	3.361
Tyrrhena	426 x 660	2+	0.16 - 0.28°	45	1.176	1.764	2.117

#assumes  $\rho = 1500\ kg/m^3$  in deposits      \*  $1\ kg\ m^2/sec^2 = 10^7\ ergs$

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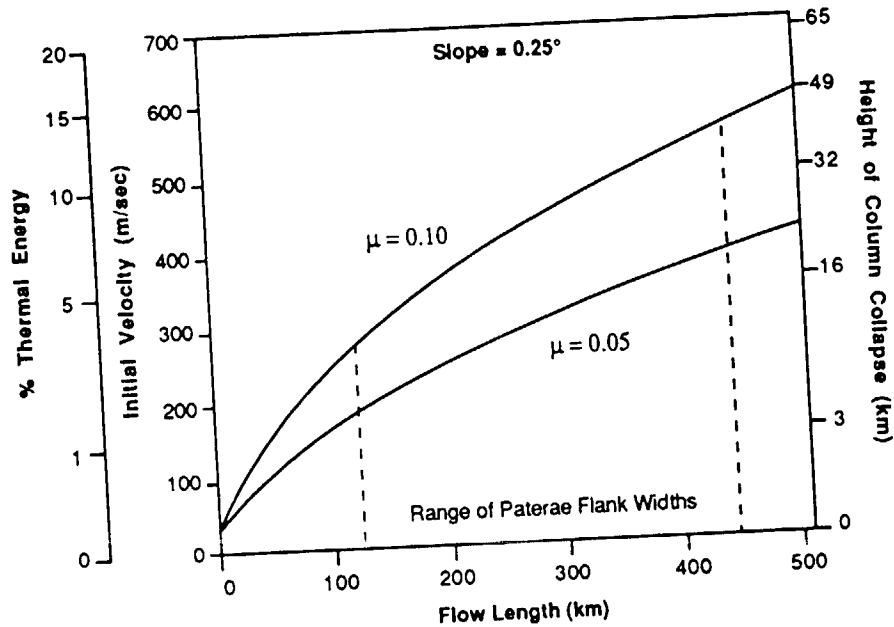


Figure 3.

Figure 1. Viking mosaic of Hadriaca Patera showing (a) dissected flanks and (b) central caldera complex. A large outflow channel and several irregular depressions are evident to the south and west.

Figure 2. Viking image (087A14) of Tyrrhena Patera showing (a) dissected lower shield materials, (b) smooth upper shield materials, (c) central caldera complex, and (d) sinuous rille-like features.

Figure 3. Relationship between flow length and initial velocity for martian pyroclastic flows over a 0.25° slope for coefficients of friction equal to 0.05 and 0.10 [6]. The initial velocity is correlated with the % of the available initial thermal energy of the magma and with a corresponding height of column collapse (by conversion of kinetic energy to potential energy). The observed range of flank widths for Hadriaca Patera and Tyrrhena Patera is indicated.

**Ejecta Deposits of Large Martian Impact Basins: A Useful Geologic Tool and Window to Early Martian History?** *Kenneth S. Edgett, Department of Geology, Arizona State University, Tempe, AZ 85281.*

The recognition of basin ejecta deposits is considered to be an important key to the interpretation of martian cratered highlands geology [1]. Because most of the basins are very ancient (Early and Middle Noachian [2]), they are important to understanding the nature and composition of the early martian crust. If recognized, basin ejecta deposits could be used as stratigraphic tools to correlate the timing of early events in the martian cratered highlands, as was done on the Moon [3,4] and Mercury [5,6]. Basin ejecta are also considered to be desirable materials to return to Earth for study, because they permit the sampling of deep, ancient crustal materials and can aid in the absolute dating of the martian stratigraphic record [7,8].

The main problem with the study of impact basin ejecta on Mars is that the deposits are generally not recognized through photogeologic study. Features characteristic of basin ejecta (radial valleys, troughs, grooves, and secondary crater chains) are not observed near most basins. Although King [8] has reported to the contrary, this is particularly true of Hellas Basin and the other large basins, Argyre and Isidis.

A survey of martian basins  $\geq 200$  kilometers in diameter has been completed, in order to assess the preservation state of their ejecta deposits. The very ancient and highly degraded basins described by Schultz *et al.* [9] are not included here, because they were treated in that earlier study. There are about 19 basins  $\geq 200$  km in diameter, excluding those of [9]. Three of these are Hellas, Isidis, and Argyre. For the most part, their ejecta are not recognized. A fourth is tentatively labeled "South Polar" (82.5°S, 267°W; 850 km diam.) [10], and it, too, shows no ejecta deposit. The remaining basins of concern in this study are those 200 - 500 km in diameter. Of these 15 basins, only two have significant observable ejecta deposits. One of these is the youngest basin on Mars, Lyot (50°N, 331°W; 200 km diam.), considered to be Early Amazonian in age [2]. The other basin is Herschel (14°S, 230°W; 300 km diam.), which has at least 95,000 km<sup>2</sup> of preserved continuous ejecta; and it is older (Middle Noachian) than Lyot [11]. The basins Newton (41°S, 157°W; 300 km diam.), Huygens (14°S, 304°W; 460 km diam.), and Schiaparelli (3°S, 344°W; 470 km diam. [1]) are the only others which definitely show the types of radial features which are landforms associated with the emplacement of ejecta. The other basins  $\geq 200$  km in diameter show a range of states of preservation, but most seem to lack ejecta material. Some notes on these basins are listed below.

Consequently, the use of martian basin ejecta as a photogeologic stratigraphic tool is impractical for two reasons. The first is that the basins are widely separated from each other [12]. Even in Arabia, where there are five basins with diameters  $\geq 200$  km, the continuous ejecta deposits could not overlap. Secondly, most of the ejecta deposits have been greatly modified by depositional and erosional processes [13, 14, 15]. Subsequent meteorite impacts and aeolian processes have probably dominated the erosion of ejecta. Fluvial or sapping processes have also operated (*eg.* radial channels in Herschel Basin ejecta [11]). Much of the ejecta of these basins has probably been *buried* by volcanic and aeolian deposits [13, 16, 17].

As a potential material for unmanned sample return, basin ejecta may be a poor candidate. There are three reasons for this: (1) basin ejecta deposits are largely buried or eroded, as stated above, (2) an ejecta deposit would not likely provide a smooth, safe site for robotic landers and rovers, and (3) because these basins are mostly very ancient (Early-Middle Noachian, 3.85 or 4.2 to 4.6 billion years old [2]), the ancient crustal rocks they have exposed will likely be chemically altered and weathered.

The potential for using martian basin ejecta deposits as windows to martian geologic history is not diminished, however. A careful study of the erosional and depositional processes which have acted upon them may reveal clues to local, global, and historical climatic conditions. Study of weathered samples of ejecta will provide additional insight to these issues, as well as information about the crustal composition and age. Careful field studies of buried ejecta deposits by geologists *on* the surface of Mars will eventually make it possible to correlate the timing of basin-forming events, and to study early crustal materials exposed in basin ejecta and within the basin walls and central peaks.



## IMPACT BASIN EJECTA

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**MARTIAN BASIN EJECTA MODIFICATION:  
PRELIMINARY NOTES FOR BASINS  $\geq 200$  KM,  $\leq 500$  KM  
(Except Lyot)**

- Schmidt (-72°, 79°) Diam. ~200 km [MC-30-D]  
 -some radial texture preserved  
 -appears to be buried by (layered?) mantling deposits  
 -also has several superimposed craters
- "Unnamed" (-42°, 123°) Diam. ~200 km [MC-24-NE]  
 -tectonically disturbed in NE  
 -volcanics in NE, SE  
 -channeled 'ejecta?' in SE
- Secchi (-58°, 258°) Diam. ~205 km [MC-28-SE]  
 -radial "fluvial" channels  
 -lies in heavily fractured Hellas Mts. material  
 -some smooth materials mantle area
- Kepler (-47°, 219°) Diam. ~210 km [MC-29-NC,SW]  
 -modified by impacts, some radial channels  
 -tectonic disturbances in SE and E?  
 -might be partly buried in N,NE  
 -possible secondary chain to NW
- Flaugergues (-19°, 341°) Diam. ~220 km [MC-20-SW,NW]  
 -some radial grooves, valleys  
 -burial and tectonic disturbance to SE  
 -ejecta difficult to recognize
- Galle (-51°, 31°) Diam. ~220 km [MC-26-SW,SE]  
 -located in Argyre Mts. material  
 -some radial features (ie. grooves to E)  
 -much ejecta may be eroded  
 -probably mantled, as well
- Koval'skiy (-30°, 141°) Diam. ~280 km [MC-24-NC, MC-16-SE]  
 -greatly modified by volcanic materials from S. Tharsis area
- Copernicus (-40°, 169°) Diam. ~290 km [MC-24-SW,NW]  
 -tectonically disturbed in NE  
 -no 'ejecta' visible?
- Herschel (-14°, 230°) Diam. ~300 km [MC-22-NE,SE]  
 -ejecta preserved in NW,N,NE,SE  
 -channels radial to basin  
 -possible burial ? in SW  
 -several superimposed impacts  
 -(see: Edgett *et al.*, 1988)
- Newton (-41°, 157°) Diam. ~300 km [MC-24-NW,NC]  
 -some radial features  
 -secondary chains to SE, N  
 -tectonic disturbance in W  
 -channels in E

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Cassini (24°, 328°) Diam. ~400 km [MC-12-NE]

- numerous superimposed impacts
- burial by volcanics (?) in SW,W,NE

Antoniadi (22°, 299°) Diam. ~400 km [MC-13-NW]

- buried by volcanics in SE
- disrupted by impact (Baldet Crater) in NE
- radial grooved to S,N
- S,W,N- hard to distinguish ejecta from surrounding terrain

Huygens (-14°, 304°) Diam. ~460 km [MC-21-NW,SW]

- numerous radial grooves, channels
- some burial by volcanics in N,NE,W?

Schiaparelli (-3°, 344°) Diam. ~470 [MC-20-NW, MC-12-SW]

- radial grooves ESE, SSW (some ejecta present?)
- burial in SE, SW, N, NW(?) by volcanics
- radial channels evident
- (see Mouginiis-Mark *et al.*, 1981)

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ORIGIN OF THE MARTIAN CRUSTAL DICHOTOMY; Herbert Frey and Richard A. Schultz, Geophysics Branch, NASA/Goddard Space Flight Center, Greenbelt, MD 20771

#### INTRODUCTION

The fundamental first order characteristic of the martian crust is its two-fold nature: old, heavily cratered highlands in the south, lower-lying, more sparsely cratered plains in the north. This crustal dichotomy is not unique among the terrestrial planets, but its origin on Mars remains controversial. Both endogenic and exogenic processes have been suggested. Wise et al. (1) originally proposed that the origin of the crustal dichotomy was linked with the origin of the Tharsis Rise: an early single-cell convection system was envisioned to have subcrustally eroded the northern one-third of the planet which subsequently foundered isostatically and was overplated with volcanic plains. Problems with this model include whether early convection actually occurred and was vigorous enough to erode the necessary thickness of crust (at least 10 km), whether a single-cell convection pattern could ever occur and persist (with the associated delay in core formation that would be required), and whether or not some large scale structural evidence of the process (beyond the lowered topography) should have persisted over martian history (2).

Wilhelms and Squyres (3) proposed the other extreme, suggesting the dichotomy was due to a giant impact which formed a 7700 km wide "Borealis Basin" in the northern one-third of Mars. It is clear that large impacts have played an important role in the evolution of the martian crust, but there is little direct evidence for an impact of the size proposed (2). The "rim" of the proposed basin is missing for one-half of the structure. The topography of the northern one-third of Mars is not basin-like (even allowing for subsequent relaxation of the original shape). In particular, there is high standing relic cratered terrain in the knobby material in Elysium-Amazons with crater retention age as old as that of the cratered terrain outside the proposed basin (4,5). Finally, if the Borealis Basin had occurred and were the largest member of a  $D^{-2}$  population of impact basins, then there remains to be found on Mars a very large number (>40) of large (>1000 km diameter) impact basins, many of which would have to be in the well-preserved cratered highlands (6).

Although the question of the origin of the crustal dichotomy on Mars is sometimes described as "endogenic or exogenic?", we do not believe this is an either/or problem. Large impacts are capable of rearranging the entire structure of the lithosphere and producing lowland topography. But large impacts may also trigger internal mantle responses and may concentrate long-lived thermal and tectonic processes (7,8,9,10). We suggest that the crustal dichotomy of Mars is the result of overlapping large (but not giant) impacts, many of which were concentrated in the northern one-third of Mars, which produced low-lying basin topography but also led to the prolonged volcanism which characterizes the northern plains and perhaps even to the formation of Elysium and Tharsis (6).

#### MISSING IMPACT BASINS ON MARS

The largest recognized impact basin on Mars (besides the proposed Borealis Basin) is the Chryse Basin (11). If the Chryse and Borealis Basins are each separately considered to be the largest member of a proposed original  $D^{-2}$  population, then very different populations are predicted (460 basins larger than 200 km for Chryse-largest, 1440 basins

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larger than 200 km for Borealis-largest). We compared the actual number of observed impact basins on Mars (37 or 38, depending on whether Borealis is "observed") with the number predicted in order to crudely estimate the number of "missing" basins which subsequent geologic processes have presumably removed (6). For Chryse-largest there could be 1 basin larger than 2000 km and 7 larger than 1000 km missing. The total area required to accommodate these missing basins is less than 30% the surface area of Mars. In principle (though unlikely in practice) all of these could be located in (and have helped to produce) the northern lowlands.

It is more difficult to imagine how the very much larger number of missing Borealis-largest basins (1 larger than 5000 km, 10 larger than 2000 km and 47 larger than 1000 km) could have been easily removed. The area required to accommodate these (allowing for overlap) is 80% the surface area of Mars. Many large impact basins would have to be located in the cratered highlands, where to date they have not been recognized. Alternatives to the mega-impact hypothesis should be examined.

Chryse may not be the largest impact basin on Mars but if it is, the D-2 distribution through this 4300 km diameter basin predicts at least 1 basin larger than 2000 km and 7 larger than 1000 km remain to be found. If Hellas is larger than generally assumed (12,13), these numbers may double. McGill (14) has suggested a large impact basin beneath the plains in Utopia. The diameter of the proposed Utopia Basin is at least 3000 and perhaps 4500 km and it overlaps both the Isidis and Elysium (12) impact basins. In addition to producing much of the low-lying topography in the northern plains in eastern Mars, the Utopia Basin may also help explain the existence of the Elysium volcanic complex. Both Elysium Mons and Hecates Tholus lie at the intersection of Utopia and Elysium Basin inner rings, which may be centers for prolonged volcanic activity (12). The high standing relic cratered terrain appearing as knobs and partly buried craters in Elysium-Amazonis generally lies outside the Utopia Basin rings where they intersect the Elysium Basin. Perhaps the formation of an Elysium Basin centered on the rim of a previously formed Utopia Basin provided the necessary conditions for volcanism inside the Utopia Basin while allowing preservation (to a degree) of cratered terrain outside.

The distribution of knobs and detached plateaus and topography in the northern plains is more compatible with a number of impact basins than with a single basin (15). A closed circular depression 1000-1500 km across and 3 km deep lies at 60°N, 30°W. Scattered knobs define the southwest and southeast boundaries of this depression which overlaps the northern edge of the Chryse Basin. A semi-circular distribution of knobs of Noachian-Hesperian age (16) in Amazonis at 60°N, 168°W has a diameter of 700-900 km and may intersect an outer ring of the large Elysium Basin as well as the smaller Mangala Basin at 0°N, 147°W. The 600 km wide arc of masifs centered at 37°N, 167°W may be an inner ring of a larger structure which would overlap the inner rings of the Elysium Basin.

Overlapping large impacts maybe more effective than a single giant impact in producing the variable topographic structure of the northern lowlands and the different styles of volcanism found there. The mechanical thinning of the lithosphere would be greater at the overlapping event. Greater fracturing of the deep lithosphere might produce better access to the surface for magma, leading to prolonged emplacement of volcanic materials. Because the thermal decay time for very large impacts is  $10^7$  to  $10^8$  years (7,9,10), overlapping impacts will likely have significantly

enhanced temperatures at depth for greatly extended periods as the impact heat for the second event adds to that remaining from the first. Given the observational evidence for old impact basins in their vicinities, it is reasonable to ask whether Elysium and even Tharsis owe their existence to the clustering (and timing?) of large impact basins.

#### AGE AND EVOLUTION OF THE CRUSTAL DICHOTOMY

It is sometimes assumed that the crustal dichotomy predates nearly all observable martian history, but this may not be the case. Wise et al. (1) set the crater retention age  $N(1)$  (cumulative number larger than 1 km diameter per  $10^6 \text{ km}^2$ ) for the lowering and resurfacing of the northern one-third of Mars at  $N(1) = 100,000$  to  $50,000$  with the formation of highland cratered terrain at  $N(1) = 400,000$  to  $1,000,000$ . Detailed study of the resurfacing history of the highland/lowland boundary (transition zone), the adjacent cratered terrain and smooth plains (5) and other areas around and within the northern lowlands (4,17) places tighter constraints on these ages. The oldest age we find lies in the cratered terrain south of the transition zone, the cratered terrain in Tempe, and best preserved knobby terrain in Elysium and has an age of  $N(1) = 250,000$ , consistent with that of Wise et al. (1). No relic of early cratered terrain exists within the transition zone, a consequence of the efficiency of later resurfacing events. The first of these, which ended at  $N(1) \sim 85,000$  in the cratered terrain, knobby terrain and in Tempe, is also recorded in the transition zone. But the diameter range over which the event is marked includes larger craters (up to 80 or 100 km diameter) in the transition zone than in the adjacent cratered terrain ( $< 40$  km). This suggests that the 85,000 age resurfacing, which we associate with the formation of cratered plateau material, occurred after some structural difference between the transition zone and cratered terrain had already been established. Lowered topography in the transition zone is likely.

A subsequent resurfacing at  $N(1) \sim 25,000$  was even more widespread and was coeval with the emplacement of ridged plains in Lunae Planum, Tempe and elsewhere (16,17). Estimates of the thickness of the resurfacing materials, using the smallest surviving crater which defines the ages of the different crater counting surfaces, show that the Lunae Planum Age materials were significantly thicker in the transition zone west (220 to 325m) than east of Isidis (135 to 210m). In cratered terrain in Tempe and adjacent to the transition zone this layer is only 100m thick or less. Topography again likely plays a role, and the larger number of middle size impact basins along the transition zone west of Isidis (Nilosyrtis Mensae Basin, basin south of Renaudot, basin south of Lyot, Deuteronilus A and B) by comparison with those east of Isidis (Al Qahira, basin south of Hephaestus Fossae) may be important. Subsequent resurfacing appears to always be more efficient (involving greater thickness of material) west of Isidis (5). The likely topographic control on these events is much easier to understand if the origin and evolution of the crustal dichotomy were a consequence of a series of events, including multiple impacts.

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EARLY RESURFACING EVENTS ON MARS; H. Frey, J. Semeniuk, and T. Grant, Geophysics Branch, NASA/Goddard Space Flight Center, Greenbelt, MD 20771

## INTRODUCTION

The style of geologic activity on Mars changed at the Lower Hesperian/Upper Hesperian boundary in Tanaka's (1) stratigraphy from planetwide to progressively more localized at a few major centers (Tharsis, Elysium, the northern lowlands). Two planetwide resurfacing events were particularly important in early martian history. The first, in the Late Noachian, is recorded in the intercrater plains. The second, involving the emplacement of ridged plains, occurred in the Early Hesperian.

Major resurfacing events produce a depopulation of impact craters that will appear as a departure from a standard production curve in a cumulative frequency plot. Neukum and Hiller (2) developed a technique to identify such events by breaking the cumulative frequency curve into separate branches at these departure points and determining the crater retention age for both the older, underlying and newer, overlying surfaces. We have used this approach to identify major resurfacing events at widely different locations on Mars (3,4,5,6). Because new branches are determined by subtracting away older (survivor) craters, this method has advantages over total cumulative crater counts alone. If a planet-wide resurfacing event ended everywhere on Mars at the same (real) time, the surface corresponding to the end of the event would have the same crater retention age everywhere (using the Neukum and Hiller approach) independent of the previous history. Dating (of crater retention age) of common stratigraphic horizons should be possible.

Limitations include dependence of the results on the choice of standard production curve, applicability only where resurfacing has been a relatively discrete and efficient process, and dating only the termination of the event. No information is available on the beginning or duration of the resurfacing. Despite these, the approach has important potential for identifying individual resurfacing events and for dating the surfaces produced, even when those surfaces are buried by later events. In addition, craters dating from a given surface can be located and used to infer the minimum extent of now-buried surfaces (7). It is also possible to obtain on the thickness of layers representing the resurfacing events from the smallest crater surviving from each event.

## COMMON-AGE RESURFACING EVENTS

A number of common ages occur when the Neukum and Hiller technique is used at widely different locations around Mars. For heavily cratered terrain south of the highland-lowland transition zone in eastern Mars, in Xanthe Terra adjacent to Lunae Planum, in Tempe Terra and Arabia Terra, an old branch defined by craters 40-100 km diameter with a crater retention age  $N(1)$  (cumulative number greater than 1 km diameter per  $10^6$  km<sup>2</sup>) about 250,000 is common. This Early Cratered Terrain age is also found in the knobby terrain of Elysium, but does not occur in the highland-lowland transition zone (3) nor in the ridged plains of Lunae Planum (4) or Tempe Terra (5). This old age may be a resurfacing of still older crust: the Arabia Npl<sub>1</sub> cratered unit curve has an well-determined branch at  $N(1) \approx 800,000$ . This same ancient age occurs in Xanthe Terra. Craters which define this age are from 70-150 km across.

A major resurfacing event in cratered terrain in eastern Mars and in Tempe Terra occurred at  $N(1) \approx 85,000$ . We associate this with the formation of

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cratered plateau material and Tanaka's (1) Late Noachian intercrater plains resurfacing (3). Relic surfaces of this age do occur in the highland-lowland transition zone in eastern Mars and in the Elysium-Amazonis knobby terrain (3,6). Within the transition zone this age surface is better preserved east of the Isidis Basin, probably due to a thinner overlying cover from subsequent resurfacing events. In Xanthe Terra the first major resurfacing event has a crater retention age  $N(1) \approx 50,000$ , considerably lower than the 85,000 described above. If this is the same cratered plateau material resurfacing event seen elsewhere, it ended later in Xanthe Terra. Perhaps proximity to the Tharsis-Valles Marineris-outflow channel complex played a role.

The most widespread common event appears to be of Lunae Planum Age,  $N(1) 25,000$ . This event occurred in cratered terrain, the transition zone, the knobby terrain in Elysium-Amazonis, and in ridged plains in Tempe, Lunae Planum and elsewhere. A spread in the ages appears real, from  $N(1) < 32,000$  in cratered terrain Mars to less than 18,000 in ridged plains south of the Valles Marineris. There appears to be a progression in the Lunae Planum ages in this region, younger westward toward Tharsis.

#### EXTENT AND THICKNESS OF BURIED UNITS

The individual branches which define the different resurfacing events in the cumulative frequency curves provide information on the extent of buried surfaces (7). From Figure 1, the early cratered terrain surface at  $N(1) 250,000$  in Tempe Terra is defined by craters in the 40-100 km, and the cratered plateau age surface at  $N(1) \approx 89,000$  is marked by 20-35 km wide craters. By locating these diameter craters it is possible to map out the minimum extent of these cratered surfaces. In Figure 2 note the absence of any craters dating from these two surfaces within the ridged plains in Tempe Terra. The Lunae Planum Age surface is found throughout the region.

The smallest diameter crater which defines a given branch provides control on the total overlying thickness of materials due to all subsequent resurfacing events. From Figure 1, the early cratered terrain surface can not lie any deeper than 435 m or the rim heights (8) of craters 40 km across would be buried. Likewise for the cratered plateau age surface, because 20 km craters survive all subsequent resurfacings, the total overlying thickness of these materials is less than 315 m. The difference between these yields a thickness of material associated with the cratered plateau age resurfacing of 120 m in the HCR unit. This is only a statistical average based on assumed uneroded rim heights of craters surviving from a given age surface. It is also possible to estimate the depth to surfaces not preserved in the cumulative frequency curves. For example in Figure 1 there is no branch for the ridged plains unit RPL dating from  $N(1) \approx 85,000$ . But in the area occupied by those plains there could be, if that surface does lie below the plains, 10 craters larger than 48 km and one as large as 90 km not seen because of the thickness of the plains. Thus the total thickness probably lies between 475 and 645 m, or about 560 m. Figure 3 shows cross-sections for these units based on the above approach.

The thickness of the Lunae Planum Age resurfacing materials in cratered terrain is generally thin ( $< 100\text{m}$ ). In the transition zone values range from 150-235 m, while in ridged plains  $> 250$  m. For the cratered plateau age resurfacing the materials appear to be of more uniform thickness, 100 to 150 m except in plains-forming units which exceed 150 m. The thickness of materials post-dating the Lunae Planum Age resurfacing at is less than 200 m in cratered terrain,  $> 250$  m in ridged plains and transition zone units, and  $> 450$  m in the lowlying northern plains.

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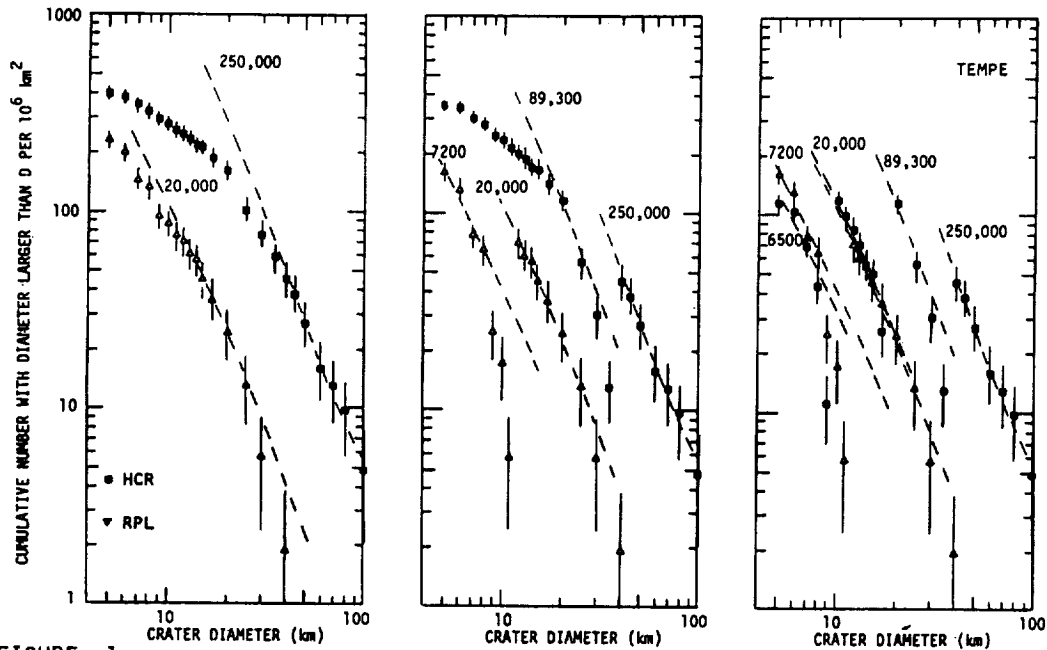
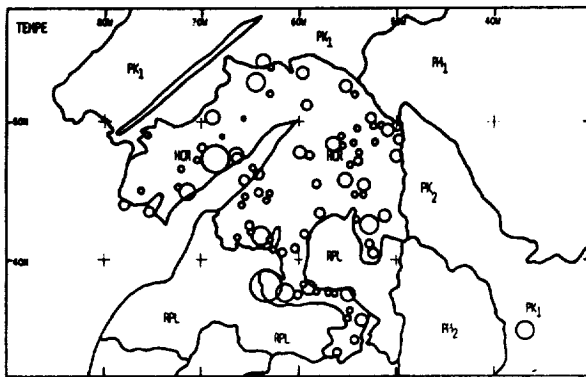


FIGURE 1

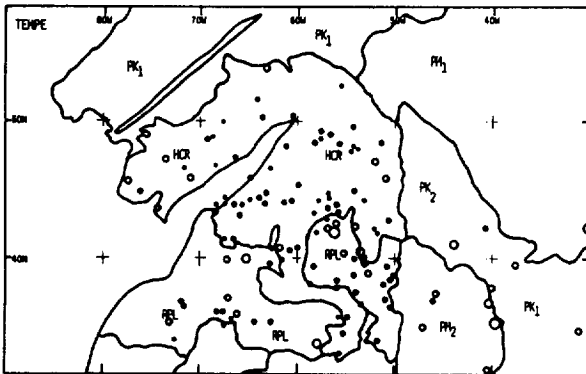
Application of the Neukum and Hiller technique to cumulative frequency curves for heavily cratered terrain (HCR) and ridged plains (RPL) in the Tempe Terra region. Note especially the common age surface at  $N(1) = 20,000$  (Lunae Planum Age) and the absence of older surfaces in the ridged plains plots.

FIGURE 2A



Distribution of craters predating the Lunae Planum Age resurfacing event in Tempe Terra region.

FIGURE 2B



Distribution of craters defining the Lunae Planum Age resurfacing event in Tempe Terra region.

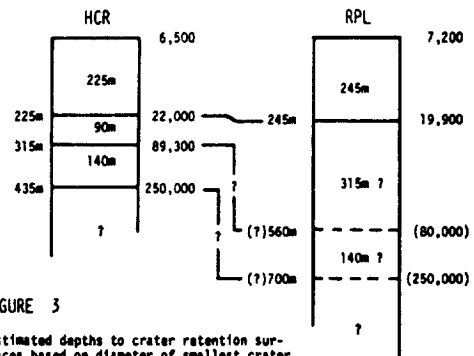


FIGURE 3

Estimated depths to crater retention surfaces based on diameter of smallest crater which defines that age surface. HCR and RPL represent heavily cratered and ridged plains unit in the Tempe Terra region. Note the greater thickness of overlying units in the ridged plains. Depth to early cratered terrain surface could exceed 700 m.



**EARLY VOLCANISM ON MARS: AN OVERVIEW; Ronald Greeley,**  
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Geologic study of Mars has been underway for more than two decades. Photogeologic mapping, aided by analyses of other remote sensing data and considerations of geophysical constraints, has revealed a remarkably diverse planet with a complex history. Global synthesis of these studies in the form of three maps published at 1:15 million scale provides a uniform framework for assessing many aspects of martian geologic history (1,2,3).

Volcanism has played a key role in the evolution of the martian crust. It is estimated that half the planet is surfaced with volcanic materials (4), most of which appear to be of mafic compositions. Moreover, with volcanic features included in all ages of surfaces, volcanism spans the entire "visible" history of Mars. However, volumetrically, more than half of the identified volcanic materials were erupted and emplaced in the first billion years, with a "peak" in activity 3.5-4.0 billion years ago ("absolute" ages are dependent upon models of meteoritic flux). Following planetary accretion and the development of a lithosphere, loss of volatiles via impact erosion and hydrodynamic escape may have led to a thin or nearly absent atmosphere at the start of the "visible" history of the planet ~4-4.4 by ago. Explosive volcanism in this environment would have been enhanced, perhaps leading to the generation of extensive ash deposits. Although definitive evidence of such an origin is lacking because of deep erosion and modification, much of the early upland plateau may have been mantled with explosive volcanic products. Large quantities of volcanic dust injected into the thin, evolving martian atmosphere would have blocked solar heating of the surface, lowering the temperature and perhaps resulting in a "volcanic winter". Low temperatures would have driven volatile materials, including water, into the subsurface as ground water and ground ice.

At least some intercrater plains consist of thin sheets of lavas that may represent ultra mafic lava flows, perhaps similar to komatiitic lavas erupted during Precambrian times on Earth. Such flows are considered to have erupted at very high temperatures and to have been extremely fluid. Estimates of flow rates and viscosities suggest that the flows would have been turbulent and capable of eroding and assimilating older rocks. Such flows on Mars could have released volatiles to the atmosphere, not only from the deep interior but also through erosion into ground water and ice zones in the regolith. Other eruptions involved flood-style lavas and their interbedding with deposits of fluvial, aeolian, and impact origin. Some of these lavas contain sets of ridges similar in morphology to lunar "wrinkle ridges" and may be analogous to mare flood lava flows.

Highland paterae, such as Tyrrhena Patera, represent the earliest recognizable "central" volcanic constructs. Most of the highland patera occur in the southern hemisphere. Four, Tyrrhena, Peneus, Amphitrites, and Hadriaca paterae are found adjacent to the Hellas basin. Peterson (5) proposed that circumferential ring fractures resulting from the basin impact could have served as eruption conduits. As suggested by Greeley and Spudis (4), magma rising through water-saturated mega-regolith would have led to phreatomagmatic eruptions (6 and this volume) to produce early-phase ash eruptions and the broad shield morphology of the paterae. With time, the water supply from the groundwater system diminished or was sealed, leading to effusive activity to produce the younger lava flows observed in the summit of Tyrrhena Patera.

The martian Eastern Volcanic Assemblage (2) includes the central volcanoes and lava plains of the Elysium province and of Syrtis Major. The earliest phases of eruptions occurred during the Hesperian Period and are approximately contemporaneous with the formation of the highland patera. Syrtis Major is a low-profile central volcano which produced a series of flows extending radially from the vents as far as 900 km. The caldera vents for Syrtis Major are located on ring fractures of the Isidis impact basin. In many respects, the Elysium volcanic province is an older, smaller version of the Tharsis volcanic region. Earliest volcanism occurred in the Middle

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Hesperian Period to produce Hecates Tholus, followed by eruptions which formed Albor Tholus. Mouginiis-Mark et al. (7) suggested that a smooth region on Hecates Tholus is a mantling airfall deposit from a plinian eruption. Late-stage (Amazonian age) eruptions led to the formation of Elysium Mons and the series of flows which partly buried the flanks of Hecates and Albor mons.

In summary, most of the recognizable volcanic units on Mars were emplaced in the first 1 to 1.5 aeons and involved flood eruptions; lavas appear to have had rheological properties similar to lunar mare flows and to komatiitic lavas on Earth, and are inferred to be mafic to ultramafic in composition. Central-vent volcanism occurred later, and in some cases was associated with impact-generated crustal deformation. Some central-vent volcanism appears to have provided pyroclastic deposits.

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## CHEMICAL AND PHYSICAL PROPERTIES OF PRIMARY MARTIAN MAGMAS; John R. Holloway and Constance M. Bertka\*, Departments of Chemistry and Geology, Arizona State University, Tempe AZ 85287. \*Also at the Geophysical Laboratory, Carnegie Institute of Washington.

**Introduction.** Primary magmas are in equilibrium with a mantle source region phase assemblage. They constitute the starting point composition for the magmas ultimately reaching the Martian surface or intruded into the crust. The purpose of this paper is to review current understanding of the chemical and physical properties of primary magmas in Mars and to speculate on some of the consequences of those properties for the early volcanic evolution of Mars.

**Mineralogy and Chemistry of the Martian Mantle.** The most recent estimates of the bulk chemistry of the mantle (1,2) have been tested experimentally for consistency between phase equilibria and estimated density (3). The good agreement found suggests that the proposed bulk compositions are reasonable. The two compositions are nearly identical and are similar to estimates of the Earth's mantle except for significantly higher ratios of Fe/Mg and Fe+Mg/Si. These compositions yield the mantle mineral assemblages shown in Figure 1. The position of the spinel-lherzolite to garnet-lherzolite transition is well known (4), while the other transitions shown have not been determined for the Martian mantle composition and are estimated based on studies of terrestrial compositions (5,6).

**The Volatile-absent Solidus for Mars.** Also shown in Figure 1 is the solidus temperature as a function of pressure. The only two points experimentally established are at one bar and 23 kbar. The rest of the solidus is based on analogy with terrestrial studies (7).

**Primary Magma Composition.** The chemical composition of the primary magma formed by small (< 2 wt%) degrees of melting in the garnet-lherzolite field is given in table 1. The normative mineralogy is very rich in olivine. The most striking

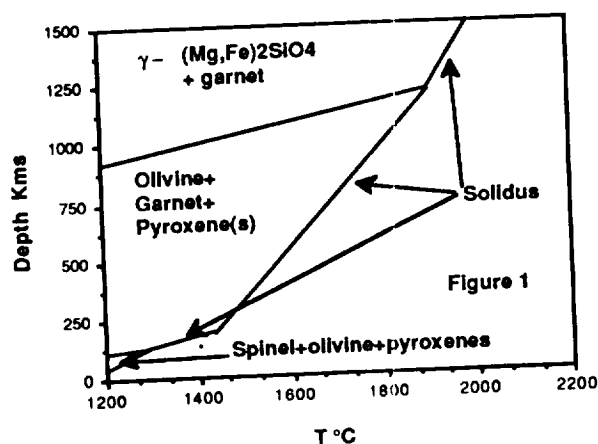


TABLE 1

Oxide	Wt%	Normative Mineral	Wt%
SiO <sub>2</sub>	41.9	ab	7
Al <sub>2</sub> O <sub>3</sub>	11.9	an	22
FeO	23.3	di	15
MgO	12.7	hy	1
CaO	9.4	ol	48
Na <sub>2</sub> O	0.8		
mg#	49.3		

features are the very high FeO/MgO ratio and the low Al<sub>2</sub>O<sub>3</sub> content. Melting at pressures greater than 23 kbar in the garnet-lherzolite field will result in even higher normative olivine contents. Larger degrees of partial melting will result in loss of garnet in the residual phase assemblage, a decrease in normative olivine, and further decrease in Al<sub>2</sub>O<sub>3</sub>. Melting in the spinel-lherzolite field will result in lower normative olivine (hence higher silica) and higher Al<sub>2</sub>O<sub>3</sub> contents. The actual Martian mantle probably has higher C and H abundances than the Earth and so the composition of melts formed at low degrees of partial melting will be strongly influenced by CO<sub>2</sub> and H<sub>2</sub>O, whereas melts at high degrees of partial melting will not be significantly affected by volatiles. Regardless of the myriad possibilities for variations in the composition of primary magmas on Mars, they will all have very high FeO/MgO ratios compared to those on Earth, and this feature will be retained in most derivative magmas.

#### Primary Magma Physical Properties.

The low silica and alumina and the very high iron content of these primary melts results in unusual physical properties. Viscosity is very low with calculated values in the 2-10 poise range at atmospheric pressure and preliminary experimental results suggesting viscosities of 1-2 poise. A striking feature of the melts is their high density, caused primarily by their high FeO content. Calculated densities (based on 8) are shown as a function of FeO/MgO ratio in Figure 2 and as a function of pressure in Figure 3. The calculations predict that the primary magmas will sink relative to pyroxenes and olivine at some pressure between 25 and 35 kbar. Addition of either H<sub>2</sub>O or CO<sub>2</sub> to the melt will lower its density (9,10) and thus the melt will only sink at greater pressure, possibly 40-50 kbar. The point at which the melt sinks relative to pyroxenes and olivine occurs at much lower pressures in the Martian mantle than in the Earth's (compare 11). The pressure at which primary magmas sink relative to olivines and pyroxenes puts an effective lid on source region depth which

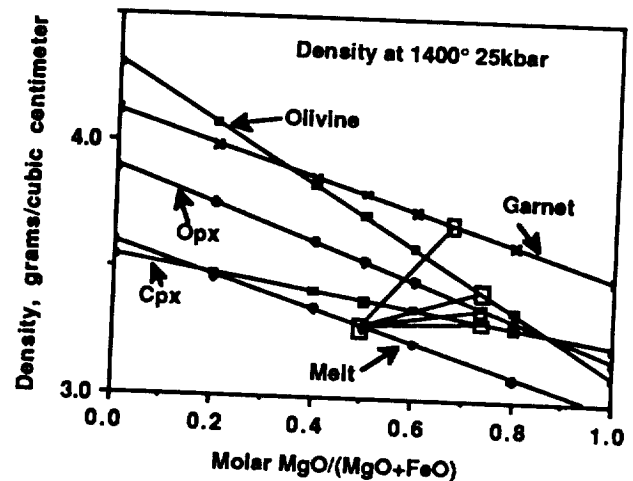


Figure 2. The tie-lines connected to boxes

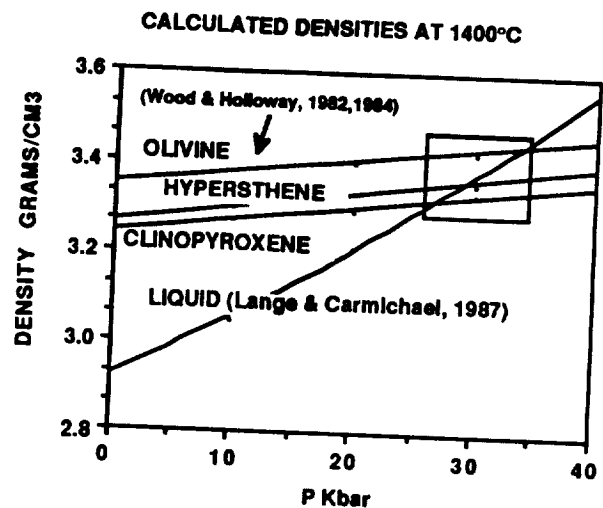


Figure 3. The box includes the pressure range in which melt begins to sink.

PRIMARY MAGMAS  
Holloway, J.R. & Bertka, C. M.

is estimated to be between 190 to 375 kilometers.

**Fractionation of Primary Martian Magmas.** Any primary magma originating in the Martian mantle is likely to crystallize olivine as it ascends to lower pressures (12). The effect of such fractionation is shown in Figure 4. Olivine fractionation of the volatile-absent primary magma from the garnet-lherzolite Martian mantle results in  $\text{SiO}_2$ , FeO and MgO levels similar to those in SNC meteorite compositions, but the  $\text{Al}_2\text{O}_3$  levels deviate significantly. This suggests that, if the SNCs are Martian then they are products of either large degrees of partial melting, or of two-stage melting of an  $\text{Al}_2\text{O}_3$ -depleted mantle as previously argued (13).

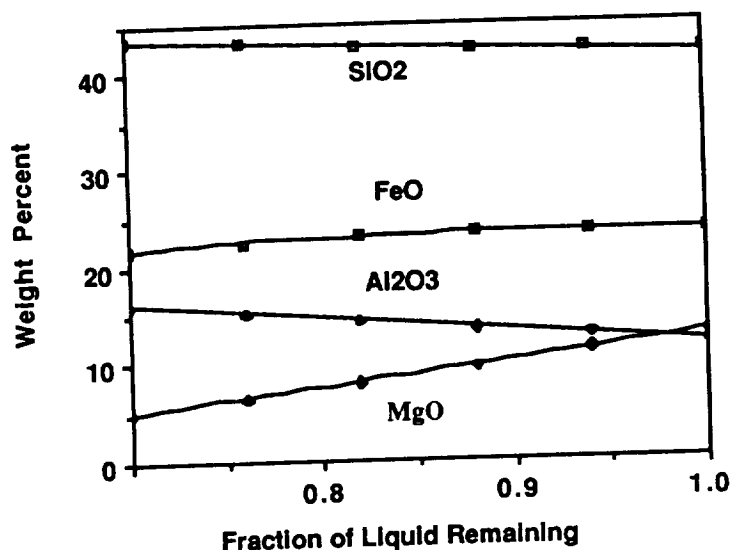


Figure 4.

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**VOLATILE INVENTORY, OUTGASSING HISTORY, AND THE EVOLUTION OF WATER ON MARS.** Bruce M. Jakosky and Ted A. Scambos, Department of Geological Sciences and Laboratory for Atmospheric and Space Physics, University of Colorado, Boulder, CO 80309.

The outgassing history of Mars and the subsequent evolution of water at the surface are intimately connected to the initial volatile inventory, as present at the surface or within the interior of the planet at the end of planetary formation. Recent observations and models have significantly changed our viewpoint on the volatile history of Mars and its connection to the history of volcanism; the relevant analyses are reviewed here, with an eye toward reconciling the different models.

Working backwards, let us start with the history of water at the surface. Carr calculates that about 50 m (global equivalent layer) of water is required to carve the catastrophic flood channels and explain the other surface features indicative of the presence of water. Greeley suggests that the observed volcanism will also account for about 50 m of water. Measurements by Owen et al. of the atmospheric D/H suggest a significant loss of water to space over geologic time; depending on the size of the non-atmospheric reservoir with which atmospheric water can exchange, Yung et al. and Jakosky calculate this loss to be between 3 and 60 m of water. The latter value is most consistent with other information about the Mars climate. Although water has clearly been important in the formation of surface geologic features, there is no evidence to contradict the simple model that all of the water released to the surface has escaped to space except for that which is present today in the two polar caps, chemically and physically bound within the near-surface regolith, and in the atmosphere (about 15 m of water).

Outgassing of water (and other volatiles) to the surface occurs during planetary formation, as a result of the violent impacts of accreting planetesimals, and during subsequent planetary evolution, associated with volcanism and other processes which can release interior water to the surface. The recent Earth-accretion model of Zahnle et al., if extended to the martian case, suggests that Mars probably did not end accretion with an atmosphere containing a substantial portion of its volatiles; this is based on the fraction of accreting material expected to be outgassed as well as on the ability of a small accreting planet to retain an atmosphere against escape to space. Measurements of atmospheric  $^{36}\text{Ar}$  support this hypothesis, suggesting the retention of a small fraction of the planetary volatiles. Surface extrusive volcanism occurring over geologic time will release water to the atmosphere, as will shallow igneous intrusions associated with volcanism. The history of surface volcanism described by

VOLATILE INVENTORY  
Jakosky B. M. and Scambos T. A.

Greeley shows planetary activity increasing to a maximum intensity about one billion years after Mars' formation; this result is completely consistent with thermal history calculations. If planetary outgassing is proportional to volcanic activity, then this curve also shows the relative intensity of outgassing as a function of time. From this history, along with the measured atmospheric  $^{40}\text{Ar}$  abundance and various estimates of the initial bulk planetary  $^{40}\text{K}$  abundance, we can calculate the release factors of volatiles. For the probable range of  $^{40}\text{K}$  abundances, between about 0.01 and 0.1 of the  $^{40}\text{Ar}$  has been released to the surface, and between 0.02 and 0.2 of the water has been released. For Greeley's 50 m of water released by volcanism, this suggests that at least 250 m, and possibly as much as or more than 2.5 km, is still retained within the planet.

Although formulated in a different manner, this value is consistent with the conclusion by Carr that as much as 500 m of water remains locked up beneath the surface. Geochemical evidence for the water abundance of the planet comes from comparison of volatile abundances within the atmosphere or within the SNC meteorites. Unfortunately for the former, conditions within the solar nebula are sufficiently uncertain that the amounts of volatiles incorporated into planetesimals which eventually accumulated to form Mars are difficult to estimate. Comparing trace element abundances in the SNCs to terrestrial, lunar, and meteoritic abundances, Dreibus and Wanke suggested that Mars is relatively volatile rich. They estimate only about 130 m of water incorporated within the whole planet, however; this amount must be considered extremely uncertain given the assumptions regarding the composition and volatile abundances of the accreting planetesimals and the conditions extant on the forming Mars as compared to those expected based on thermal history calculations.

In summary, the available observations and models are consistent with the presence of several hundred meters or more of water within Mars, with most of this still being in the interior. Of that which has been outgassed, much of it has escaped to space, and most of the rest may be locked up in the polar caps. The relationship between the water abundance and the history of volcanism is clearly an important one, both in the sense of water within the magma affecting the volcanic style and of volcanism serving as both a conduit and a measure for outgassing of the interior.

THE ROLE OF MANTLE CONVECTION IN THE ORIGIN OF THE THARSIS AND ELYSIUM PROVINCES OF MARS, Walter S. Kiefer and Bradford H. Hager (Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena CA, 91125)

The Tharsis province of Mars is 6000 to 7000 km in diameter and contains four major shield volcanos as well as a number of smaller volcanic structures. Although the long-wavelength topography of Mars is not well known, Tharsis rises approximately 7 to 9 km above the plains to the north and west (1-3). The peak nonhydrostatic geoid anomaly is 1.5 km for spherical harmonic degrees 2 to 18 (4). The Elysium province appears to be a somewhat smaller scale version of Tharsis, with a peak topographic uplift of 3 to 5 km, a peak geoid anomaly of 300 m (degrees 3 to 18), and three main volcanic structures.

Existing models of Tharsis and Elysium have generally treated these regions in terms of either isostatic or flexural models (5-11). However, the large horizontal scale of these regions and the large volume of volcanic materials suggests that mantle processes, particularly convective flow, also must have played a major role in the evolution of these regions. Some existing models for Tharsis have treated thermal anomalies in the upper mantle in terms of Pratt compensation. We have shown (12, 13) that Pratt models and convective models predict very different relationships between geoid and topography, and thus Pratt models should not be regarded as an adequate substitute for convection modeling.

We propose that Tharsis and Elysium overlie regions of broad-scale upwelling caused by the internally heated component of convection in the mantle of Mars. Within these broad upwelling regions, a number of mantle plumes are formed by the bottom heated component of convection. These plumes fed the various shield volcanos. In some cases, a single plume may have fed several closely spaced volcanos. Uranus Patera, Uranus Tholus, and Ceraunus Tholus may be an example of such a cluster of volcanos with a single plume source. The proposed concentration of Martian mantle plumes within the Tharsis and Elysium regions is reminiscent of the situation on Earth, where hotspots show a pronounced bimodal distribution (14).

Cratering statistics indicate that eruptive activity ceased at different times at the various volcanos in Tharsis and Elysium (15). Eruptive activity at some of the smaller Tharsis volcanos ceased soon after the end of the heavy bombardment, while activity at the four largest Tharsis shields continued for a much longer time. The youngest flow units at Olympus Mons and Ascraeus Mons may be less than 100 million years old. In Elysium, volcanic activity may have ceased more than 1 billion years ago (15), although some recent studies indicate that at least some volcanism in the Elysium region may be comparable in age to the youngest volcanic activity at Olympus Mons (16-18). The extent of young volcanism in Elysium is in dispute, however, with Plescia (18) arguing for extensive young volcanism and Tanaka and Scott (17) favoring only a limited amount of recent volcanism. Models for the thermal evolution of Mars indicate that the heat flux dropped sharply during the first 1 billion years after accretion and has continued to decline at a slower rate since that time (19). We suggest that as the heat flux decreased, the number of plumes in the Martian mantle which were



sufficiently active to lead to volcanism may also have decreased, leading to the cessation of eruptive activity at many of the volcanos.

In the past, dynamic support for the topography of Tharsis and Elysium has sometimes been rejected on the grounds that it can not be maintained for a significant fraction of Martian history (7,9). However, Mars is sufficiently large that mantle heat transport must be currently dominated by convection. In the absence of plate tectonics, Tharsis and Elysium could stay over mantle upwellings for an indefinite length of time. The evidence for recent volcanism in Tharsis supports the idea that mantle upwelling and dynamic support of topography is currently important for Tharsis. Hall et al. (10) argued that an apparent lack of recent volcanism in the Elysium region implied that convective upwelling is not currently important in supporting topography there. However, as noted above, some recent volcanism apparently has occurred in the Elysium region. In any case, a lack of young volcanic extrusions should not be interpreted as ruling out currently active mantle upwelling. Convection-induced igneous activity could also occur in the form of intrusive bodies, which would not be detectable in Viking Orbiter imagery.

We have used finite element calculations to assess the possible contribution of mantle convection to the geoid and topography of Tharsis and Elysium. In our preliminary modeling, we have used a cylindrical axisymmetric geometry and a non-temperature-dependent rheology. The neglect of spherical geometry may not be important for models of Elysium but clearly is important for Tharsis and will need to be included in more definitive models. In scaling our models, we assume a mantle thickness of 1600 km and a mean mantle heat flow of  $30 \text{ mW/m}^2$ . We have filtered our results to include only wavelengths which are comparable to those included in the geoid model of Balmino et al.(4). Our results are sensitive to a number of unknown parameters, including the Rayleigh number, the variation of viscosity with depth, and the ratio internal heating to bottom heating. In models with a horizontal scale appropriate to Tharsis, we find that convection can cause up to 700 m of geoid and 10 km of topography. An identical model with a horizontal scale appropriate to Elysium produces 380 m of geoid and 7 km of topography. These results show that the current long-wavelength geoid and topography of Tharsis and Elysium may be dominantly the result of convective processes. They also suggest that the large differences in the observed amplitudes of geoid and topography for the Tharsis and Elysium swells may be largely due to differences in the horizontal sizes of their underlying convection cells. These results do not include the possible effects of membrane stresses in the elastic part of the Martian lithosphere. However, at the long wavelengths considered here, an elastic lithosphere with a mean thickness of less than 100 km should reduce the calculated topographic uplift by less than 30% (20). Geoid values would be reduced by a larger amount which depends on the variation of viscosity with mantle depth as well as the elastic layer thickness.

Another constraint which has sometimes been applied to models of Tharsis and Elysium is to compare predicted lithospheric stresses with observed tectonic features (6,7,8,10). Our models predict that at the surface, the azimuthal normal stress is more extensional than the radial normal stress. Thus, any extensional features which form should be oriented radially to the center of the uplift, in agreement with observations of extensional features in Tharsis. Because the equations which relate convective

Mantle Convection on Mars  
W.S. Kiefer and B.H. Hager

velocities and strain-rates, and hence velocities and stresses, take similar forms in spherical and cylindrical geometries, we believe that this result will also hold for convection models performed in spherical geometry. We have not yet analyzed how these stress field results may be modified by the inclusion of an elastic lithosphere.

Our results show that convective processes played an important role both in the formation of Tharsis and Elysium and in maintaining the current geoid and topography of these regions. Clearly, isostatic and flexural processes also play a role in maintaining the geoid and topography of Tharsis and Elysium. Ultimately, it will be necessary to combine both convective and lithospheric processes into a coupled, self-consistent model.

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GROUND ICE ALONG THE NORTHERN HIGHLAND SCARP, MARS; B.K. Lucchitta and M.G. Chapman, U.S. Geological Survey, Flagstaff, Arizona 86001.

The northern highland boundary on Mars is characterized in many places by a steep scarp that is locally dissected into mesas and buttes. These are particularly well developed in the region from about long  $280^{\circ}$  to  $350^{\circ}$  and lat  $30^{\circ}$  to  $50^{\circ}$  N, the "fretted terrain" of the Protonilus and Deuteronilus Mensae. In this latitude belt, mesas are typically surrounded by debris blankets that are generally thought to have formed by ice-lubricated flow [1]. This ice came either from frost deposition [1] or, as many researchers believe now, from ground ice [2,3,4]. The debris blankets are conspicuous features of this latitudinal belt and are young [2]. They are generally thought to indicate that ground ice may currently be present close to the surface [3]. Lucchitta [2] explained the occurrence of ice-charged debris blankets in this latitude belt by (1) the maintenance of ground ice near the surface, which, under current atmospheric conditions, is possible only poleward of about lat  $30^{\circ}$  to  $40^{\circ}$  N [5,6] and (2) the ambient temperature that is warm enough to permit ice to flow by dislocation creep, even though at very slow rates [2]. Farther north, where the atmosphere is colder, this type of ice deformation is less likely to occur [7].

We have observed a number of mesas in the Kasei Valles area that resemble fretted mesas in size and shape and that appear to have formed by similar processes. These mesas are located farther south (lat  $25^{\circ}$  N) than the Deuteronilus or Protonilus Mensae and are not presently surrounded by debris blankets. Instead, several of the mesas are surrounded by moats composed of flat-floored valleys (about 2 to 5 km wide) bordered on the outside by low, inward-facing scarps (Fig. 1, arrows). The moats have the same planimetric shape and dimensions as the debris blankets surrounding the fretted terrain farther north. The plains surrounding the moats are composed of probable Tharsis lava flows, whose relative crater density of 1300 craters  $>1 \text{ km}/10^6 \text{ km}^2$  indicates that they are about 2.5 b.y. old (conversion of relative to absolute age from Neukum and Hiller, model 2 [8]). The similarity of the shape of the moat to the planimetric shape of debris blankets surrounding the fretted mesas farther north suggests that the mesas in Kasei Valles once were also surrounded by debris blankets and that these blankets must have been in place at the time of lava flooding; the lavas apparently embayed the debris blankets. Later, the debris blankets disappeared completely, leaving the moats.

The disappearance of the debris blankets suggests that they contained more ice than is commonly attributed to them (10% ice content according to Lucchitta [3] and Carr [5]). In order to disappear completely, only minor amounts of rock could have been incorporated, unless all the rock was of a grain size easily picked up and removed by the wind. However, no other wind erosion features are seen in the area. Also, such a uniform grain size is unlikely unless very unusual circumstances are invoked, such as debris blankets reworking largely fine-grained ash or sandy deposits. Considering, then, that the debris blankets were mostly composed of ice and that this ice was derived from ground ice underlying the mesas [2,3], the ground ice must have had a concentration so high that it may have resembled segregated ground ice in the Northwest Territories of Canada [10]. This observation suggests that the highland margin and associated mesas in fretted-terrain regions locally were underlain by nearly pure ice masses. These ice masses could have been segregated from ice-rich highland-type breccia, or they

## ICE, HIGHLAND SCARP

Lucchitta, B.K. and Chapman, M.G.

could represent frozen water from ancient channel-mouth regions. Furthermore, if the debris blankets were composed largely of ice, sublimation of this ice on fresh exposures within the slowly churning debris would eventually deplete the ice, especially at the tip of the debris blankets where the ice is replenished only slowly. Sublimation of debris blankets in this manner would explain why most of the blankets on fretted mesas extend no farther than about 20 km from their source. Sublimation of debris blankets on gradually diminishing mesas would also explain the disappearance of material from the plains separating the mesas. Thus, the generation of the plains may have been caused by the fretting process [2] involving large-scale sublimation of ice and does not have to be ascribed to an unrelated earlier erosional episode [1].

The former existence of debris blankets in the Kasei Valles area also suggests that ground ice once occurred near the surface in a latitudinal belt that is now desiccated. This observation supports Fanale et al.'s [6] proposition that the southern limit of near-surface ground ice shifted gradually northward with time. Ground ice at lat 25° N 2.5 b.y. ago falls right on the equatorward limit permitted in Fanale et al.'s models of near-surface ice retention.

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## ICE, HIGHLAND SCARP

Lucchitta, B.K. and Chapman, M.G.



Figure 1

STRUCTURAL MODIFICATION ALONG THE CRATERED TERRAIN BOUNDARY, EASTERN HEMISPHERE, MARS. Ted A. Maxwell, Center for Earth and Planetary Studies, National Air and Space Museum, Smithsonian Institution, Washington, D. C. 20560

Initial investigations of the Mars cratered terrain boundary (CTB) suggested that it formed progressively by erosional backwasting of the cratered terrain (1). Such conclusions were based on the presence of isolated knobs and mesas in the boundary zone adjacent to the southern cratered terrain, particularly in the Nilosyrtis Mensae region, and by the fact that the boundary is present now as an elevated scarp in the eastern hemisphere. Evidence for mass wastage of both the main bounding scarp and along the edges of isolated plateaus also supported such interpretations, further indicating that most of the erosion took place during a period(s) of wetter climate. Based on more recent investigations of the structure, topography and timing of the CTB, however, it is now apparent that the sequence of events that are responsible for the present appearance of the CTB, at least in the region between Nilosyrtis Mensae and Aeolis (Long. 300° to 180°), occurred within a discrete time interval, indicating that the northern border scarp of the cratered plateau material is not the result of longterm migration of a previous structurally controlled topographic disruption. Based on structural mapping, detailed shadow measurements of plateaus and knobs, and age dating of the exposed surfaces in this region, the tectonic record of the eastern hemisphere is one of modification of the cratered terrain boundary, with little evidence that can be used to constrain theories for the original creation of the CTB.

Structural mapping of the CTB indicates little correlation between presumed compressional features (ridges and scarps) and the orientation of the boundary itself (2). Instead, compressional features are aligned in a primary NS direction, with local deflections where underlying topography reaches the near-surface in plains units (as judged by structural trends in nearby highland areas). The notable exception is in a wide trough that extends to the southwest between Isidis and the impact crater Herschel. Here, scarp orientations are predominantly axially symmetric, suggesting continued deformation by downdropping of this major trough past the times of plains emplacement (3). The 300 km wide block of cratered plateau material on the north side of the trough acted as a discrete structural unit during deformation, and may be analogous to the subsurface terrace that lies beneath the shallowly flooded region of plains to the north of the CTB.

Shadow-length elevation measurements of individual plateaus, knobs, and graben in and north of the cratered terrain boundary suggest that downdropping of the "terrace" also occurred as a discrete crustal block rather than as individual faulted fragments of crust. If formed by erosional retreat, isolated remnants of the cratered plateau material

STRUCTURAL MODIFICATION OF MARS DICHOTOMY BOUNDARY  
Maxwell, T. A.

would display a greater degree of erosion with distance from the present boundary. Such is not the case, due to either the presence of a significant resistant layer, or downdropping as a terrace. In the absence of confirmation of a widespread resistant unit from the limited Viking multispectral data (4), the latter interpretation is preferred here.

The timing of downdropping suggests structural modification as a discrete event, though for reasons listed above, not the event that created the dichotomy boundary itself. The faulting that produced the broad, 300 km wide terrace occurred in late Noachian or early Hesperian time (around N(5)=190), younger than the faults surrounding the Isidis basin (5). In other regions, however, faulting of plateau material occurred at different times, suggesting a protracted period of structural adjustment to the original boundary.

Current theories for creation of the Mars crustal dichotomy are 1) single mega-impact (7), 2) multiple large impacts (8), and 3) crustal overturn (9). Hypotheses 1 and 2 would be expected to show either concentric rims or radial fractures, neither of which occur in the highlands between Isidis and Aeolis. The large crustal block north of the graben-like trough east of Isidis suggests endogenic processes (regional extension) rather than impact for at least the modification of the presently observed boundary, which is consistent with hypothesis 3, though not conclusive. By looking at modification processes and prior geometry of the boundary region, more precise constraints can be placed on the geometric form of the original structure.

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POTENTIAL INDICATORS OF PYROCLASTIC ACTIVITY NEAR ELYSIUM MONS, MARS. Kathleen McBride, Univ. of Houston-Clear Lake and Lunar and Planetary Institute, James R. Zimbelman, National Air and Space Museum, Smithsonian Institution, Washington, D.C., 20560, and Stephen M. Clifford, Lunar and Planetary Institute, 3303 Nasa Rd. 1, Houston, Tx., 77058.

The Elysium region contains abundant evidence of volcanism, most of which is interpreted to predate the Tharsis Montes eruptions [1]. Also, the shape and morphology of Elysium Mons suggests that the magmas at Elysium may be more chemically evolved than those in the Tharsis region [2].

Preliminary photogeologic mapping of Elysium Mons (22.5-27.5°N, 210-220°W) on a 1:500,000 scale has revealed numerous features indicating the possible occurrence of both pyroclastic and effusive volcanic activity. Analysis of Viking Infrared Thermal Mapper data reveals a strong correlation between low thermal inertia values and a linear fracture located at 23°N, 216.5°W, indicating a possible volcanic vent. Image resolution for this area is poor (250 m/pixel) and no individual lava flows are visible in the vicinity. Craters are irregularly shaped and appear subdued as if mantled by dust or a volcanic airfall deposit.

Close inspection of high resolution images (~44 m/pixel) of the Elysium Mons caldera has disclosed five levels of collapse. A crater measuring five km in diameter south of the caldera rim has been interpreted to be of impact origin [2]. However, the crater is not completely circular and the eastern rim shows evidence of slumping or mass wasting. The crater also contains a large central pit which is uncharacteristic of an impact crater of this size. On the northern side of the caldera is a tectonic depression that breaches the caldera rim. This "impact" crater and the depression are connected by a linear chain of craters across the caldera floor indicating possible tectonic control. There are also other numerous pits and crater chains radiating outward from the caldera. These are believed to be endogenic in origin, formed by the collapse of lava tubes or collapse along rift zones.

The eastern flank of Elysium Mons appears relatively smooth and only slightly hummocky with no distinct lava flows visible at a resolution of ~44 m/pixel. Some impact craters in this area appear mantled by some surficial deposit. The northwestern flank of the volcano consists of massive undifferentiated lava flows and ridges that grade into the lava flood plains. The only mappable lava flows are a few smooth, flat lobate flows. Examination of high resolution (~35 m/pixel) images of the western side of Elysium Mons has disclosed numerous small lava channels, some of which breach the rims of small impact craters.



POTENTIAL INDICATORS OF PYROCLASTIC ACTIVITY  
McBride K. et al.

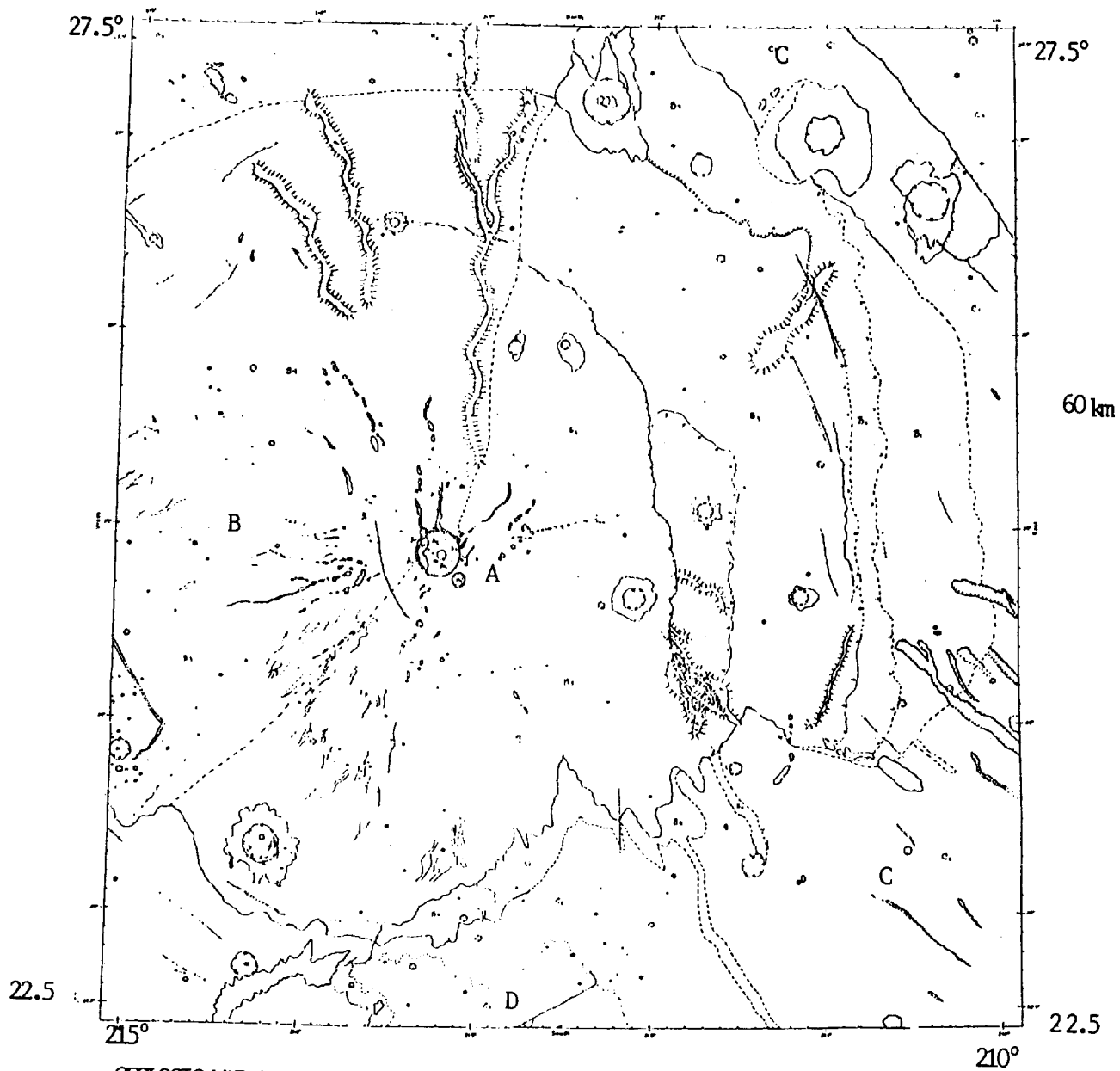
Four aligned domes resembling terrestrial cinder cones have been identified at 27°N, 218-219°W. These domes possess circular summit depressions and are aligned roughly parallel to the concentric fractures circumferential to Elysium Mons, indicating possible structural control. However, there are many other dome-shaped features in this area that are randomly oriented and lack any evidence of summit craters at a resolution of ~150 m/pixel. Small, randomly distributed dome-like structures are also present in the area around Elysium Fossae (23.5-25°N, 219-220°W). These features vary in size from 1-5 km in diameter. Most lack summit depressions and a few appear as elongated ridges.

There have been several investigations into possible volcano-ground ice interactions in Elysium Planitia. Mouginiis-Mark *et al.* [1,3] found a variety of morphological features indicative of an interaction of ground ice with erupting magma. The apparent cinder cones may be related to the pseudocraters Frey *et al.* identified in Cydonia [4]. These pseudocraters are dome-like structures Frey *et al.* believe result from phreatomagmatic eruptions.

The area centered at 26°N, 220°W is believed to be a fourth volcanic center, in addition to Elysium Mons, and Albor and Hecates Tholii. It has been called the "Elysium Fossae Complex" by Mouginiis-Mark *et al.* and is comprised of a series of vents, lava flows, and domes [5]. They believe the morphology of the domes in this area is similar to terrestrial rhyolite domes.

Further analysis is currently in progress and will include interpretation of geologic units and structural features to determine the geologic history and eruption characteristics of Elysium Mons.







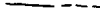
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GEOLOGIC MAP OF ELYSIUM MONS

Geologic Units: (A) caldera (B) undifferentiated lava flows (C) flood lavas  
(D) lava plains

Map Symbols:

	graben		break in slope
	depression		lava channel
	crater		lava flow
	contact, dashed where inferred		

**THE MARTIAN CRUSTAL DICHOTOMY;** George E. McGill, Department of Geology and Geography, University of Massachusetts, Amherst, MA 01003.

The physiographic and stratigraphic contrasts between the northern lowland plains and the southern cratered uplands of Mars are generally interpreted to imply a first-order dichotomy in the martian crust. Although this dichotomy is as fundamental as the continental/oceanic crustal dichotomy on Earth, no evidence has yet been found to suggest a similar origin. Hypotheses for the origin of the martian crustal dichotomy are of two general types: 1) "catastrophic" convective overturn of the mantle, probably associated with core formation (1); or 2) impact, either a single giant impact (2), or focussing of several large impacts (3). All of these hypotheses are envisaged by their perpetrators as involving unusual events that occurred very early in martian history. Each implies physiographic, stratigraphic and structural consequences that may be compared with observation. LOS gravity from Viking 2 (4) provides additional constraint.

By analogy with the moon, one might expect mascons to be associated with large impact basins; mascons are clearly present over the Isidis and Utopia basins (Fig. 1), but not over the proposed giant Borealis basin (Fig. 2). It has been suggested informally (Squyres, Pers. Com.) that the absence of a mascon is due to the uniqueness of the giant Borealis impact, but this argument needs to be developed formally. Substituting several large impacts for the single giant one is not a tenable alternative. There is no reasonable mechanical explanation for the required focussing of large impacts on 1/3 of the surface of Mars. Even if an explanation could be devised, several closely spaced large impact basins must produce a topography that includes areas higher than the original surface as well as areas lower; there is no way they could form a single large hole, all of which is lower than the original surface.

If the giant Borealis basin really exists, then it must be older than any surviving features within its rim, including both the Isidis and Utopia basins. This requires that it be older than the oldest defined time-stratigraphic units (7). The presence of inliers of old upland terrain over much of the northern plains (e.g., 5) supports

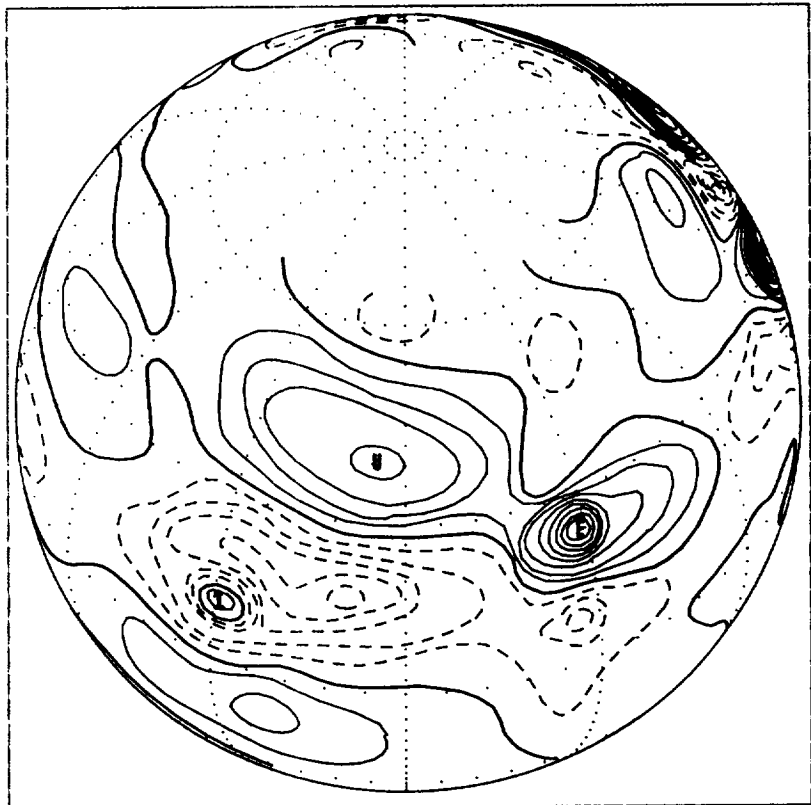


Fig. 1. LOS gravity (4). Projection centered on Utopia basin (5) at 48N, 240W. Contour interval=10 mgal; dashed lines negative. U and I are mascons associated with Utopia and Isidis basins; anomaly E lies over Elysium Mons.

this conclusion by indicating that ancient crust lies beneath most or all of the northern plains. Furthermore, this ancient crust cannot have suffered significant thinning by removal of surface materials since early to middle Noachian. The oldest craters now buried beneath northern plains deposits have suffered extensive fracturing and erosion before burial, surviving as rings of knobs defining their rims. The maximum depth to which surface materials could have been removed since early to middle Noachian thus can be estimated by applying the crater dimensional equations of (8) to the smallest surviving "knob ghost" craters ( $D=10-15$  km). This exercise indicates a maximum of ~200 meters of lowlands-wide erosion of the ancient

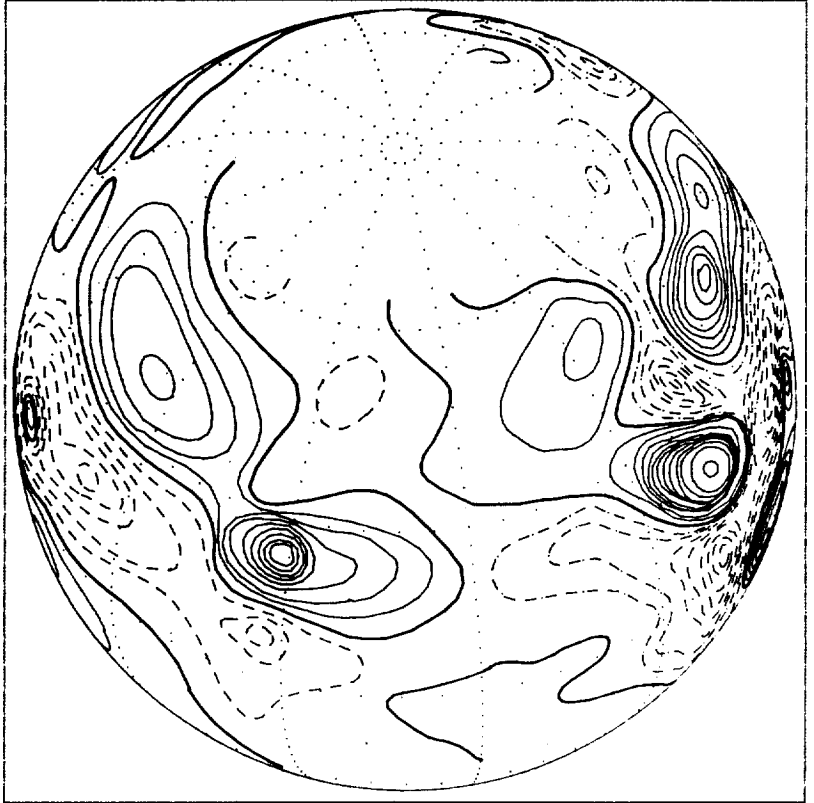


Fig. 2. LOS gravity centered on putative Borealis basin (2) at 50N, 190W (no mascon). Large anomalies at right are Olympus Mons and Tharsis.

surface buried beneath the northern plains. It also invalidates suggestions that the present dichotomy boundary is significantly south of its original position because of erosional scarp retreat (9) requiring the removal of the top 2-3 km of the ancient crust. In the eastern hemisphere, small "knob ghosts" are present on the lowland side of the boundary right up to the scarp, indicating that no more than ~200 meters can have been removed, and thus that most if not all of the relief along the present dichotomy boundary in the eastern hemisphere is structural.

The impact hypothesis for the origin of the dichotomy boundary thus requires a single mega-event that occurred very early in martian history and left no mascon. The present dichotomy boundary scarp is probably very close to its pre-Noachian position, based on counts of "knob ghost" craters buried beneath the plains just north of the boundary. Hence local lack of correspondence between the dichotomy boundary and the proposed Borealis basin rim cannot be explained away as due to later scarp migration by erosion; either the dichotomy boundary is not the rim of a Borealis basin, or the basin is significantly "out of round". Despite arguments by (3), the very large size of the putative basin is not a problem, but its apparent lack of roundness and the implication of stability along the dichotomy boundary since pre-Noachian time are problems.

A number of recent studies suggest that the dichotomy boundary is not a stable ancient feature. There is clear evidence of relatively young faulting and major erosion in northern Mars, some of it demonstrably in early Hesperian (10), the rest less precisely dated as late Noachian through Hesperian (11,12). In fact, it is probable that the widespread fretted terrain and mesas were in

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large part formed at this time. It has been proposed (13) that most of the small drainage networks developed at this time in response to pervasive volcanism and sill intrusion. Crater counts indicate that there was widespread resurfacing by lava flows at about the Noachian/Hesperian boundary (14). Finally, detailed analysis of Viking LOS gravity (15) indicates that there is a narrow negative anomaly associated with the dichotomy boundary, suggesting activity more recent than the presumed early origin of the dichotomy required by the mega-impact hypothesis.

Thus it appears as if the northern part of Mars experienced several possibly related events in late Noachian and/or Hesperian time: fracturing and faulting of the crust in the northern third of Mars, widespread erosion with formation of channels and valley networks, and extensive extrusive and intrusive igneous activity, most of it occurring near or on the dichotomy boundary. An internal process that thinned the crust from below is the most likely cause, but the specific mechanism envisaged by (1) is ruled out because if a "catastrophic" core-formation event occurred, it must have taken place very early in martian history. Nevertheless, the cause must be internal because the geological and geophysical observations discussed above require that a mega-impact occur much earlier in martian history.

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**THARSIS AND THE EARLY EVOLUTION OF MARS;** Roger J. Phillips, Dept. of Geological Sciences, SMU, Dallas, TX 75275; and Norman H. Sleep, Dept. of Geophysics, Stanford, CA 94305

The Tharsis and Elysium provinces are the result of focused heat and mass transport in the Martian lithosphere that commenced early in the history of Mars. The oldest structural evidence unequivocally tied to the Tharsis structure is a set of circumferential fractures and graben associated with Claritas Fossae [1]. Some of the graben are seemingly interrupted by large craters, suggesting that they are approximately contemporaneous with the underlying stratigraphic units, rather than being significantly younger. These underlying units are Noachian in age, implying that the first Tharsis events were relatively early in Martian history.

Central to discussions on the origin of Tharsis are the relative roles of structural uplift and volcanic construction in the creation of immense topographic relief. The circumferential fractures of Claritas Fossae can be interpreted in terms of flexural uplift, but this does not imply, necessarily, that much of the elevation of Tharsis was created by this mechanism. It is also possible that uplift was minor and followed a major episode of volcanic construction [2].

Conditions leading to anomalous heat transport in the Martian lithosphere have been classified as active [3,4] or passive [2,5]. In the former case, a mantle plume provides the heat source for the creation of Tharsis, although the existence of such a plume does not necessarily favor uplift mechanisms over constructional mechanisms. In the passive case, the pre-Tharsis lithosphere starts out anomalously thin (perhaps due to a large impact), which concentrates stresses; this leads to sustained transport of magma into and through the lithosphere. There need not be anything remarkable about the mantle beneath Tharsis in this instance.

We can condense the broad scale geophysical issues concerning Tharsis to the following questions: How was the great elevation of Tharsis achieved? How much of the elevation is due to "permanent" (as opposed to transient thermal) uplift? The second question is complicated semantically because constructional processes (i.e., processes that add volume to the crust by the addition of igneous material) can be intrusive and lead to uplift. This has been suggested as a mechanism for plateau uplift on the Earth [6]. We call this process "intrusive uplift". Another type of uplift is "compensation uplift", wherein lateral mass loss from a region leads to upward isostatic adjustment. This process is familiar on the Earth as erosion followed by isostatic rebound [7].

Recently, a new class of Tharsis models have been formulated [1,8] that are "petrologically mass conservative". This means that the mass of all elements of a partial melting process (residuum, melt products, lateral mass loss) is equivalent to the mass of the original source region that undergoes partial melting. *Phillips and Sleep* [1] additionally emphasize the importance of the isostatic constraint in such a system. Perhaps the most surprising result of their analyses is that the so-called "isostatic effect", the consequences of the difference between isostatic balance and mass balance, can profoundly effect the achievable elevation in a magmatic system when the depth of compensation exceeds a few percent of the planetary radius. The process can be self-defeating: Mass transferred toward the surface contributes to increased positive elevation with increased partial melting because of the decreased density of solidified melt products. At the same time, due mostly to the effects of membrane stresses, mass transferred to the surface becomes a load increasingly in excess of isostasy, and this contributes to downward isostatic adjustment. Often, it is the density of the residuum that makes the largest net contribution to positive elevation.

To achieve the required elevation of Tharsis seems to require a depth of compensation of about 150 km or less and a depth extent of partial melting less than twice this, or about 250 km (also taken as the thickness of a Pratt lithosphere). This value is in agreement with an estimate, based on a simple hydrostatic argument, from the height of the youngest shield volcano, Olympus Mons [9]. Thus the 250 km depth might mark the last vestiges of magma generation beneath Tharsis, which we expect to have deepened with time. This result is in conflict with gravity models that are purely isostatic [e.g., 10,11] and require lithospheric thicknesses of 400 km or more. However, a small flexural load added to an otherwise isostatic configuration

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will satisfy *both* the total elevation and gravity anomaly constraints. The formation of the radial fractures and graben on the periphery of Tharsis also seems to require a flexural load [12,13].

As mentioned, we argue that the fracture and graben patterns seen at Claritas Fossae demand that flexural uplift played a role in achieving the total elevation of Tharsis, and that very early in the history of Tharsis the lithosphere failed in this mode. Subsequently, the elevated region of Tharsis was in radial deviatoric tension and tended to fail isostatically. Lateral mass loss (compensation uplift) may be capable of providing some of the uplift, but does not seem to be a mechanism capable of achieving a significant fraction of the total elevation of Tharsis. Elevation gain by intrusion and thickening of the crust (intrusive uplift) is a part of the constructional process, and we see no reason why some significant fraction of Tharsis magmas could not end up in the crust (and upper mantle), as opposed to on the surface. This process has been invoked for plateau uplift on the Earth [6], and it may be even more likely on Mars, given that relatively low magnesium numbers are likely for its mantle. Further, successful gravity models require a densification of the lower crust.

In summary, the model we propose for Tharsis has the following characteristics: (i) Partial melting in the mantle started at shallow depths and never proceeded below a depth of about 250 km; (ii) The mechanical properties of the lithosphere were largely isostatic; (iii) A significant fraction of the magma ended up as intrusive into the crust (and upper mantle) and helped create uplift of the surface; and (iv) Due to the isostatic effect, the late magmatic products formed a flexural load. This load is in part responsible for the stresses that formed the peripheral or outer radial fractures, but thermal contraction may have also contributed.

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**TECTONIC IMPLICATIONS OF MARTIAN WRINKLE RIDGES**  
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Wrinkle ridges are common physiographic features on the terrestrial planets that have been interpreted as forming in compressional environments. Recent work (Plescia and Golombek, 1986; Watters, 1988) has begun to address the structural style and specific mechanisms of formation of wrinkle ridges. We have previously interpreted wrinkle ridges as structures forming under compressional stresses that are composed of low-angle reverse (thrust) faults forming an anticlinal fold at the surface.

Topographic data for wrinkle ridges allow an estimate of the magnitude of shortening due to folding and due to faulting. Shortening due to folding is obtained by unfolding the ridge (the difference between the integrated length along the ridge surface and the horizontal distance across the ridge). Shortening due to faulting is proportional to the dip of the fault and the elevation offset across the structure, which characterizes virtually all wrinkle ridges. This estimate is model-dependent because the dip of the fault is unknown. However, compressional faulting of prefractured rocks is expected to occur along thrusts that dip about 25 degrees, along which the frictional resistance to sliding is minimized (Brace and Kohlstedt, 1980; Byerlee, 1978). Using these estimates, the total shortening across a wrinkle ridge can be calculated as well as the strain.

High-resolution topographic data derived from stereo photogrammetry are not available for Mars making determination of the vertical relief of Martian wrinkle ridges difficult. However, monoscopic photoclinometric techniques (Davis and Soderblom, 1984) can provide topographic information (profiles). About 250 topographic profiles have been collected for wrinkle ridges in Lunae Planum and Amazonis Planitia using the photoclinometric method. These data allow an estimate of the shortening and strain associated with wrinkle ridges.

Ridges in Lunae Planum and Amazonis exhibit about 100 to 300 m of total relief. These ridges also show an elevation offset across the ridge. That is, the plains on one side of the ridge have an elevation distinctly different from the plains on the opposite side. The offsets are about 100 m and are consistent along the ridge. In northern Lunae Planum, plains on the east side of the ridge are typically higher than plains to the west.

The regional elevation offset is a critical element in the understanding of the structural origin of wrinkle ridges. Neither folding (e.g., Muehlberger, 1974) nor shearing (e.g.,



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Tija, 1976) can readily produce the regionally consistent elevation offset associated with the ridge. Vertical faulting (e.g., Lucchitta, 1976; Sharpton and Head, 1988) does not provide a mechanism for producing the broad low positive relief structure characteristic of wrinkle ridges. A combination of folding and low-angle faulting, however, can produce both the observed ridge morphology and the regional elevation offset.

Assuming the model discussed above, the amount of shortening and strain due to faulting and folding can be calculated. The shortening due to folding is about 5 to 25 m; that due to faulting is about 100 to 300 m. These are similar to values for lunar wrinkle ridges. The ratio of shortening to due faulting versus that due to folding, indicative of the relative contributions of faulting and folding, is between 5 and 20. These data indicate that faulting is the dominant mechanism for accommodating shortening in Martian wrinkle ridges; folding is subsidiary.

Strain across wrinkle ridges can be calculated at two scales; strain across an individual ridge and strain on a regional scale across a basin or province. For Lunae Planum, individual ridges exhibit a local folding strain of about 0.05 to 0.5 %, a faulting strain of 1.0 to 5.0 % and a total combined strain of 1.4 to 5%. The average regional strain across Lunae Planum (at about 20 degrees N) is estimated to be 0.2 to 0.5%. By comparison, on the Moon, the local strain is about 0.36% for folding and about 1.1% for faulting and the combined strain is about 0.26 to 3.6% (Golombek et al., 1988). An estimate of the strain across the Humor basin on the moon is about 0.22%.

In summary, wrinkle ridges are interpreted to be tectonic in origin--a low-angle thrust fault and an associated overlying anticlinal fold. Martian wrinkle ridges characteristically exhibit a regional elevation offset (the plains on one side are higher than on the other). Shortening amounts to 5-25 m due to folding and 100-300 m due to faulting. Regional strain across northern Lunae Planum is estimated to be 0.2 to 0.5%.

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SOLID STATE CONVECTION AND THE EARLY EVOLUTION OF MARS.  
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Recognition of the fundamental role of solid state convection in the mantles of the terrestrial planets, other than the Earth, has been much delayed because of the absence of great horizontal displacements of their lithospheres similar to plate motions. The latter has, of course, been generally viewed as the primary evidence for convection in the Earth's mantle. However, it can be argued that this is not the most fundamental evidence for convection in a planetary mantle: rather the existence of nonhydrostatic low harmonics in their gravitational fields is the most clear evidence for conductive motions. Only the high harmonics can plausibly arise from the density variations in the lithospheres retained by finite strength.

A correlation appears to exist between high topography and positive gravitational anomalies but attempts to explain this by isostatic model have not been successful and some kind of dynamical support has been favoured which again points to the importance of solid state creep.

From asymmetries detected in many craters and from supposed polar sedimentary deposits palaeoequators have been traced which are different from the present. Thus evidence exists for ancient reorientations of Mars with respect to its axis of rotation: the necessary change in the moment of inertia tensor must be ascribed to changes in mass distribution probably associated with the volcanic processes producing the tharsis uplift. But again such "polar wandering" depends fundamentally on the hydrostatic equatorial bulge keeping in step with the migrating pole, i.e. at  $90^\circ$  to it. This is only possible if the flow occurs in the interior. Thus solid state creep in the mantle is its fundamental mechanical property over long times.

The discrepancy between the dynamical ellipticity and that which is calculated from various studies of its shape ( $1/120$  compared with about  $1/170$ ) appears to require the existence of a second degree convection pattern at present. It is not unreasonable to suppose that a single cell convection pattern might have been important in the earliest processes of differentiation in Mars and does seem to be the only satisfactory way of explaining the hemispherical asymmetry and the displacement between the centre of figure and the centre of mass: an analogous phenomena to that observed in the Moon.

NORTHERN PLAINS OF MARS: A SEDIMENTARY MODEL; R.  
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Several considerations suggest that a sedimentary model should be applied in the interpretation of the surface geology of the northern plains of Mars. In this model the upper few hundred meters of the plains are made up of weathered fluvial sediment, reworked locally by eolian processes. Volcanism plays a minimal role. The material underlying the sedimentary cover is similar to that of the ancient cratered terrains.

The northern plains of Mars are generally thought to have a diverse makeup and a complex geologic history. Some earlier observations based on Viking images led to divergent interpretations. Carr et al. (1) argued that a deep permafrost layer controlled much of the plains morphologic expression, as evidenced by giant polygons in Utopia. Guest et al. (2) interpret the fractured plains as a pediment surface, where the overlying mesa material has been stripped away. Various investigators have suggested cooling lava to explain the polygons. This interpretation was shown by Pechmann (3) to be unlikely on the basis of material properties. Frey and Jarosewich (4) analyze the numerous subkilometer conical features that occur on the northern plains, often on the polygonal plains. After discussing various possible origins, including pingos, they conclude that these features are volcanic, either cinder cones or pseudocraters involving lava and ice interaction. This interpretation of the activity should also be present. More recently, Lucchitta et al. (5) have favored a thick sediment deposit to explain the giant polygons. Lucchitta et al. (5) point out that the occurrences of giant polygons are in low areas that would have collected sediment from the outflow channels. They also map sinuous ridges along the margins of the polygonally fractured terrain and compare them to Antarctic ice ridges formed as pressure or flow ridges. DeHon (6) maps other linear features on the Acidalia plains and interprets them in a general framework of sediment-ice phenomena. Rossbacher (7) mapped curvilinear patterned ground associated with the boundary between uplands and northern plains and interprets them to be produced by differential erosion in the presence of a substantial amount of ground ice.

The depositional mechanism for this material presents a problem. McGill (8) argues that the sediments were deposited in sediment-charged floods from NW Elysium, somewhat downplaying his alternate suggestion of pyroclastics. Jons (9) also argues a sediment slurry origin of the plains, even to the extent of a mud ocean covering the northern lowlands. Lucchitta et al. (5) suggest a large standing body of water as an alternate hypothesis, but find the polygon deposits do not correspond to the deepest areas except locally and that the topographic map would require a 4-km deep ocean in one region. They allow that later tectonic warping may have occurred. However, in comparing the mapped occurrences of polygonal features of Lucchitta et al.

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with the topographic map of Christensen (10), in fact, the lowest regions of plains, lying below the -2-km contour, have the giant polygons.

Consideration of the history and nature of the surface material of the northern plains raises a number of questions. There are some fundamental issues, such as whether the plains surface and geologic features are volcanic or sedimentary, erosional or constructional. Also there are important secondary questions. What was the fate of the sediment that was removed from the channels of the Chryse Trough, including its initial and final state? How much total water was required to move that sediment? Why are fans and deltas not generally observed? Was the climate wet and warm before, during, or after the channel-forming events?

The climate issue has been addressed by Pollack et al. (11) and more recently by Owen et al. (12). These authors provide arguments for an early temperate period in Martian history when liquid water could have been stable. During this time, the silicate regolith of Mars would have been thoroughly weathered. Extensive carbonate deposits may also have formed. Carr (13) develops a reasonable scenario for catastrophic flooding involving sudden release of a confined aquifer from beneath a saturated permafrost table. In this case, the climate at the time of the floods may have become cold, and if there were a northern sea it would have been ice covered.

Carr (14) has estimated the volume of material removed in the formation of the valleys of the Chryse region as  $5 \times 10^6 \text{ km}^3$ . The sink area for this sediment, represented by the northern plains, is approximately  $5 \times 10^7 \text{ km}^2$ . Allowing for some change in porosity, this sediment would cover the northern plains with an average thickness of about 100 m from Chryse drainage alone, neglecting channels in other areas. Low regions such as Utopia could have accumulated much thicker deposits than the average. It is possible that the upper few km of the regolith were thoroughly weathered prior to the channel forming events. The regolith may have been already decomposed to friable clay-size grains and easily suspended and transported in the floods that swept down the channels. That the large channels are not cut into the northern plains suggests a base level for channel cutting. On Earth, this base level is provided by sea level, where the change in current velocity allows suspended sediment to be deposited in deltas and along continental margins. On Mars the channels may never have been cut into the plains or they may have been cut and then covered by later materials, volcanic or sedimentary.

Were there temporary lakes, or an ocean, on the northern plains of Mars? In spite of the dilemma posed by Lucchitta (5) based on incomplete topographic data for the northern plains, that an Acidalia ocean would be some 4-km deep, the topographic data basis of support for this concern is not reliable. Hypsometric analysis (15) of the global topographic map by Christensen (10), based on Earth-based radar topography, spacecraft radio occultation, IR, UVS, and optical measurements,

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provides some estimates of the volume of the northern basin so that the mass balance of water, sediment and basin size can be approximated. There are more recent maps, but no new data for the northern region. The mean elevation of the region below the base of the highland margin lies between the 0.5-km and 1.0-km contour. The mean elevation of the topography below this level is between 0 km and -1.0 km. The volume of this basin is  $4-5 \times 10^7 \text{ km}^3$  with great uncertainty, but probably not a factor of two. Carr (14) states that at least  $5 \times 10^6 \text{ km}^3$  are required to remove the volume of material represented by the Chryse valleys and allows that "the actual figure may have been substantially higher." If the actual ratio of water to sediment were 10, the northern basin could have formed an ocean.

One difficulty may be that this amount of water is high in comparison to any previous estimates of the Martian water budget. Greeley's (16) estimate of  $6.6 \times 10^6 \text{ km}^3$  of water released by volcanics is somewhat higher than Carr's minimum required for valley erosion. However, Greeley used only volcanic material presently observed. Based on our knowledge of the early volcanic history of the Earth and Moon, it is unlikely that volcanic rates were lower on Mars in its earliest history than later. Greeley's observations show a peak in Early Hesperian (~3.8 BY before present). It seems likely that earlier rates were much higher but the geologic record has been obscured. Thus much more water could have been available.

The model for the northern plains that arises from the above considerations would have the following features: (1) The epeirogenic basin aspect of the northern lowland predates the period of major channel formation. The fretted terrain is a consequent feature to the northern basin, formed by later erosion. (2) The northern plains are floored by sediment of variable thickness with provenance mainly in the upland Chryse and NE Elysium valleys. (3) The sediment and substrate of the northern plains originally contained considerable quantities of water. (4) The entire northern basin was occupied by an ocean. (5) The present surface features are predominated by the nival conditions that evolved with the changing climate and the life cycle of the northern sea.

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FAULT AND RIDGE SYSTEMS: HISTORICAL DEVELOPMENT IN WESTERN REGION OF MARS; David H. Scott and James Dohm, U.S. Geological Survey, 2255 N. Gemini Dr., Flagstaff, AZ 86001.

The tectonic history of Mars can be traced, in part, by mapping faults and ridges that were formed during successive geologic periods. This has been accomplished in the western equatorial region by determining the relative ages of these structures based on their occurrence in rock units of known stratigraphic position. Three maps (Figs. 1-3) at reduced scale (1:50,000,000) show the areal extent of geologic units that were emplaced, and faults and ridges that originated during the Noachian, Hesperian, and Amazonian Periods. We extracted and compiled these data from the geologic map of the western equatorial region of Mars (Scott and Tanaka, 1986), considering that (1) faults and ridges of Amazonian age may extend across the boundaries of older rock units; (2) structures of Hesperian age may extend into Noachian but not Amazonian rocks; and (3) Noachian structures only occur within the Noachian units. We also recognized, however, that older rocks may contain younger faults and ridges whose ages cannot be determined unless they also transect younger rocks. To make these discriminations where the geologic base (1:15,000,000 scale) did not clearly reveal the age relations, Viking photomosaics (1:2,000,000 scale) were examined. In Figures 1-3, the rocks exposed within each Martian period are enclosed within hachured lines. Faults and ridges in each period are shown by lines marked by ball and diamond respectively. The locations of volcanic centers are shown for reference only, and do not necessarily reflect their time of origin.

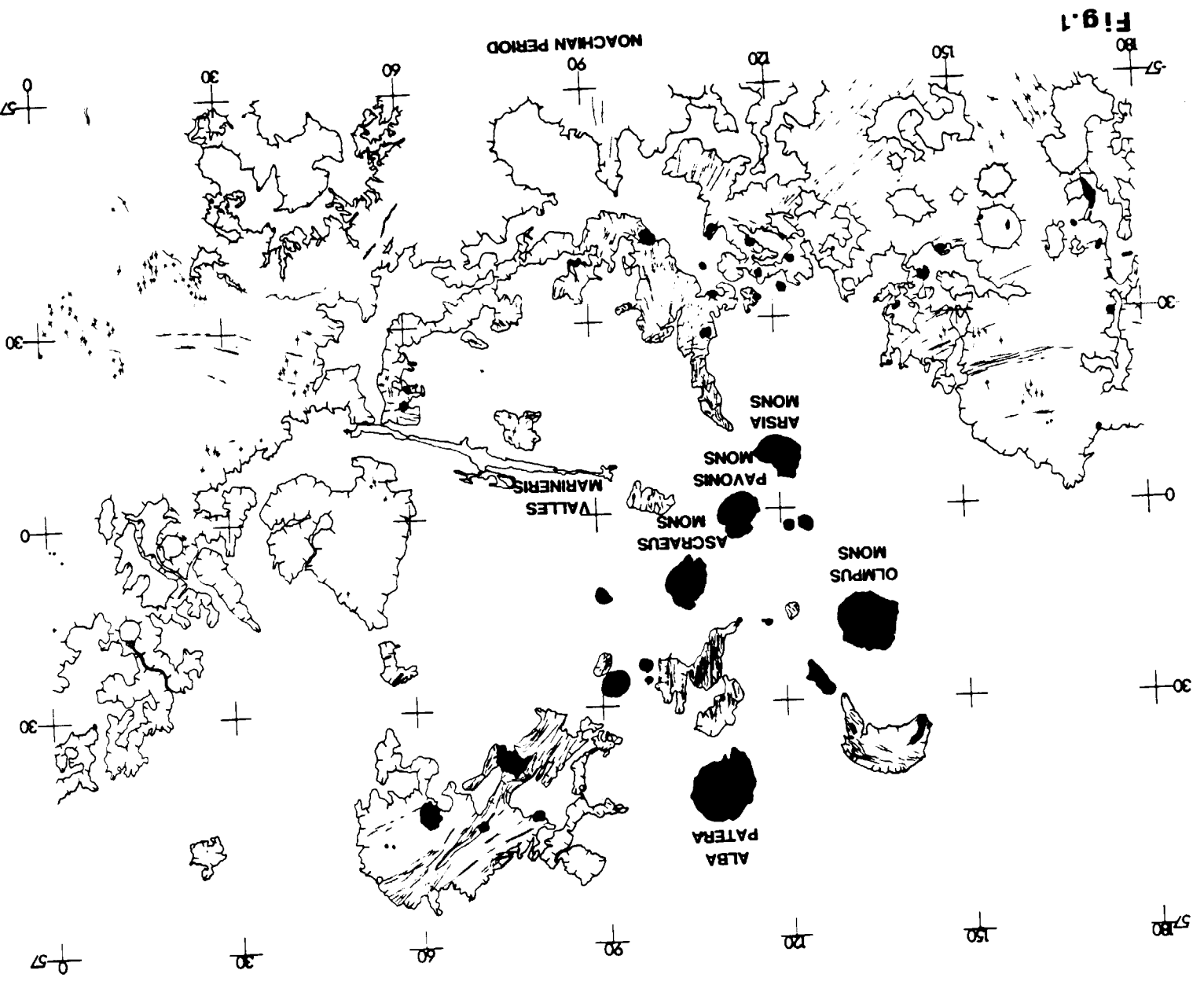
Figure 1 and previous studies by Tanaka (1987) indicate that most faulting during the Noachian Period was associated with three centers: 1) the Tharsis Montes axial trend and its northeastern and southwestern extensions; 2) the Syria Planum rise ( $-15^{\circ}$  lat,  $105^{\circ}$  long); and 3) the Acheron Fossae structure north of Olympus Mons. Clusters of ridges also occur and are confined within Noachian rocks but may have originated during the Hesperian Period.

Figure 2 shows that extensive faulting continued during the Hesperian along the Tharsis Montes trend and in Syria Planum, and was initiated radial to Alba Patera. Fault activity ceased, however, at Acheron Fossae. This period marked the culmination in western Mars of formation of ridges where they occur in a broad oval pattern around the major volcanic centers; most ridges are older than the faults of this period, possibly suggesting that those that occur in Noachian rocks were formed during an earlier period.

Figure 3 shows the great general decline in tectonism during the Amazonian Period. Only Alba Patera remained as a major fault center where older radial systems were rejuvenated and concentric faulting was initiated around the crest of the volcano. Minor faults are associated with Olympus Mons and its aureole materials; some are very young as they cut Late Amazonian lava flows.

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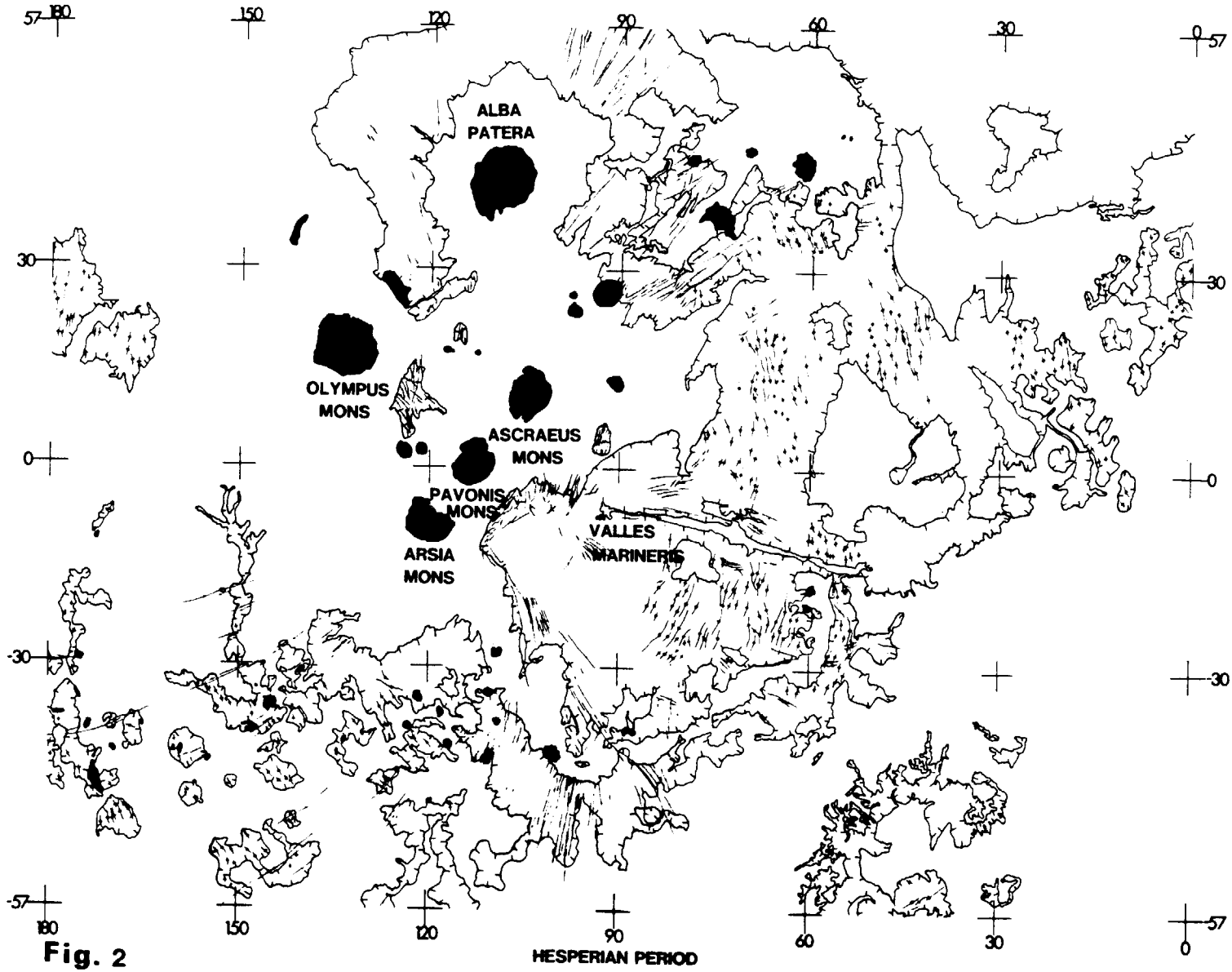


Fig. 2



Mars; Faults and Ridges  
Scott, D.H., and Dohm, J.

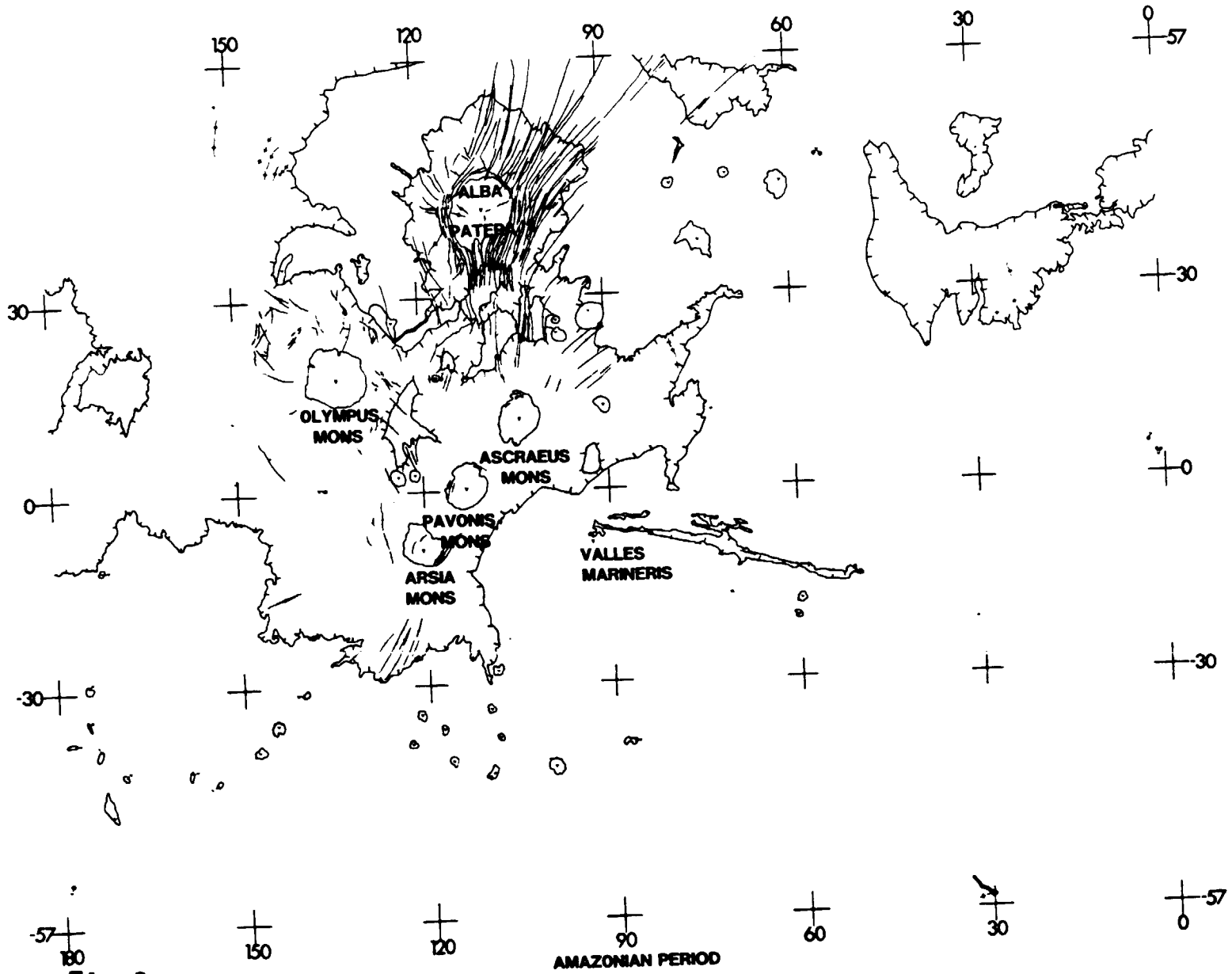


Fig.3

LATE IRON LOSS FROM THE MARTIAN LITHOSPHERE: NOT A LIKELY CAUSE OF VERTICAL TECTONICS, N. H. Sleep, Depts. Geology and Geophysics, Stanford University, Stanford CA 94305, and K. J. Zahnle, NASA Ames Research Center, Moffett Field, CA 94035

A density contrast of  $200 \text{ kg m}^{-3}$  throughout a 400-km thick lithosphere is needed for Tharsis to be isostatically compensated [Sleep and Phillips, 1985]. A conceivable mechanism is late loss of metallic iron originally accreted in a cool lithosphere or added subsequently by projectiles. Direct constraints on Mars are obtained from lead isotopes in shergottites which indicate iron removal very early in the history of the solar system [Chen and Wasserburg, 1986]. Conversely, the lead isotopic data indicates that abundant metallic iron did not remain in or become added to the Martian lithosphere at late times (at least for the part sampled).

Other constraints are obtained from estimating the total mass of projectiles hitting Mars subsequent to the main accretion. Comets (or outer solar system planetesimals) were sufficiently rare that the ferrous iron in the Martian mantle was not extensively oxidized. As this is hard to quantify, concentration is given to asteroids. Attention needs to be given to the time period covered by each estimate. The one direct estimate is obtained from the abundance of platinum group elements in shergottitic meteorites [e.g., Treiman et al., 1986, 1987]. The total mass of projectiles hitting Mars and becoming mixed in the mantle after the core became a closed system is equivalent to a layer about a kilometer thick. Unfortunately, the end of this accretion is poorly constrained as there is too much indigenous Pb in the martian mantle for a small addition of primitive lead to have much effect. The total Ir in the earth's mantle is equivalent to adding a layer 18 km thick [Chou et al., 1983]. Again the beginning of Ir retention is poorly constrained and may predate retention on Mars.

The best indirect constraints are obtained from studying the lunar regolith. The most useful element for determining the amount of meteorites is probably Ni, but the interpretation is not agreed on to a factor of 2 [e.g., McKay et al., 1986; Korotev, 1987; cf., Ringwood and Seifert, 1986]. That is, there is 2-5% meteorites in average highland upper crust [e.g., Drake, 1987], and the thickness of the mixed layer is around 30 km [e.g., Spudis, 1984]. The upper limit is also constrained by Mg/Al ratios as meteoritic components become evident around 8%, as in lunar sample 63507,15 of McKay et al. [1986]. These ranges agree with the 1 km equivalent thickness obtained from Mars. The mass of projectiles hitting the moon (mass of Vesta) is so great that reliable total masses can be obtained from elemental ratios; appeal to a special class of projectiles is unrealistic. The beginning of the sampling interval is probably around 4.4 B.Y. when the upper crust became a nearly closed system and much of the flux was over by 4.24 B.Y., the age of old granulites formed from breccia [Lindstrom and Lindstrom, 1986]. Two more model-dependent constraints in agreement with a kilometer layer are rarity of impacts on the moon that excavated into the deep crust and Pb isotopes in the lunar regolith.

The amount of energy in a 1-km thick layer of projectiles is enough to melt much of the material in the lunar or Martian megaregolith sometime during its history and also to stir the megaregolith. Conversely, the total energy from impacts is also not enough to significantly heat the deeper parts of the Martian lithosphere except locally. The total mass of projectiles is also too little to greatly perturb the density of the crust. Late loss of iron from the lithosphere is thus not a likely cause of uplift to form either the north-south dichotomy or Tharsis. Internal processes are therefore favored for the formation of Tharsis and other Martian tectonics after 4.4 B.Y.

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**HETEROGENEITIES IN THE THICKNESS OF THE ELASTIC LITHOSPHERE OF MARS: CONSTRAINTS ON THERMAL GRADIENTS, CRUSTAL THICKNESS, AND INTERNAL DYNAMICS.** Sean C. Solomon, Dept. of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139, and James W. Head, Dept. of Geological Sciences, Brown University, Providence, RI 02912.

*Introduction.* The thickness of the elastic lithosphere on a planet is essentially a measure of the reciprocal of the vertical thermal gradient in the lithosphere, i.e., the depth to a temperature at which ductile behavior replaces brittle behavior at typical geological strain rates. Under flexure there is an elastic "core" of the lithosphere occupying the depth interval over which the bending stress is less than an envelope of "strength" versus depth defined by a frictional failure curve at shallow depths and a ductile flow law at greater depth [1,2]. At the shallowest depths, lithospheric bending leads to faulting to a depth that is dependent on the load, the flexural rigidity, and the failure law. The depth of the lower limit to "elastic" behavior is governed primarily by temperature and also by strain rate, composition, and load magnitude. Estimates of elastic lithosphere thickness derived from simple models of flexure have been quantitatively related to the average vertical thermal gradient of the lithosphere on the Earth [e.g., 3] and Moon [4], and similar concepts have been used to constrain the thickness of the elastic lithosphere on Venus [e.g., 5]. In this paper we apply these concepts to Mars.

*Elastic Lithosphere Thickness.* The thickness  $T_e$  of the elastic lithosphere of Mars has been estimated from the tectonic response to individual loads [6,7] and from the global response to the long-wavelength load of the Tharsis rise [8,9]. The spacing of graben circumferential to the major volcanoes Ascraeus Mons, Pavonis Mons, Arsia Mons, Alba Patera, and Elysium Mons indicate values for  $T_e$  of 20 - 50 km (equivalently, values of flexural rigidity  $D$  of  $10^{30}$  -  $10^{31}$  dyn cm) at the times of graben formation [7]. For the Isidis basin region the elastic lithosphere thickness exceeded 120 km ( $D > 10^{32}$  dyn cm) at the time of mascon loading and graben formation [7]. The absence of circumferential graben around Olympus Mons requires the elastic lithosphere to have been at least 150 km thick ( $D > 3 \times 10^{32}$  dyn cm) at the time of loading [6,7]. Models of the flexural response of Mars to the long-wavelength topography of the Tharsis rise provide a reasonable fit to the geoid and to the distribution of tectonic features in the Tharsis province if the global elastic lithosphere of Mars is 100 to 400 km thick, corresponding to  $D = 10^{32}$  to  $7 \times 10^{33}$  dyn cm [8,9].

The values for  $T_e$  derived for individual loads (Figure 1) are not consistent with a simple progressive increase with time in the thickness of the elastic lithosphere of Mars. The largest estimates of  $T_e$ , for instance, are for perhaps the oldest (Isidis mascon) and youngest (Olympus Mons) features considered. Spatial variations in elastic lithosphere thickness must have been at least as important as temporal variations [7]. In particular, there appears to have been a dichotomy in lithosphere thickness that spanned a significant interval of time, with comparatively thin elastic lithosphere ( $T_e = 20$  to 50 km) beneath the central regions of major volcanic provinces and substantially thicker elastic lithosphere ( $T_e$  in excess of 100 km) beneath regions more distant from volcanic province centers and appropriate for the planet as a whole.

*Implied Thermal Gradients.* The values of  $T_e$  may be converted to the mean lithospheric thermal gradient, given a representative strain rate and a flow law for ductile deformation of material in the lower lithosphere. Formally, this is done by converting  $T_e$  to  $T_m$ , the depth to the rheological boundary marking the base of the mechanical lithosphere. This conversion is accomplished by constructing models of

bending stress consistent with the strength envelope and finding for each model the equivalent elastic plate model having the same bending moment and curvature [3]. The choice of rheological boundary is to some extent arbitrary; we take it to be the depth at which the ductile strength falls below 50 MPa [3]. We take the representative strain rate for the flexural response to each local load to be the quotient of the maximum horizontal strain given by the elastic model and the growth time of the load, taken to be  $10^{8\pm 1}$  yr. The uncertainty in growth time contributes only a small uncertainty to the derived value of  $T_m$ .

A considerably larger uncertainty arises from the poorly known value for the thickness of the martian crust and the distinct flow laws for crustal and mantle material. The mean crustal thickness consistent with global topography and gravity must be at least 30 km [10], which corresponds to zero crustal thickness beneath the Hellas basin. Models of the Viking line-of-sight (LOS) residuals over the Hellas basin and the 370-km-diameter crater Antoniadi are consistent with complete local Airy compensation only if the crust is 120-130 km thick [11,12]. LOS data over Elysium Planitia and Olympus Mons can be fit with varying degrees of Airy isostatic compensation and crustal thicknesses of 30-150 km [13,14].

We assume that the large values of elastic lithosphere thickness determined from the local response to the Isidis mascon and Olympus Mons and from the global response to the Tharsis rise exceed the thickness of the martian crust. Because flexurally-induced curvature is modest for these loads, the depth  $T_m$  to the base of the mechanical lithosphere is approximately equal to  $T_e$  [3] and is determined by the

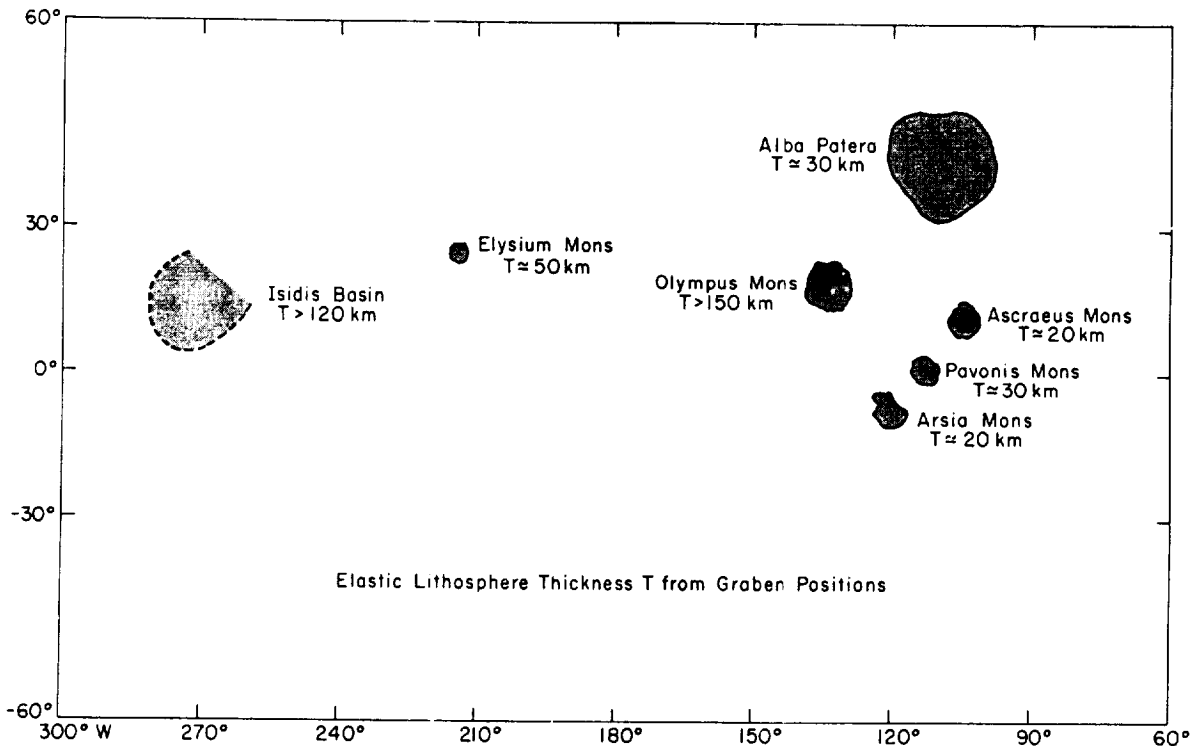


Figure 1. Summary of estimates of the thickness  $T_e$  of the elastic lithosphere beneath major volcanic loads on Mars from the radial distance of prominent circumferential graben [7]. The thicknesses shown correspond to the times at which the graben formed.

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ductile strength of the mantle, assumed to be limited by the creep strength of olivine [15]. The minimum values of  $T_e$  for the Isidis mascon and Olympus Mons correspond, by this line of reasoning, to mean lithospheric thermal gradients of no greater than 5-6 K/km.

While the heat flow and thermal structure of Mars are not known, we may compare these gradients with values derived from scaling arguments. For instance, if Mars loses heat at the same rate per mass as the Earth [16], then the mean heat flux would be about 30 mW/m<sup>2</sup>-K. For a representative value of lithospheric thermal conductivity of 2-3 W/m-K, the mean lithospheric thermal gradient would be 10-15 K/km. These figures would be reduced somewhat if the relative contributions to global heat loss of radioactive heating and secular cooling differ between Earth and Mars.

The values of  $T_e$  derived from the Tharsis Montes and Alba Patera (Figure 1) are less than or comparable to the thickness of the crust. The mechanical lithosphere thickness  $T_m$ , which exceeds  $T_e$  for these loads [3], is likely governed by the strength of crustal material, taken to be limited by the creep strength of anorthosite [17]. The mean thermal gradients consistent with the values of  $T_m$  for these loads under this assumption are in the range 11-18 K/km. The thermal gradient corresponding to the value  $T_e = 54$  km determined for Elysium Mons [7] depends strongly on the thickness of the martian crust. The required gradient decreases with increasing crustal thickness, but generally falls between those for Olympus Mons and Isidis and those for the Tharsis Montes and Alba Patera.

*Possible Causes of Lateral Variations.* As noted above, the differences in lithospheric thermal gradients implied by the different values of  $T_e$  must be at least in part due to lateral variations in temperature within and beneath the lithosphere. These variations can be due to lithospheric reheating beneath the centers of major volcanic provinces, thermal differences remaining from major pre-volcanic events such as large impacts [18], or some combination of these two effects [19]. The temperature differences, at least 400 K at 30 km depth, are too large to be solely the effect of large impacts of order 10<sup>9</sup> yr earlier [20]. They are, however, similar to the temperature variations associated with lithospheric reheating beneath hot spot volcanic centers on Earth [21]. The temperature anomalies beneath the central regions of major volcanic provinces on Mars must be similarly related to mantle dynamic processes, such as convective upwelling and magmatism. Whether the pattern of mantle dynamics on Mars during the era of volcano growth and lithospheric loading has a heritage from the earlier time of heavy bombardment remains open.

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**VOLCANOTECTONIC PROVINCES OF THE THARSIS REGION OF MARS:  
IDENTIFICATION, VARIATIONS, AND IMPLICATIONS;** Kenneth L. Tanaka and James M. Dohm, U.S. Geological Survey, 2255 N. Gemini Dr., Flagstaff, AZ 86001.

Previous histories of the Tharsis region of Mars have analyzed the sequence of faulting and volcanism in the context of regional development [e.g., 1-3]. Differences in volcano morphology as a function of location and age led Scott [4] to distinguish five volcanic provinces in Mars' western hemisphere. The recent geologic map of the Martian western hemisphere [5] shows further diversity in the volcanic and tectonic record of the area surrounding Tharsis Montes, the center of the Tharsis region. The map shows six formations consisting of local volcanic rocks (as well as many other units of possible volcanic origin) that distinguish discrete centers of Tharsis volcanic activity. Most volcanoes in the region are located on or near regional structures (Fig. 1). Volcanic centers, in turn, commonly appear to have stimulated local faulting [e.g., 6].

We and our colleagues have been examining the detailed volcanotectonic relations within three areas of the Tharsis region. Thus far, Tanaka and Davis [7] have distinguished 13 fault sets in Syria Planum (subquadrangles MC 17NE and SE), and Scott and Dohm [8] have identified 9 sets in Tempe Terra (MC 3SE, 3NE, 4SW, and 4NW) and 11 sets in Ulysses Fossae (MC 9SW and NW). We interpret less than half the sets to be made up of long regional faults that radiate from Tharsis Montes or Syria Planum; the remainder are shorter and are related to local volcanotectonic centers. Identification of these local centers, as well as of variations in volcanic and tectonic style within the Tharsis region, enables us to divide Tharsis into volcanotectonic provinces. This division provides a new perspective that will assist in discriminating between local and regional faulting and will also stimulate reassessment of the volcanotectonic evolution of the Tharsis region.

Our preliminary analysis identifies eight volcanotectonic provinces (Fig. 1). All of these provinces include faults and grabens radial to the center of Tharsis and display locally distinctive volcanic and tectonic features. Characteristic volcanic styles include broad, high shields (e.g., Tharsis Montes); broad, low centers (e.g., Alba Patera); and dispersed, old, moderate-size volcanoes (e.g., those on Terra Sirenum). Tectonic styles can be differentiated by density, morphology, and orientation of faults and by the magnitude of crustal uplift or downdrop. A summary of these characteristics for each province is provided in Table 1.

We stress that local volcanotectonic activity developed along regional structures stemming from Tharsis-centered activity, generating large volumes of volcanic material and forming local fault sets. The provinces' considerable differences in styles of volcanism and tectonism confirm the heterogeneity of Tharsis evolution. Further work is intended to complete identification of fault sets in local exposures and to distinguish regional and local fault trends. This work will lead to a refined mapping of the volcanotectonic provinces and will yield a more detailed history of the Tharsis region than that previously attained.

VOLCANOTECTONIC PROVINCES, THARSIS REGION  
Tanaka K.L. and Dohm J.M.

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**Table 1. Characteristics of Volcanotectonic Provinces in the Tharsis Region of Mars.**

Province	Volcanic style	Tectonic style
Tharsis Montes	Large, high shields; extensive flow fields	Sparse radial grabens; a few ring faults surrounding volcanoes
Alba	Broad, low shield; extensive flow fields	Dense radial and some concentric grabens surrounding volcano
Tempe	Several small, low shields; local flow fields	Dense radial and some concentric grabens; several local centers
Valles Marineris	Interior pyroclastic rocks (?); extensive plateau flow field(?)	Broad uplift; deep rifting and collapse structures
Syria-Thaumasia	Broad region of fissure activity; small, outlying volcanoes	Local uplifts; radial and concentric grabens and normal faults; collapse structures
Sirenum	Dispersed volcanoes	Broadly spaced grabens radial to Tharsis Montes
Elysium	Large shields and rille sources; extensive flow fields	Broadly spaced grabens radial to Tharsis Montes; local tension cracks
Olympus	Large shield; local fissure sources; local plains flows	Older radial grabens; concentric grabens of Acheron Fossae



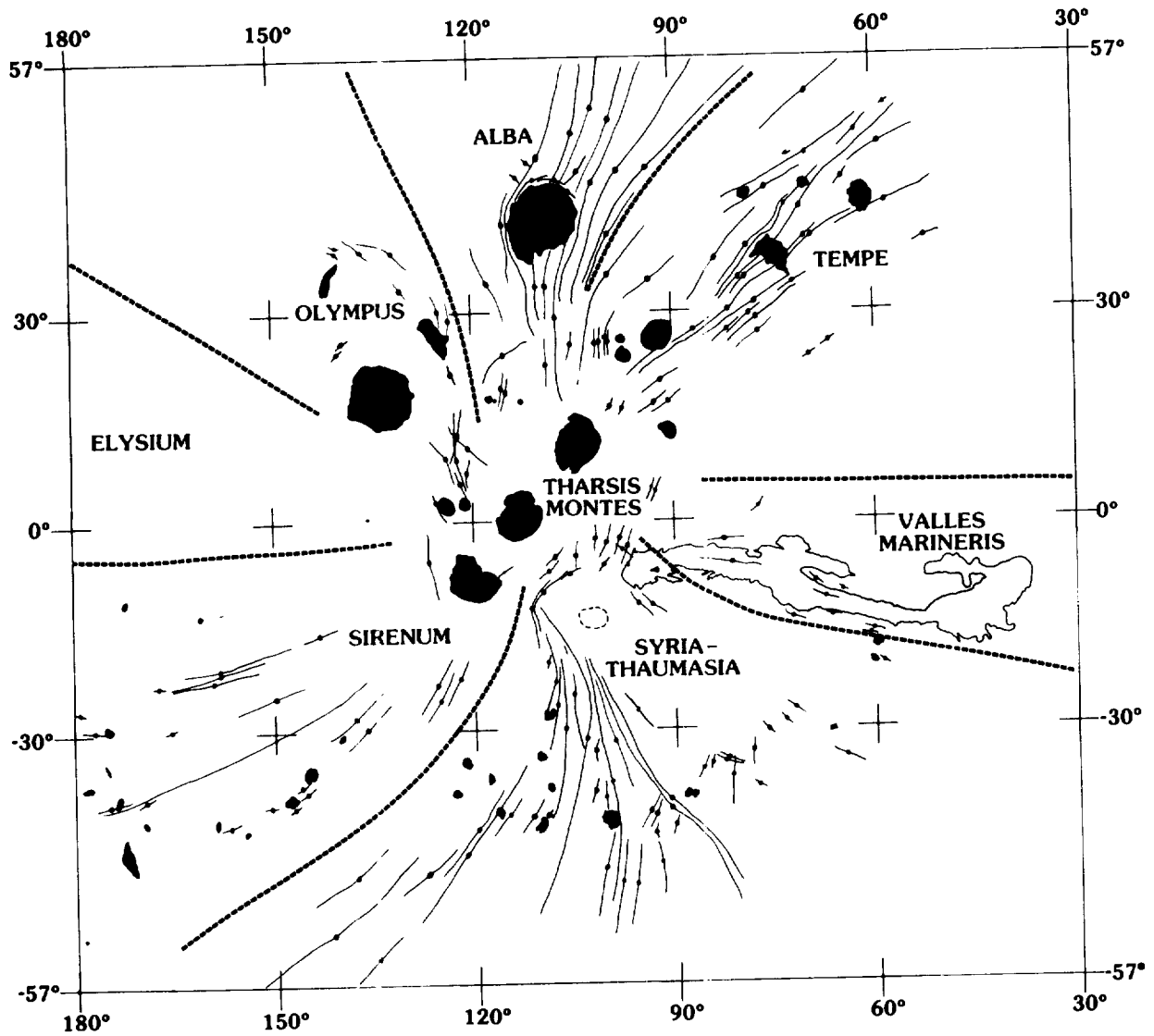


Figure 1. Volcanotectonic provinces of the Tharsis region of Mars. Heavy dashed lines delineate provinces. Solid areas are volcanoes; lines with balls are faults and grabens. Solid line outlines Valles Marineris canyon system; light dashed line encloses source area of Syria Planum flows. Volcanoes and grabens of the Elysiium province occur outside map area.

## PRIMORDIAL GLOBAL DIFFERENTIATION, MARS-STYLE

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A strong case can be made, albeit from circumstantial evidence, that in primordial times all of the terrestrial planets were substantially molten, at least in their outer few hundred km. The predominance of extensional tectonics on Mars suggests that it was less thoroughly melted than most of the other terrestrial planets [1]. Indeed, it might be argued that Mars "must" have been less intensely heated, because it is far from the Sun and apparently has a relatively low core/silicate ratio. Even so, isotopic results from the putatively martian SNC meteorites [review: 2] suggest that the planet underwent global differentiation at or very near 4.5 Ga. The density of craters in the martian southern hemisphere highlands also attests to a differentiation that was mostly completed within the first few hundred My of planetologic time. Assuming that a state of partial interior melting at least roughly comparable to a "magma ocean" developed on primordial Mars, it behooves us to examine what sort of crust would have been produced. How does the predicted crust compare with the actual ancient crust of Mars? We already have some constraints on the composition and thickness of the ancient martian crust, and impending USSR and USA Mars missions should provide many important new observations. Thus, simple predictions should be testable, if not now then at least within the next few years.

As discussed at length by Warren [3], planet size has a tremendous influence over primordial global differentiation. In large terrestrial planets, high pressures toward the bottom of the magmasphere affect the relative stabilities of the various silicate minerals. In the massive Earth, pressure-stabilized garnet and Al-rich pyroxene must have severely curtailed the ultimate yield of crust. The essential difference between planetary "crust" and "mantle" is concentration within crust of the low-pressure aluminosilicate mineral: feldspar. From a mantle perspective, the celebrated  $\text{SiO}_2$  enrichment of the Earth's continental crust is minor compared to the crust/mantle enrichment in  $\text{Al}_2\text{O}_3$ . Aluminum sequestered into deeply buried garnet and Al-rich pyroxene is aluminum that never contributes to the ultimate volume of the crust. This effect has been examined [3] with quantitative models of high-pressure fractional crystallization; including, in some models, periodic replenishment of the melt zone with fresh melt from the deep interior. Crystallization sequences were modeled as functions of both melt composition and pressure, with constraints from a variety of recent ultra-high-pressure phase equilibrium experiments, most notably the results of Takahashi [4] for peridotite KLB-1 at pressures up to 14 GPa. KLB-1 has a bulk composition remarkably similar to many estimates for the bulk composition of the Earth's upper mantle. Either suitably high-pressure partial melts of KLB-1 [4], or else (for models with initial magma ocean depth  $\geq 500$  km) KLB-1 itself, were used for initial melt compositions. The ultimate composition and thickness of the crust is calculated by assuming (as a first approximation) that all feldspar crystallized floats, and that it buoys up enough contemporaneously-crystallized mafic silicates to bring the net density of the crust up to the 0.1-MPa density of the underlying melt zone. Results indicate that a plausibly deep ( $\sim 200$ -500 km) magma ocean would produce a crust comparable in both composition and thickness to the total crust of the modern Earth.

This same model can be adapted for other terrestrial planets, including Mars. If we assume the same KLB-1 (or KLB-1 partial melt) compositions for the initial melt zones, the resultant crusts are relatively constant in composition among all the terrestrial planets, but (assuming a moderately deep initial magma ocean) the final thickness of crust is much lower for the Earth than for smaller planets, including Mars (Fig. 1A). This result reflects the much more extensive high-pressure crystallization of garnet and Al-rich pyroxene within the Earth. Depending upon assumptions regarding melt zone replenishment, garnet constitutes 0-1 wt% of the final crystallization products of a martian KLB-1/500-km model. In contrast, 47-51 wt% of the total  $\text{Al}_2\text{O}_3$  in the terrestrial KLB-1/500-km model ends up in garnet. Of course, this discrepancy in crustal yield is even greater in terms of volume percentages of the respective planets. The Earth's 21-km (global average) crustal thickness comprises only 0.9 vol% of the planet; whereas the outer 21 km of smaller Mars comprises 1.8 vol% of the planet.

Fig 1B shows results from models with initial magma ocean thickness again 500 km, but with initial melt compositions adjusted based on the estimated compositions of the different planets. Venus is too Earthlike in size, and too uncertain in composition, to warrant a separate discussion here. The Moon's magma ocean was modeled as having an initial composition corresponding to a mixture of 65 wt% Warren's [5] "Standard Initial" composition (a.k.a. "SI," derived as a model for lunar Mg-suite parent melts by simulating high-degree equilibrium partial melting of a high-mg chondritic silicates composition) and 35 wt% KLB-1. The SI composition is diluted with KLB-1 because mass balance suggests that a 500-km deep magma ocean, encompassing 64 vol% of the Moon, would have to be a higher-degree partial melt than assumed for Warren's [5] derivation of SI (pure SI would be appropriate for a lunar magma ocean 200-300 km deep). For Mercury, a 500-km deep magma ocean would probably encompass 5/6 of the entire non-core portion of the planet [1]. The bulk composition of Mercury's mantle+crust is poorly constrained, but most cosmochemists favor a composition akin to the highly reduced (i.e., Fe<sup>0</sup>-rich, FeO-poor) enstatite chondrites. For Fig. 14 the initial melt was modeled as a 65%/35% weighted mean of (a) an average for the non-metallic, non-FeS, portions of EH group enstatite chondrites [6], and (b) KLB-1. Results using a Mercury mantle+crust composition proposed by Morgan and Anders [7] are similar, except the crustal thickness result increases from 25 to 40 km.

The composition used for martian models was an average of six proposed compositions for the mantle+crust portion of Mars from Table 4.3.2d of [8]. This composition has the same Al<sub>2</sub>O<sub>3</sub> content as KLB-1, but is much richer in FeO, mainly at the expense of SiO<sub>2</sub> and MgO. Since [8], tentative identification of SNC meteorites as derivatives of Mars [2] has strengthened the case for a relatively FeO-rich bulk-planet composition. Also tested were compositions diluted with the relatively Al-rich "martian mantle minimum melt composition" of Table 11 (column 1) of [9], but such models only enhance the surprisingly high result for crustal thickness, while having no significant effect on the result for crustal Al<sub>2</sub>O<sub>3</sub>.

The different outcomes from the models with fixed initial composition (Fig. 1A), vs. the models with diverse, more realistic initial compositions (Fig. 1B), arise from mechanisms that are readily understood. In comparison to the Earth, the lunar initial melt composition is more Al-rich. This disparity arises not from any assumed difference in mantle+crust composition, but as a consequence of the vastly different pressure at which melting occurs. The same factor (low pressure) leaves more Al for late-stage melts on the Moon, and causes less extreme FeO enrichment. Less FeO enrichment translates into lower melt density, and low-density melts yield a crust that is comparatively free of rafted mafic silicates (i.e., more nearly pure plagioclase); hence the relatively high crustal Al<sub>2</sub>O<sub>3</sub> result. Mercury's "reduced" composition leads to a high proportion of pyroxene to olivine, and the pyroxene crystallization occurs mainly at pressures between 3 and 7 GPa, over which pressure range  $D_{Al}(px)$  is especially high. Extensive pyroxene crystallization severely limits the ultimate yield of Al-rich, crust-generating melt. Another consequence of Mercury's bulk composition is relatively slight FeO enrichment: FeO contents of crust-forming melts are typically 0.5-0.6 × the FeO contents of analogous Earth model melts. These low-density melts engender a crust that is comparatively free of mafic silicates, hence the high crustal Al<sub>2</sub>O<sub>3</sub> result for Mercury (this density effect also works to limit crustal thickness).

Results for Mars, which is assumed to be FeO-rich, are just the opposite: relatively little pyroxene is generated, and only a trace of garnet. The resultant high degree of Al<sub>2</sub>O<sub>3</sub> enrichment in late-stage melts leads to a remarkably thick model crust (Fig. 1B). The arbitrary choice of 500 km as the initial depth for the martian magma ocean model of Fig. 1B may be too high, in which case the crustal thickness would be commensurately overestimated. Recall, however, that a conservatively Al-poor initial composition was used. Plagioclase is especially buoyant over the extremely FeO-rich late-stage melts that the martian magmasphere generates, and consequently the crust is relatively impure (low in Al<sub>2</sub>O<sub>3</sub>). Model results for a range of initial depths and detailed compositions indicate that the magmasphere-generated crust of Mars should be decidedly less anorthositic than its counterparts on the Moon, Mercury, and even the primordial Earth (Fig. 1B).

A relatively low-Al composition for the ancient martian highlands is also consistent with the regolith compositions (average  $\text{Al}_2\text{O}_3$  in silicate fraction = 8.8 wt%) measured by Viking [10]. This regolith is probably well mixed on a global scale by dust storms. Assuming the average  $\text{Al}_2\text{O}_3$  of the lowlands is at least 4 wt%, and at least 50% of the regolith's mass is derived from the highlands, the average  $\text{Al}_2\text{O}_3$  content of the highlands (which cover roughly 60% of the planet) is unlikely to be  $\gg 14$  wt%. A relatively thick crust for Mars is also favored by the scant observational evidence currently available: Anderson et al. [11] infer from Viking seismic results that the mean global thickness of crust is at least 30 km, and gravity modeling suggests that crustal thickness in the Elysium dome region (possibly an area of atypically thick crust) is roughly 50-80 km [12]. Acquisition of more precise constraints on the composition and thickness of the martian crust will allow more definitive judgements regarding its origin and evolution.

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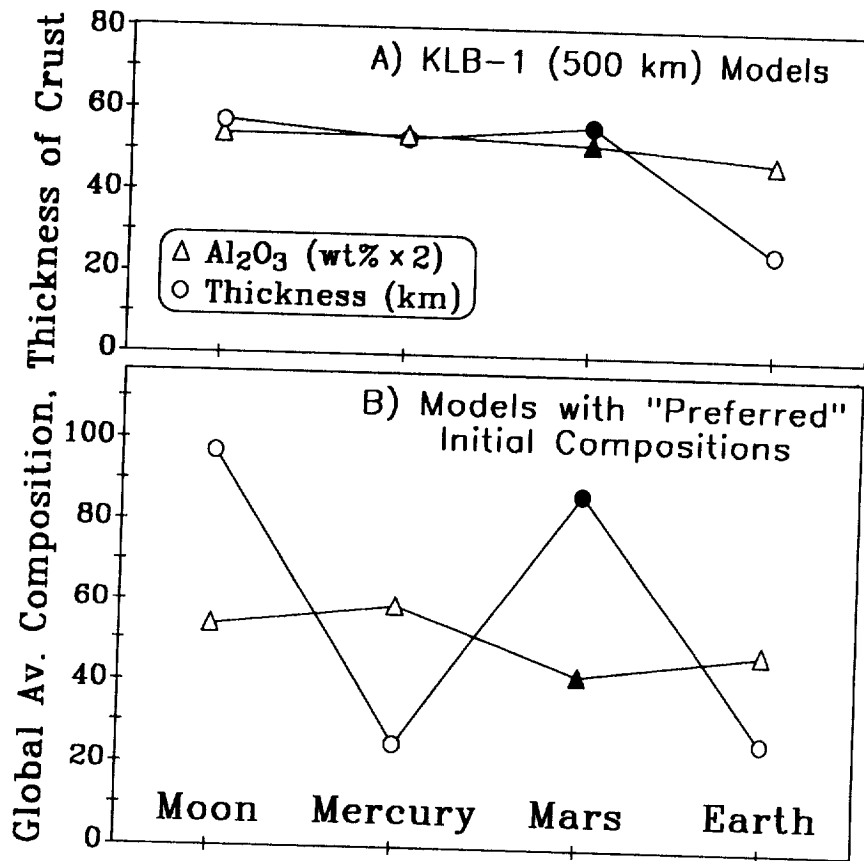


Fig. 1

PERIODICALLY SPACED WRINKLE RIDGES IN RIDGED PLAINS UNITS ON MARS. Thomas R. Watters, Center for Earth and Planetary Studies, National Air and Space Museum, Smithsonian Institution, Washington, D.C. 20560.

Ridged plains units cover over 3% of the surface of Mars and range in age from Noachian to middle Hesperian. These units are characterized by smooth plains and the presence of landforms analogous to mare wrinkle ridges. Although the exact nature of the ridged plains material has yet to be determined, photogeologic evidence of volcanic landforms, comparisons with terrestrial flood basalt provinces and lunar mare, proximity to major volcanic centers, and presence within large impact basins suggests that they are the result of flood volcanism (1, 2, 3). The origin of the wrinkle ridges has been the subject of a number of recent papers and some debate over the role of buckling and/or reverse or thrust faulting (3, 4, 5). However, the general consensus is that wrinkle ridges are tectonic in origin resulting from horizontal compressive stresses.

A characteristic of the ridged plains first noted by Saunders and Gregory (6) is the periodic nature of the wrinkle ridges. The regular spacing of the co-parallel ridges has been analyzed by dividing ridged plains provinces into domains based on factors such as: 1) ridge orientation; 2) borders with major tectonic features and; 3) contacts with other geologic units. Statistics on the ridge spacing were determined using a series of sampling traverses spaced at roughly 12 km intervals oriented perpendicular to the mean orientation of the ridges within a defined domain (Table 1).

The periodic spacing of the ridges suggests a deformational mechanism involving a dominant wavelength. Viscous buckling models have been suggested (6, 7) but these models ignore the influence of gravity and relevant boundary conditions. The model proposed here assumes that: 1) deformation has occurred at the free surface; 2) the ridged plains are a multilayered sequence that behaves as a linearly elastic material resting on a mechanically weak regolith of finite thickness which is in turn resting on a rigid boundary (figure 1); 3) gravity is not a negligible factor. The model also assumes that units of flows are separated by thin regolith interbeds that allow free slip between the individual units. This is consistent with evidence of subsurface radar reflectors detected in the ALSE data over Mare Serenitatis and Crisium interpreted to be deep-lying density inversions consisting of a regolith layer (8, 9). Recent data from deep wells near several of the anticlinal ridges of the Columbia Plateau, which are good analogs to planetary wrinkle ridges (3), show evidence of sedimentary interbeds separating groups of individual flows. Based on these assumptions the critical wavelength  $\lambda_c$  is given by:

$$\lambda_c = 2\pi[(h_0 \text{Ent}^3 / 12(1-v^2)E_0) + (\text{Ent}^3 / 12(1-v^2)\rho_s g)]^{1/3}$$

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where  $E$  and  $E_0$  are the Young's modulus of the plate and substrate respectively,  $h_0$  is the thickness of the substrate,  $t$  is the thickness of an individual group of flows,  $n$  is the number of groups,  $\rho_s$  is the density of the substrate,  $g$  is the acceleration due to gravity and  $\nu$  is Poisson's ratio. The critical wavelength as a function of the ratio of elastic modulus  $E/E_0$  for values of  $t$  of 250, 350 and 500 m using an estimated total thickness of the ridged plains material of 4 km and regolith substrate of 4 km is given in figure 2. Many of the observed ridge spacings can be explained provided that the contrast in elastic modulus between the ridged plains material and the regolith  $E/E_0$  is on the order of 500 to 5,000. Such a high range of  $E/E_0$  is not unreasonable given the degree of brecciation and high porosity of a typical lunar regolith (10). The critical stress  $\sigma_c$  to achieve buckling is given by:

$$\sigma_c = [ (tEE_0/n^3(1-\nu^2)h_0) + (tE\rho_s g/n^3(1-\nu^2)) ]^{1/2}$$

Critical stresses for most of the wavelengths shown in figure 2, between an  $E/E_0$  of 500 to 5,000, range from about 0.6 to 1.8 kbars. Assuming the ridged plains material is basalt-like, the compressive strength of the material at a depth of 4 km using Byerlee's law is approximately 1.7 kbars. Thus, based on this model the ridged plains material would be expected to deform at least initially by buckling when subjected to a sufficiently large horizontal compressive load.

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Table 1.

Location	Mode (km)	Observations
Amazonis	30	472
Lunae Planum	30	940
Coprates	50	900
Chyrse	40	308

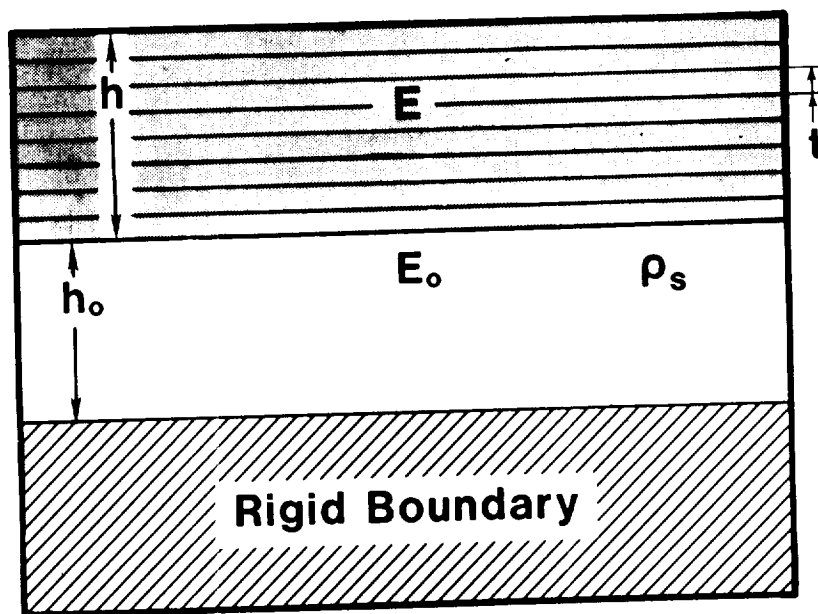


Figure 1.

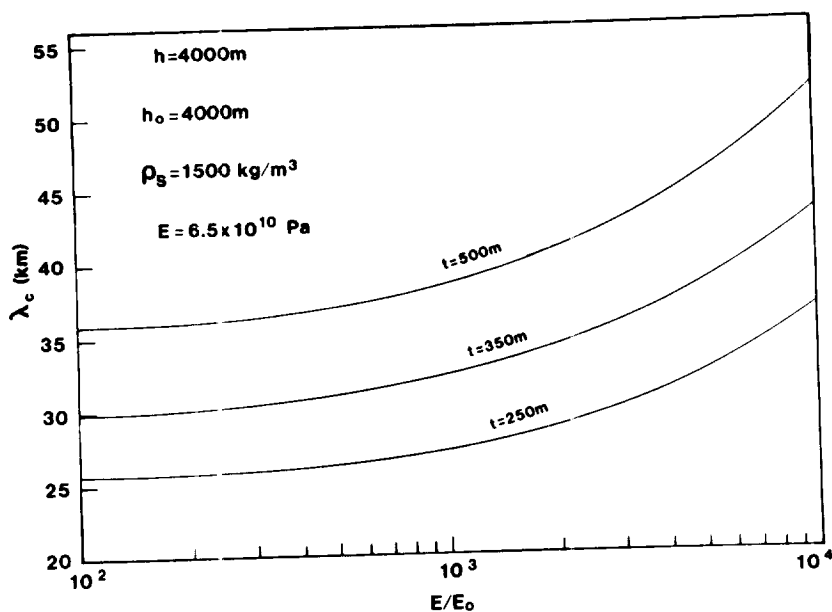


Figure 2.

AN ANCIENT VALLES MARINERIS?; R.W. Wichman and P.H. Schultz, Dept. of Geological Sciences, Brown University, Providence, RI 02912.

Is Valles Marineris (VM) on Mars unique to processes associated with Tharsis relatively late in martian geologic history? Or is it only the last and best preserved major canyon system on the planet? A shallow trough system concentric to Hellas (1, 2, 3) may represent such an ancient canyon system comparable in width and extent to Valles Marineris. In this contribution we compare these two systems and consider the proposal (4) that Valles Marineris is an analogous Chryse-centered trough system rejuvenated by mobilization of trapped ground-ice during Tharsis tectonism/volcanism.

The shallow concentric Hellas canyons (HC) occur between 1900 and 2500 km from the basin center, outside both the extrapolated Hellas boundary scarp and the associated topographic expression of the Hellas basin (5). The individual troughs range from 30–100 km in width and from 300 to 800 km in length. The bounding scarps are generally slightly furrowed in appearance and rectilinear in plan with reliefs on the order of 0.5 km from earth-based radar data (6). To the south, the scarps lose relief and merge into the cratered plains, although ridge/scarp systems on Malea Planum (Chalcoporos and Pityusa Rupes in particular) may represent a continuation of the trend. To the northeast, the trend is lost near the crater Huygens, perhaps overprinted by Huygens-related fractures. The troughs have been floored by intercrater plains units, and a back-slope away from the troughs is suggestive of block rotation or relaxation after canyon formation.

The geometric similarity between the HC and VM relative to the Hellas and Chryse impact basins, respectively, are documented in Figure 1. The Hellas canyons are 1900–2500 km from the Hellas basin center, whereas the Valles Marineris canyons exhibit elements 1850–2500 km in distance from and concentric with the Chryse basin center (4). The characteristic lengths of the two systems are also similar: the Hellas canyons extending over 2600 km; Valles Marineris extending ~3300 km before disappearing among the chaotic terrains arching around Chryse to the east. Although Valles Marineris exhibit a broad spectrum of canyon widths in general (from 50 to 150 km), less modified elements mimic the size distribution of the narrower Hellas canyons (Figure 2). The histograms in Figure 3 further indicate a correlation in both continuous scarp length and linear segment length of the HC structures with observed lengths in the shorter, discontinuous VM canyons (Gangis, Juventae, Hebes). Deviations of boundary scarp strikes relative to the mean system strike between the two canyon systems are also similar.

The HC scarps are more subdued than the VM canyon walls, but possess similar spur and groove textures. Although the plains units cut by the VM canyons are regionally elevated along the VM trend (7), a raised lip along the VM walls is apparent in some topographic maps (8), thereby indicating local back slopes along the VM similar to those observed in the HC. Gravity analysis of the VM further indicates a mass excess at depth and poor compensation of the mass removed by canyon formation (9). Relative uplift due to compensation along the canyon would result in apparent rotation of the bounding scarps and produce a feature analogous to the HC troughs.

The most apparent difference between the two systems is the canyon depth: VM canyons typically 4 to 8 km (8) with HC canyons only ~0.5 km (6). This may partially reflect the much older age of the Hellas system. Crater counts and stratigraphic relations along the HC scarps yield crater ages of ~1000–2000  $N(D>5)/10^6 \text{ km}^2$ , whereas VM cuts volcanic plains units with  $N(>5)$  ages as young as 125–200 (9) with possible modern volcanic activity along the VM fault scarps (10). The oldest age for



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VM has been largely consumed by backwasting although the associated outflow channels date from an N(>5) age of about 27 (11) and very ancient subparallel troughs extend to pre-Tharsis times (12, 13).

**Discussion:** We have modeled (3, 14) the Hellas canyons as impact-related fracturing of a lithosphere about 120 km thick. The geometric similarity of the Valles-Marineris/Margaritifer-Chaos system and the Hellas system with respect to basin centers (Figure 1) is consistent with a similar mechanism for the development of precursory weaknesses about Chryse. In profile, the two systems are similarly located relative to the topographically defined central impact basins. If scaled to the massif rings, however, the VM system is slightly farther from the basin center than the HC, perhaps due to thickening of the lithosphere with time or variations in global thickness.

The formation of Valles Marineris is thought to result primarily from an enlargement of graben and/or rift structures by mass-wasting, modified by subsequent deposition of interior fill deposits (15). The initial rifting episode is frequently cited as related to the evolution of the Tharsis rise to the west (7, 15) and to the Tharsis-radial trend of Valles Marineris (16, 17). Tanaka *et al.* (9) proposed more specifically that the VM developed from an initial elongate thermal anomaly beneath the VM region. Continued erosion and thinning of the lithosphere eventually mobilized ground ice by increased heat flow and magma. Thermal uplift led to doming and rifting near the surface and allowed volatile release, which further enhanced the surface expression of the rifting process.

Alternatively, the Chryse impact established a concentric zone of lithospheric failure analogous to Hellas. Subsequent erosion filled these canyons with a sequence of sedimentary and ejecta deposits capable of storing the water released by impact or volcanism into the early martian environment. After regional plains volcanism of Lunae Planum and Sinus Planum capped these units, the original impact canyon sequence would be preserved as volatile-rich reservoirs in specific zones about Chryse. Thermal reactivation associated with Tharsis would subsequently localize renewed fracturing and heating into an elongate region as in (9) and release the volatile reserves stored since the time of Chryse formation. Thus, thermal evolution during Tharsis construction could rejuvenate an older impact fracture system as the VM canyons through the mobilization of specific volatile traps.

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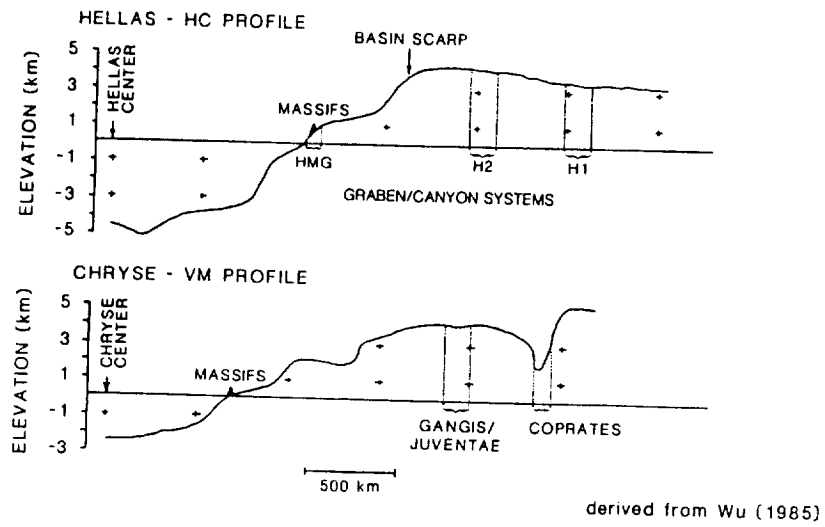


Figure 1. Basin Profiles of Hellas and Chryse with the Outer Canyon Forms.

Figure 1a. Hellas topographic profile derived from (5) extending northeast from Hellas with observed basin features and basin-concentric tectonic systems. HMG=grabens in the Hellespontes Montes; H1, H2 = systems of the Hellas canyons.

Figure 1b. Chryse topographic profile (5) extending south from Chryse between Gangis and Juventae Chasma (located on profile) to beyond Coprates Chasma.

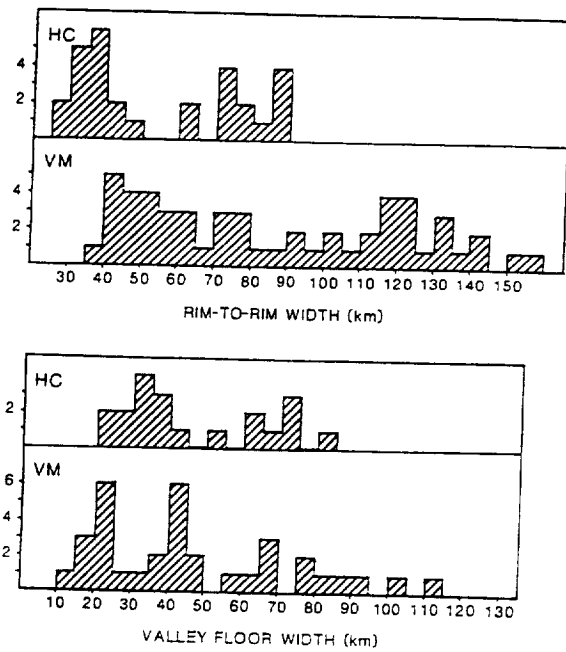


Figure 2. Comparison of VM and HC Feature Widths.

Figure 2a. Histograms of the rim-to-rim widths of the VM and HC canyon systems measured from the top of the canyon wall scarps at 55-60 km intervals.

Figure 2b. Histograms of the valley floor widths measured between the bases of the canyon wall scarps at intervals of 55-60 km. Note the common peaks at 20-40 km and 60-70 km widths.

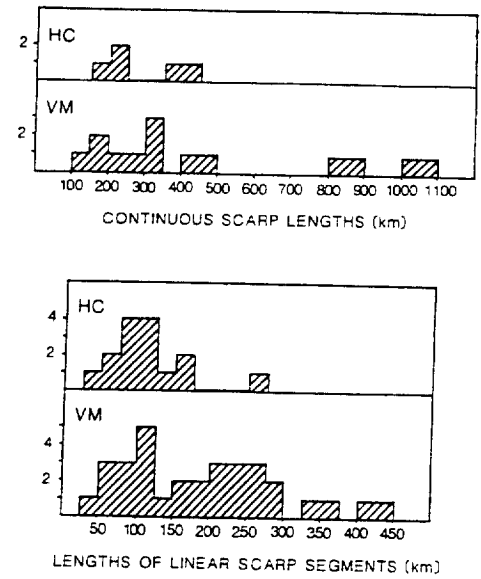


Figure 3. Comparison of VM and HC Scarp Lengths.

Figure 3a. Histograms of continuous scarp lengths comprising the VM and HC systems, measured at the top of the canyon wall scarps.

Figure 3b. Histograms of the length of linear scarp segments comprising the scarps of figure 3a.

**THE RELEVANCE OF KNOBBY TERRAIN TO THE MARTIAN DICHOTOMY**

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**The problem:** The martian uplands consist of a moonlike basement of densely packed impact craters and basins partly covered by diverse, unmoonlike deposits. The lowlands contain relatively young, sparsely cratered deposits of mixed origins. The generally southern uplands are separated from the generally northern lowlands by a distinct scarp or slope that encompasses the entire planet, except where truncated by younger basins or craters and buried by extensions of the lowland deposits. The resulting hemispheric dichotomy is a fundamental feature of Mars. Upland materials along the upland-lowland front have been reshaped into a variety of landforms, including fault blocks, mesas, knobs, and landslides. Similarly modified inliers of the uplands occur within the lowlands, separated from the uplands proper by intervening young deposits. Knobs are the most common modifiers of the inliers.

This similarity of the upland inliers to the frontal transition zone has led to a common conclusion that the same process not only modified but also created the lowlands and the front. One idea was that the uplands were converted into lowlands by erosion, but the problem of where the eroded material went was insoluble. Wise et al. [1] then suggested that the dichotomy resulted from an early first-order convection of the mantle. However, this conjecture appears to have been offered because all else had failed, and was not supported by evidence from any planet for such a massive endogenic redistribution of mass. To resolve what appeared to be an impasse in interpreting the dichotomy and the front [2], Wilhelms and Squyres [3] proposed a mechanism in keeping with the known early history of the Solar System: a giant impact. In this view the main northern lowlands occupy the interior, and the uplands the exterior, of the 7,700-km Borealis basin. The upland-lowland front, though since modified, was fixed in its basic position by the basin rim. Masifs and other concentric features support the basin's existence and indicate a center at about 50°N., 190°W.

Subsequent studies have addressed this hypothesis, mostly with skepticism [4-6] though with growing approbation [7]. A major barrier to its acceptance apparently lies in the belief that the inliers are inconsistent with the existence of Borealis. However, uplands in the geologic sense of a certain combination of craters, basins, and deposits must underlie the entire lowlands because Borealis is the basement on which all other martian features were formed. The basin is the largest member of a population defined by the size-frequency distribution of the smaller basins (despite [6]).

**Knobby terrain:** Knobby terrain occurs in two main settings: (1) along about half of the upland-lowland front and (2) in the lowlands north of the front. It has long been clear that so much of the lowland knobby terrain is arranged in circular patterns,

## KNOBS AND THE DICHOTOMY

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similar except in their knobby textures to those of the uplands, that knobby terrain must have been formed on or from upland terrain. We have dated knobby terrain by crater counts in its main lowland occurrences, Elysium and Amazonis Planitia, 0°-30°N., 157.5°-202.5°W., and distinguished three units based on spacing between knobs: high (HDK), intermediate (IDK) and low (LDK) density knobs (Table 1; updated from [8]). Crater rims are defined most distinctly by HDK. LDK includes much young inter-knob material unrelated to the knobs. We distinguished craters that predate the knobs from those that postdate them. The older craters' deposits are themselves broken into knobs whereas the superposed craters have complete ejecta blankets.

**Comparison terrains:** We have done similar counts of modified and superposed craters for two types of terrain in the relatively intact uplands south of the knob-study area and the front [9] and for a knobby, avalanche-covered deposit along the front (FK; centered at 5°S., 197°W.). One upland terrain is "primitive," that is, composed of densely packed craters and basin rings and lacking detectable superposed deposits (PT). Another contains valley networks (VT). In VT, a basement like that of PT has been partly covered by a later deposit in which the valleys were incised [9]. From the crater counts, we can compare both the oldest observed ages and the times of modifications of upland materials on both sides of the front (Table 1).

"All" craters show the minimum age of a terrain's substrate. HDK is comparable in density of "all"  $\geq 16$ -km craters to PT south of the front. Thus, HDK is ancient terrain that has been converted partly into knobs. IDK and VT are similar in densities of "all" craters. This similarity, as well as the presence of similar extensive tracts with few craters in IDK and with few craters but many valleys in VT, suggests that IDK was created out of an upland terrain having a superposed intercrater deposit like that of VT. The "all" counts for FK and LDK show that these units also originated as uplands.

The density of superposed craters shows the time that has elapsed since a terrain was modified. PT has the highest density of fresh craters, showing that it was modified first and least, if at all. LDK was modified the most recently, as is also obvious from its young, smooth, interknob deposits. The other four units are all similar in modification age.

**Conclusions:** "Upland" geology occurs in the lowlands as well as in the topographic uplands because the Borealis basin predates all other known martian features. VT, FK, HDK, and IDK all acquired their present character about the same time, the Early Hesperian Epoch ("Lunae Planum age"). This date of modification was among those detected by [10] by different methods. We believe that valleys form internally in an ice-rich deposit superposed on the primitive basement [9], and suggest here that IDK includes an originally similar deposit that has lost most or all of its ice. Craters and basins in the knobby terrain have resisted knob formation just as craters in the VT have resisted valley formation.

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Therefore, starting with similar raw material, the Early Hesperian events produced knobs north of the front and valleys south of it. We have interpreted the "event" in the uplands as warming by igneous intrusions [9]. Here we suggest that the knobby terrain was created when its ice was degraded by the higher heat flow in the Borealis basin interior. The difference in upland terrains in the two halves of the martian dichotomy supports rather than refutes the existence of the Borealis basin.

Table 1. Cumulative numbers of craters/10<sup>6</sup>km<sup>2</sup> for units defined in text. Ages [11] old to young: LN, Lower Noachian; MN, Middle Noachian; UN, Upper Noachian; LH, Lower Hesperian; UH, Upper Hesperian.

	≥5 km	≥16 km	Area (x 10 <sup>6</sup> km <sup>2</sup> )	Age
<u>Superposed craters</u>				
LDK	73 ±13	5 ±4	4.739	UH
IDK	145 ±12	20 ±6	6.160	LH
HDK	165 ±25	20 ±10	2.048	LH
FK	162 ±25	38 ±12	2.467	LH
VT	175 ±23	29 ±8	2.328	LH
PT	300 ±17	83 ±9	1.083	UN
<u>All craters</u>				
LDK	≈190	43 ±10	4.739	UN
FK	272 ±33	89 ±19	2.467	UN
IDK	525 ±23	151 ±13	6.160	MN
VT	537 ±15	172 ±9	2.328	MN
PT	665 ±26	235 ±15	1.083	LN
HDK	935 ±65	210 ±35	2.048	LN

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NOACHIAN FAULTING IN THE MEMNONIA REGION OF MARS.  
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Correlation of results from photogeologic mapping and Earth-based radar measurements of topography has identified several normal faults, each possessing hundreds to thousands of meters of vertical relief, within the ancient materials present in the Memnonia region near Mangala Valles (1). While the coverage of Earth-based topographic data for Memnonia is somewhat limited, the geomorphic characteristics of the faults next to Mangala Valles have been used to identify comparable faults throughout the Memnonia region ( $0^{\circ}$  to  $30^{\circ}$  S,  $145^{\circ}$  to  $165^{\circ}$  W; see Fig. 1). The faults are quite numerous and are all oriented approximately north-south; this orientation indicates that the faults probably are not related to the Tharsis uplift.

Geologic mapping of the area around the upper reaches of Mangala Valles, at a scale of 1:500,000, is under way as part of the Mars Geologic Mapping program (2). Results of the mapping indicate that the Mangala Valles channel system experienced at least three distinct flooding events that spread channel-related deposits and erosional scour over large portions of the highlands (1,2). The course of the Mangala Valles flooding events was constrained by the significant relief associated with normal faults in the Noachian materials (the oldest period on Mars; 3) of the southern highlands. Earth-based radar measurements (described in 4) indicate that the faults around Mangala Valles have 1 to 2.5 km of vertical relief (Fig. 2). West of Mangala Valles the cratered terrain is tilted toward the east and, if an aquifer was formed within the highland materials, the flood waters that produced the channel system may have been supplied through plains tilted during the period of north-south normal faulting (5). It is possible that an unnamed channel system along  $161^{\circ}$  W (Fig. 1) also may be the result of groundwater movement controlled by the major faults in the area.

The crater retention age for the earliest Mangala flood deposit is lower Hesperian, much older than the young Amazonian materials of the Tharsis area but younger than the Noachian materials in the highlands (3), so that the faulting likely predates most of the Tharsis Montes materials (1). This conclusion is in agreement with the results of Schultz (6) who examined several structural landforms in the heavily cratered terrain of Mars, most of which showed no relationship to the stress fields predicted for proposed formation processes for the Tharsis uplift. The distribution of ancient faults in Memnonia (Fig. 1) does not support the presence of an ancient impact basin under the present Tharsis uplift (suggested in 7). The ancient faults are most likely related to forces within the early martian crust, perhaps due to thermal stress accompanying a cooling lithosphere (6).

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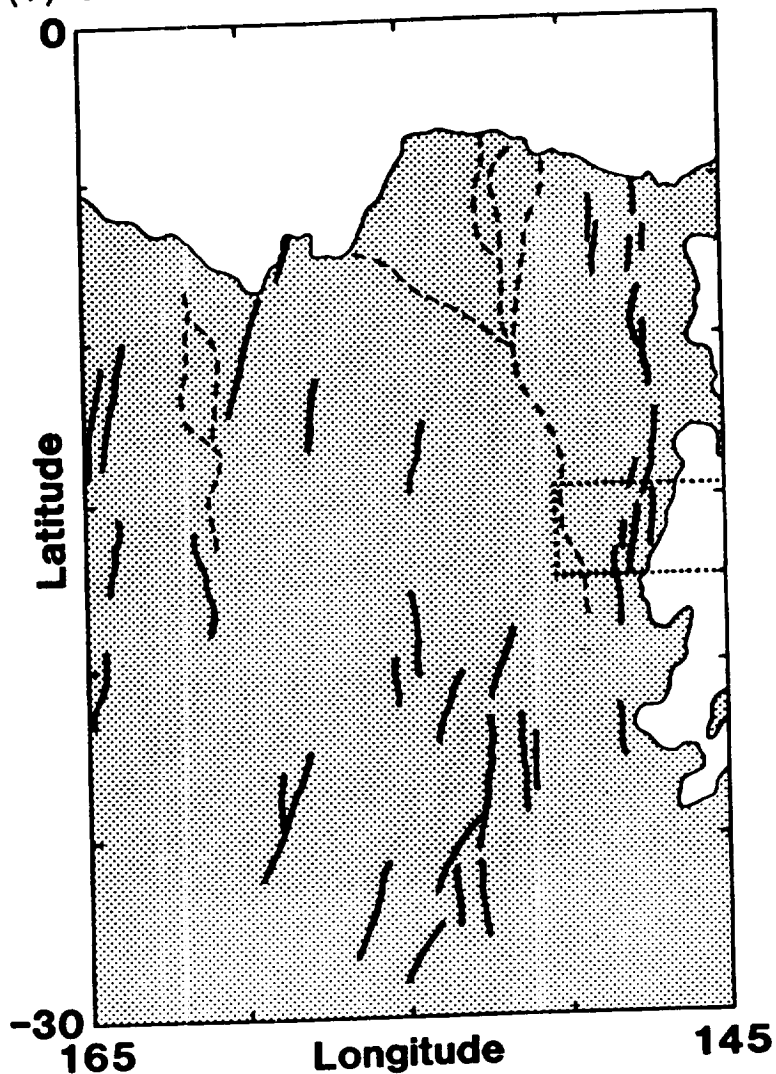


Figure 1. Location of major faults (thick lines) within the ancient highland materials of the Memnonia region (shaded region). The unshaded region north of the highland materials consists of Amazonian age materials of the Medusae Fossae Formation and the unshaded region east of the highland materials consists of Amazonian to Hesperian age materials of the Tharsis Montes Formation; the highland materials are primarily Noachian in age (3). Dashed lines indicate channels produced by flow toward the north; the channels between 149° and 157° W form Mangala Valles but the channels along 161° W are unnamed. Volatile storage and release for both channel systems may be controlled by the major faults. The dotted box shows the location of the topographic profiles in Figure 2.

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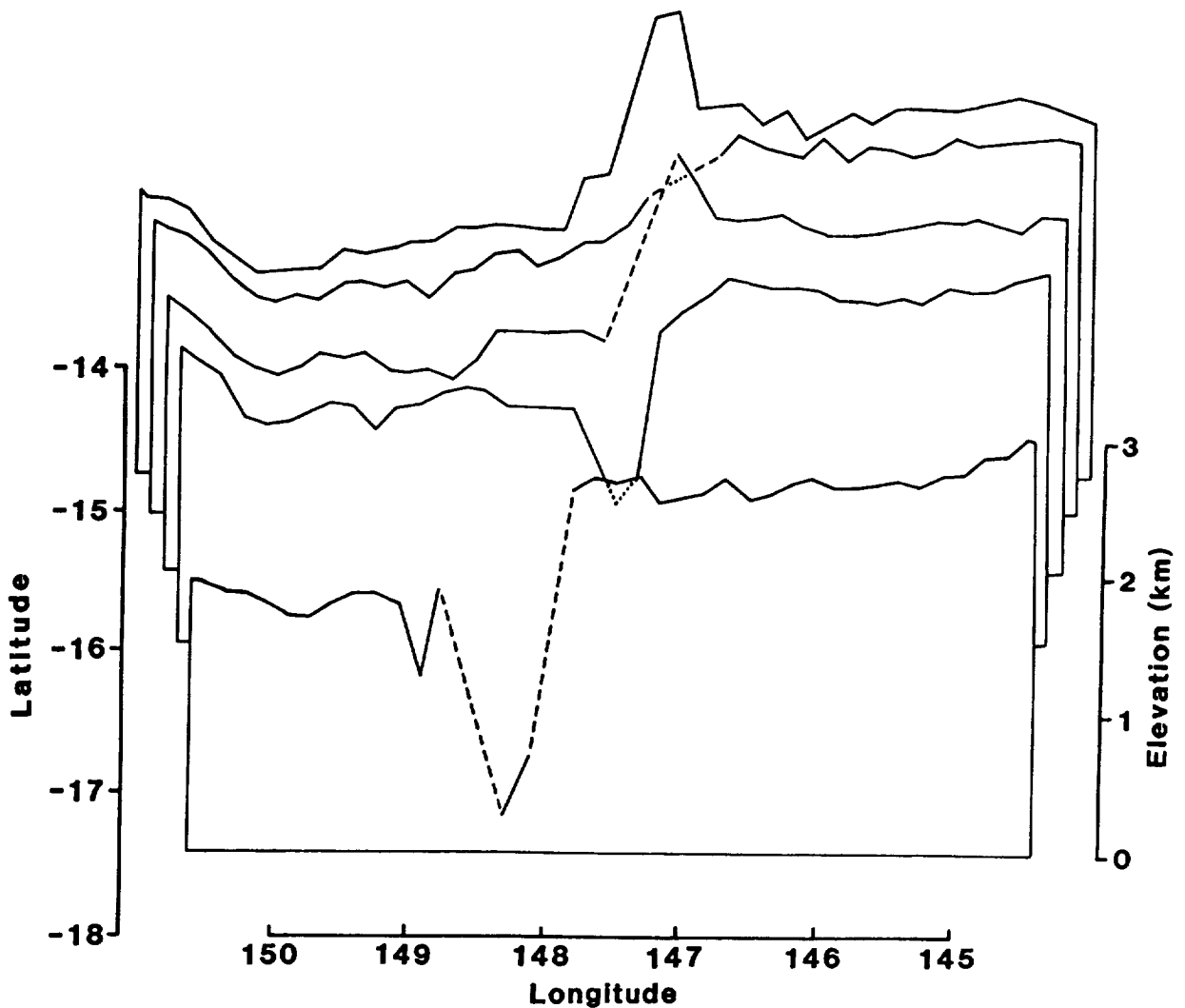


Figure 2. Five topographic profiles obtained from Earth-based radar measurements (4). The profiles have a vertical exaggeration of 57X (elevation above the 6 mbar level is shown at right) and the bottom of each profile corresponds to the data groundtrack (Latitude shown at left). The greatest relief is associated with faults within the highland materials. Mangala Valles coincides with the rather subtle topographic low along  $150^{\circ}$  W. The profiles are dashed across locations where topographic data were not available, most likely due to the large local relief, and dotted where covered by a preceding profile in this particular display. A radar resolution cell is  $0.16^{\circ}$  in longitude by  $1.3^{\circ}$  in latitude, centered along the groundtrack.



**THE SHALLOW STRUCTURE OF THE LITHOSPHERE IN THE COPRATES AND LUNAE PLANUM REGIONS OF MARS FROM THE GEOMETRIES OF VOLCANIC PLAINS RIDGES; M.T. Zuber<sup>1</sup> and L.L. Aist<sup>2</sup>, <sup>1</sup>Geodynamics Branch, Code 621, NASA/Goddard Space Flight Center, Greenbelt, MD 20771, <sup>2</sup>Department of Mathematics, University of Maryland-Baltimore County, Catonsville, MD 21228.**

Wrinkle ridges are common tectonic features on Mercury, the Moon, and Mars and are frequently found in volcanic plains units of these bodies [e.g. 1 and references therein]. On Mars, some of the most prominent assemblages of volcanic plains ridges occur in the Coprates and Lunae Planum regions [2]. These ridges most likely formed due to stresses associated with the response of the lithosphere to the Tharsis load [3-5], and form a nearly concentric pattern with a regular radial spacing of approximately 50 km [4, 6, 7]. Individual ridges have widths of approximately 2-15 km and vertical relief of 200-1100 meters [8].

A number of attempts have been made to infer the local structure of the martian lithosphere on the basis of the regular spacing of the ridges. Saunders et al. [9] applied the theory of folding in a viscoelastic medium [10] and concluded that ridge spacing is controlled by the thickness of the competent or strong volcanic surface unit and the viscosity contrast between the volcanics and an incompetent or weak semi-infinite substrate; they interpreted the substrate as the martian megaregolith. On the basis of the wavelength/layer thickness ratio predicted by the theory and the thickness of volcanics [ $\approx 1-2$  km; Ref. 11], they determined that the viscosity contrast between the volcanics and the substrate was approximately 500. Watters [12] alternatively suggested that the ridge spacing is controlled by a 15 km thick competent near-surface layer on the basis of the assumption that the dominant wavelength/layer thickness ratio ( $\lambda_d/h$ ) for elastic thin plate buckling is about four. More recently, Watters [7] recognized that the megaregolith could be more realistically represented as a layer of finite thickness rather than an infinite halfspace, but he formulated the relevant elastic buckling problem incorrectly. From his results he concluded, in agreement with Saunders et al. [9], that ridge spacing is controlled by the thickness of the volcanic surface layer.

We contend that because so little is known about the shallow subsurface structure of Mars it is necessary to consider a range of possible mechanical structures that could be consistent with the spacing of the ridges. Therefore, we are currently developing a suite of compressional deformation models that incorporate many features that have not been addressed in previous studies. On the basis of the current knowledge of the subsurface structure of Mars in the vicinity of the ridges, we invoke a baseline model of the lithosphere that consists of a thin, competent volcanic surface layer that overlies an incompetent megaregolith and a competent crustal lithosphere. We use the spacing of the ridges and the known thickness of the volcanic units as the primary constraints on the vertical rheological structure.

Thus far we have investigated two classes of simple models. The first consists of a strong layer underlain successively by a weaker layer and a strong halfspace, all of uniform strength. In this case the entire lithosphere is free to deform. This corresponds to a situation in which stresses of sufficient magnitude to induce deformation penetrate deeper than the base of the megaregolith. In the second case the lithosphere consists of a strong surface layer underlain successively by a weak layer and a rigid boundary, which represents a scenario in which deformation is confined to very shallow depths. For each of these models we have explored the effects of layer thickness, rheology, layer strength contrasts, and the effect of gravity. The rheologies that we have considered, assuming a relationship between strain rate ( $\dot{\epsilon}$ ) and stress ( $\sigma$ ) of the form  $\dot{\epsilon} = A\sigma^n$ , are

Newtonian viscosity ( $n=1$ ), non-Newtonian viscosity ( $n=3$ ), and perfect plasticity (infinite  $n$ ). The linear viscous solutions are equivalent to elastic solutions if velocities are replaced by displacements and viscosities by moduli of rigidity.

Figure 1 shows an example of the results for a uniformly compressing linear viscous model lithosphere ( $n_{1-3}=1$ ) with a strong surface layer, weak subsurface layer, and a strong substrate that is free to deform (the subscripts 1, 2 and 3 correspond to the surface layer, subsurface layer and substrate, respectively). The figure illustrates how the predicted dominant wave number ( $k_d'$ ) and wavelength ( $\lambda_d/h_1$ ) vary with the thickness of the weak layer ( $h_2$ ) for three values of the strength contrast between the surface and subsurface layers  $R_1(=\tau_1/\tau_2)$ . The strength contrast between surface layer and substrate  $R_2(=\tau_1/\tau_3)$  is assumed fixed at 0.5. Note that large  $R_1$  or  $h_2$  yield wavelength/layer thickness ratios that are compatible with the observed ridge spacing. For  $R_1=100$  or more, a regolith thickness of 5 km or greater yields acceptable solutions. This range of regolith thickness is at the upper limit of, but consistent with, current estimates determined from crater and volatile studies [e.g. 13]. The solutions are not particularly sensitive to  $n_2$ ,  $n_3$  or  $R_2$  for  $R_2 < 1$ .

The dominant wavelength for a model lithosphere consisting of a strong layer underlain by a weak layer and a rigid base is less than that for a layer underlain by a semi-infinite halfspace. Figure 2 illustrates this for a plastic layer ( $n_1=10^4$ ) that overlies a non-Newtonian layer ( $n_2=3$ ). The figure plots  $k_d'$  and  $\lambda_d/h_1$  as a function of  $h_2$ . In these calculations it was assumed that deviatoric stresses in the lithosphere are weak enough such that gravitational effects are important. Large  $h_2/h_1$  corresponds to the limiting case of an infinite halfspace. Note that none of the conditions shown in Figure 2 can explain the spacing of the ridges. For a plastic surface layer,  $\lambda_d/h_1$  is insensitive to changes in  $R_1$  [e.g. 14].

In all of the cases that we have examined with a plastic rheology, either the wavelength/layer thickness ratios are too small to explain the spacing of the ridges or the models yield multiple wavelengths of deformation. Since a perfectly plastic rheology is a continuum approximation of a medium that undergoes deformation by pervasive faulting, we conclude that such rheological behavior cannot account for the spacing of the ridges.

We can make several other preliminary conclusions about the structure of the martian lithosphere in the vicinity of the regularly spaced volcanic plains ridges. First, if the volcanic layer displayed viscous ( $n \approx 1-3$ ) or elastic behavior at the time of ridge formation, then an underlying weak regolith of uniform strength must have been at least several km thick. Furthermore, the strength of the regolith must have been at least an order of magnitude less than that of the volcanics. For a given  $\lambda_d/h_1$ , the strength contrast between the volcanic and regolith layers "trades off" with the thickness of the weak layer in a manner such that a greater strength contrast is compatible with a thinner regolith. Another implicit result is that if the megaregolith was thin (on the order of 1 km or less), then the thickness and mechanical properties of the volcanics could not have played a central role in determining the ridge spacing. The length scale in this case would have been controlled primarily by the thickness and material properties of the sub-regolith lithosphere. This would be interesting in light of the fact that the ridges (not only on Mars, but on the Moon and Mercury as well) form preferentially in volcanics.

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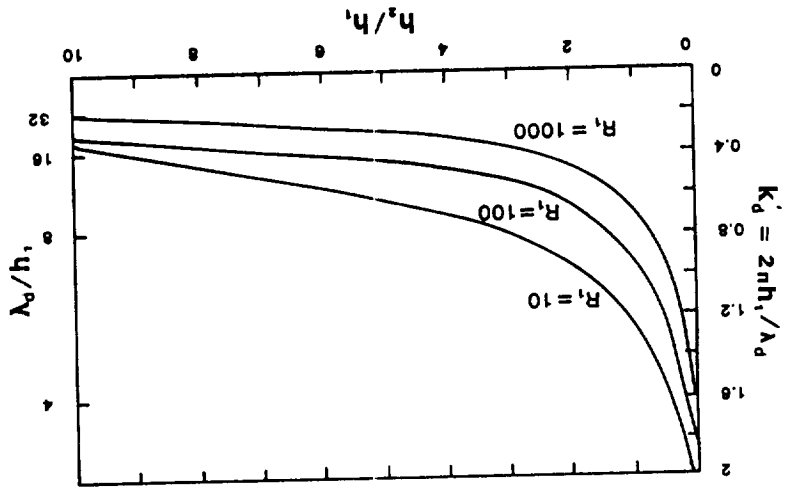


Figure 1. Dimensionless dominant wave number ( $k_d$ ) and wavelength ( $\lambda_d/h_1$ ) as a function of the thickness of a subsurface layer of thickness  $h_2$  for a model lithosphere with a competent subsurface layer over an incompetent subsurface layer and a competent substrate. Results are shown for three ratios of the strengths of the surface to subsurface layers ( $R_1 = \tau_1/\tau_2$ ).

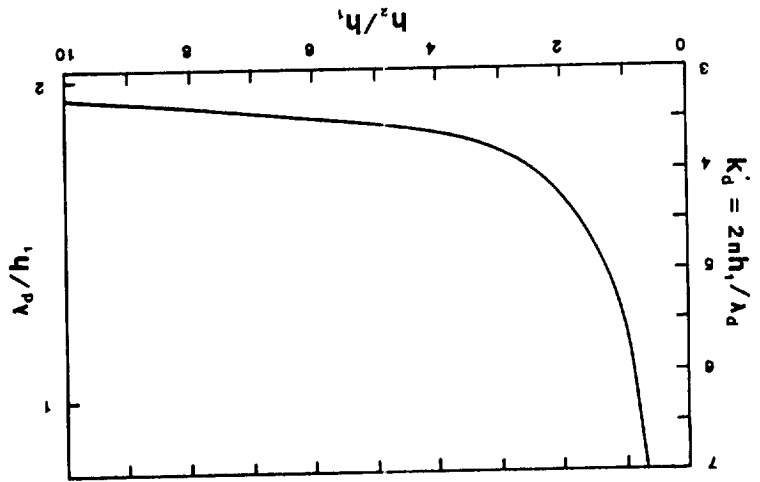


Figure 2. Dimensionless dominant wave number ( $k_d$ ) and wavelength ( $\lambda_d/h_1$ ) as a function of a subsurface layer of thickness  $h_2$  for a model lithosphere that contains a strong perfectly plastic layer over a weak viscous layer and a rigid base. All parameters are normalized by the thickness of the surface layer  $h_1$ .



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