ORIGIN OF MAGNETIC LINEATIONS ON MARS

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1. SUMMARY

The magnetic lineations discovered by MGS have been considered to be evidence of early plate tectonics on Mars. However, the lineations approximately follow lines of latitude, i.e., small circles. This presents significant geometrical problems for plate-like spreading, particularly at high latitudes. However, the sublatitudinal orientation of the lineations is consistent with meridional extension and perhaps limited crustal spreading due to a stress event centered near the geographic pole. We hypothesize that this event was the early formation of the crustal dichotomy through mantle-convective processes. This could have taken the form of a southern megaplume that formed the thick highlands crust or as subduction or downwelling in the north. Both would result in tensional stresses in the south that would form extensional fractures perpendicular to the CM-CF offset. The observed magnitude and distribution of magnetization indicates that crustal intrusion associated with this major mantle-convective event resulted in ~1000 km of extension in the southern highlands. Subsequent spin-axis reorientation due to loss of crust in the north or gain of crust in the south brought the CM-CF offset into its present N-S alignment. A portion of the ancient valley networks observed in the southern highlands are spatially associated with crustal magnetism and are quantitatively shown to be consistent with hydrothermal discharge over crustal intrusions.
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3. INTRODUCTION

The remarkable magnetic anomalies discovered by the Mars Global Surveyor (Acuña et al., 1999; Figure 1) strongly suggest that early Mars had an intrinsic magnetic field and that newly created crust was magnetized by thermoremanence. The large magnitude and size of the anomalies further indicates that the tectonic processes that influenced their formation were global in scale. Connerney et al. (1999) have suggested that this process was plate tectonics, a hypothesis that has received widespread attention. They based this hypothesis qualitatively on the apparent linear nature of the magnetic anomalies, which was considered to be similar to terrestrial magnetic "stripes" formed by seafloor spreading. Connerney and co-workers further showed through simple modeling that the intensity of magnetization had to be relatively strong compared to terrestrial analogs in order to reproduce the large magnetic fields observed from orbit.

![Figure 1](image.png)  
*Figure 1.* Magnetic field measured by MGS, from Connerney et al. (1999). Note strong sublatitudinal orientation to the lineations. Solid lines and dots denote segments used for geometric analysis (see below).
Plate tectonics has been such a successful theory on Earth because, like all good theories, it has strong predictive power. These predictions follow from the simplicity of rigid-plate motion on a sphere, which constrains the geometry of plate interactions. In general, spreading centers are not constrained to follow any particular orientation, although a great-circle path is required for normal spreading (when movement is perpendicular to the segment orientation). A second basic observation that can be made in addition to their linearity is that the anomalies approximately follow lines of latitude (Figure 1.), i.e., small circles. Especially at high latitudes, this presents strong geometrical problems for plate spreading, and suggests that other, alternative hypotheses should be considered.

Any such alternative process must be able to produce dominantly meridional extensional stresses over at least the latitudinal band in which the magnetic anomalies are observed, large strain (presumably tens to hundreds of percent or more; see below), and be limited to the Noachian era (as the lineations are observed only in the heavily cratered southern hemisphere). Potential candidates are global expansion, true polar wander, despining, and formation of the "hemispheric dichotomy." While expansion produces extensional stress, such stress is not limited to a meridional direction; furthermore the strain for plausible expansion is very small (see Solomon, 1979). Polar wander can be rejected for similar reasons (see Melosh, 1980; Grimm and Solomon, 1986). Despinning does produce a zone of east-west normal faulting at latitudes higher than 36° (Pechmann and Melosh, 1979), but the martian anomalies exceed this range, and plausible changes in flattening still produce total hemispheric extension limited to tens of kilometers.

Mars is divided by a "hemispheric dichotomy" into heavily cratered southern highlands and resurfaced northern lowlands. Leading hypotheses for this dichotomy encompass both internal (mantle convection: Lingenfelter and Schubert, 1973; Wise et al., 1979) and external (impact: Wilhelms and Squyres, 1984) mechanisms. However, MGS measurements of martian topography (Smith et al., 1999) reveal that much of the difference in elevation between the south and north is due to the center-of-mass/center-of-figure (CM-CF) offset. This degree-one offset corresponds to a steady downward slope of 0.036° between the south and north poles, and strongly favors a "smooth" internal process for the formation of the hemispheric dichotomy. We suggest that the north-south CM-CF offset is not coincidence, but the result of spin-axis reorientation due to mantle-convective formation of the hemispheric dichotomy, and that the orientation of magnetic anomalies along lines of latitude is a result of meridional extension of hundreds of kilometers or more associated with this major geological event in the early history of Mars.

The objectives of this project were to (1) Quantify the planform and intensity of crustal magnetization, (2) Assess how these geometrical patterns constrain models for the formation of the magnetization, particularly the origin of the crustal dichotomy, (3) Explore the relationships, if any, of crustal magnetization to surface geology, in particular, the valley networks.
4. MAGNETIC MODELING

Inversion of the pattern of crustal magnetization from magnetic anomalies is the foundation for determining the time scale for geomagnetic reversals and measurement of seafloor spreading (Vine and Matthews, 1963). The crustal magnetism is preferred to the magnetic anomalies themselves because superposition of the anomalies of adjacent blocks with differing magnetization intensities and/or directions results in a complex signal that may not be visually interpretable; furthermore induced moments distort the observed field and the inferred positions of reversals. Although the effects of induced magnetism may be ruled out for Mars (its crustal magnetism must be wholly remanent in the absence of a present-day intrinsic field), reduction of the observed data to the actual magnetization is an important step in understanding the origin of the magnetic anomalies.

Connerney et al. (1999) presented simple models for several individual MGS orbits. Their model is based on the following assumptions: (1) magnetization is two-dimensional, (2) magnetization is constant in crustal blocks 200 km wide by 30 km thick, and (3) the two components of magnetization $J_x$ and $J_z$ are allowed to vary independently. These models reproduce the observed anomalies quite accurately (Figure 2). As acknowledged by these authors, however, this is only one possibility for the actual crustal magnetization.

We independently modeled two of the orbits presented by Connerney et al. (1999). The data were digitized from the published paper. We also used a generalized inverse using singular-value decomposition to solve for the crustal magnetization. Our results for orbit 1999 Day 15 P6, also using an unconstrained $J_x$, $J_z$ model, are very similar to those of Connerney and co-workers (Figure 3). This result was obtained for a relatively low damping factor (SV cutoff ratio) of $10^{-2}$. A similar comparison was obtained for orbit 1999 Day 20 P7.

We performed a more thorough exploration of the parameter space for this model than reported by Connerney et al. (1999). We are examined the effects of varying block thickness, block width, and the damping factor on the goodness-of-fit, the model roughness, and the maximum magnetization. We found, as expected, that goodness-of-fit and model roughness are inversely related through the damping parameter. Because Connerney and co-workers emphasized the best fits to the data, their models are relatively rough. Connerney et al. (1991) state that "reversals in the direction of magnetization of adjacent blocks are common, if not the rule." However, strong variations between adjacent model blocks is almost always a signature of excessively high model variance or roughness. Reversals would be more believable if groups of at least several model blocks had the same sign, which can be achieved by using a narrower...
block and a higher damping. We find that goodness-of-fit is insensitive to block thickness but does improve with narrower blocks (because there are more free parameters) but that, conversely, model roughness does not vary as strongly with block width as it does with block thickness. The maximum crustal magnetizations, already apparently extreme for the nominal model of Connerney and co-workers, increase as the blocks are thinned.

While the models of Connerney et al. (1991) were useful as exploratory tools to describe the general magnitude of crustal magnetization, the spatial patterns of this magnetization are random and unrealistic (Figure 3). This is because $J_x$ and $J_z$ are both allowed to vary independently. In reality, the magnetic inclination will vary with absolute plate motion and true polar wander, but is often considered constant for some number of geomagnetic reversal cycles of seafloor spreading. Alternatively, if there is no plate motion and polar wander, the magnetizing field should closely approximate a global dipole. We have tested this latter hypothesis by constraining $J_x/J_z = 0.5 \tan(\theta)$, where $\theta$ is the geomagnetic colatitude. The magnetization amplitude (positive or negative) for each block remains as a free parameter. The solved magnetizations now vary smoothly (Figure 5) and, because the model is more tightly constrained, the required intensity of magnetization is less than half that of the free $J_x,J_z$ model. The cost is a modest decrease in goodness-of-fit (Figure 6) which, while statistically significant, does not detract from the plausibility of the model.

| Table 1. Summary of Magnetic-Modeling Fits to MGS Magnetometer Data. |
|-----------------|-----------------|----------------|
| Fit (%) | $J_{\text{max}}$ (A/m) | Pole         |
| 1999/15/P6      |                 |               |
| Arbitrary        | 98              | 41            | -             |
| Dipole +/-       | 95              | 20            | 18 N          |
| Dipole + only    | 91              | 55            | 8 S           |
| Const. Incl.     | 95              | 27            | 19 N          |
| 1999/20/P7      |                 |               |
| Arbitrary        | 96              | 34            | -             |
| Dipole +/-       | 93              | 41            | 57 S          |
| Dipole + only    | 90              | 61            | 47 S          |
| Const. Incl.     | 93              | 35            | 54 S          |
Figure 3. Best-fitting crustal magnetization for orbit 1999 Day 15 P6. The model formulation closely follows that of Connerney et al. (1999). The magnetization directions of each block are unconstrained, i.e., the x- and z- magnetizations vary independently. Note close match to the result of Connerney et al. (1999) (Fig. 2 above). However, when magnetizations are plotted as vectors, a random, unrealistic magnetizing field is inferred.

Figure 4. Data (symbols) and 2D model fit (lines) to magnetic field for model in previous figure. Goodness-of-fit is 98.4%.
Figure 5. Best-fitting crustal magnetization for same data as above, except magnetization directions are constrained to follow a global dipole. The latitude of the magnetic pole is a new free parameter, but each of the model blocks now has only one free parameter (the total magnetization) instead of two. The maximum magnetization in the constrained model is about half that of the unconstrained model (20 vs 42 A/m). The apparent magnetic equator is near x=-1000 km (where the magnetizations are nearly horizontal) and corresponds to a magnetic pole near 54°N, 16°W.

Figure 6. Data (symbols) and 2D model fit (lines) to magnetic field for model in which magnetization direction is constrained to follow a global dipole. Goodness-of-fit is 94.5%.
While the 2D models have lead to useful understanding of the crustal magnetization of Mars (see below), continuity between orbit tracks cannot be enforced, whereas the linear nature of the magnetic anomalies suggests that such continuity exists. In order to understand fully the spatial relationships of magnetization, 3D models are necessary. This was an important goal of our original proposal that was not attained. The primary responsibility lies with the PI, largely because constant supervision was not possible owing to the distance separating the PI’s workplace and the graduate student at the performing institution. Secondary issues were the lack of an available student during the first year of this project and the subsequent ability and experience of that student in performing geophysical inversions. To reduce commitments to a level that could be managed at the PI’s ability to interact with the performing institution, the program there was descoped to a single graduate student. This senior student carried out the work described in the attached reprint to this report. During the period of this contract, other groups have produced planform-magnetization models (Prurucker et al., 2000; Arkani-Hamed, 2002, Cain et al., 2002)

5. STRAIN ASSOCIATED WITH MAGNETIC ANOMALIES

The magnetization models can be used as the basis for estimating the amount of extension under meridional stress. We assume that such extension and intrusion occurred while the martian geodynamo was active, but that the intrinsic field was zero at other times. We next assume that the maximum observed magnetization intensity corresponds to a model block that is entirely composed of intruded, magnetized, igneous rocks. This is an upper limit, as even the most magnetized block could still contain unmagnetized material. It is also implicitly assumed that the scale of mixing of magnetized and unmagnetized rock is small compared to the block width. Under these assumptions, the extension of each block is proportional to the block magnetization, and the total extension over the profile can be computed. For the result shown in Figure 5, the extension is 1550 km, or a strain of 59% to produce the present-day profile length of 4000 km. While this amount of strain is large compared to that expected from vertical tectonics alone, it is only moderately larger than that inferred for deep-seated extension of the Valles Marineris (10-30%; Anderson and Grimm, 1998) and yet is small compared to the essentially unbounded extension that occurs for plate tectonics. The minimum extension estimated from the models above is 830 km, or 26% strain.

Figure 8. Schematic illustration of assumed relationship between magnetization intensity and crustal strain.
6. GEOMETRICAL TESTS

The paleomagnetic record of Mars can be used in several ways to test the predictions of the plate-tectonic and hemispheric-dichotomy hypotheses. The predictive power of the theory of plate tectonics lies in the simplicity of its geometric rules for rigid motions on the surface of a sphere (McKenzie and Parker, 1967; Morgan, 1968), and this theory can be applied to Mars. All plate movements occur along small circles about some apparent rotation pole, and where a plate boundary lies in the direction of motion a transform fault results. Therefore the orientation of transform faults is the most robust method of determining the geometry of relative rotation between two plates. Grimm and Solomon (1989) applied this method to putative fracture zones on Venus. If the crustal magnetization reconstructions described above are sufficiently robust and strike discontinuities are observed, then we will test to see if they satisfy the geometrical requirements of transform faults.

Ridges and trenches are not geometrically constrained. Where convergent or divergent motions are perpendicular to a plate boundary, however, that boundary will lie along a great circle. Regardless of spreading direction, the spreading rate must be proportional to \( \sin(\Delta) \), where \( \Delta \) is the angular distance to the rotation pole. This is the second-best method of determining plate geometry. The third method, earthquake focal mechanisms, is less reliable and is neither available for Mars nor appropriate to paleomotion studies.

![Figure 7. Map of RMS deviations (degrees) for trial southern poles about which magnetic lineations (connected diamonds) are tested as small circles. Best fit is 71 S, 8 E (asterisk); proximity to the geographic pole simply confirms the sublatitudinal orientation of the lineations.](image-url)
Working directly with the modeled planform of magnetization, we can examine along-strike variations in the widths of anomalies which could be a manifestation of a $\sin(\Delta)$ spreading-rate variation. We can also test whether segments lie along great circles, which would indicate normal spreading. Clearly smaller segments can be arbitrarily well fit to a great circle, whereas transform faults and/or oblique spreading could render the test invalid for larger segments. As the former dominate apparent ridge directions on Earth, arcs that do not fit great circles should show transform offsets if plate tectonics is still satisfied (e.g., Southeast-Indian and Pacific-Antarctic Ridges on Earth are grossly east-west, but are strongly segmented).

While there are a number of tests for plate tectonics, the global dichotomy model as presently formulated requires only that the crustal magnetizations all follow small circles. For such a test (Figure 1), we find an apparent pole at 71 S, 352 W (71 N, 172 W), with an RMS arc deviation of 1.2 degrees (Figure 7). The longitude is not particularly meaningful, as it is determined by the distribution of the data, but the latitude and goodness-of-fit confirms the visual impression that the lineations follow small circles, apparently around some former source beneath the Vastitas Borealis.

The bilateral symmetry of magnetization that was the hallmark of seafloor spreading is not immediately obvious in the martian lineations, although the subtle arc shapes to the positive anomaly bands 35 S (concave to the south) and 55 S (concave to the north) suggests symmetrical spreading about a rotation pole located close by. We have tested the bilateral symmetry of magnetization using established statistical techniques (Grimm and Solomon, 1989). Typical correlations $-0.65$ (Figure 8) do not strongly support the concept of bilateral, plate-like intrusion and spreading. Rather, the magnetic anomalies were probably formed by serial, perhaps spatially random magmatic intrusions.

![Figure 8. Bilateral symmetry test for plate tectonics. Each modeled magnetic block is assumed to be a spreading center and the correlation coefficient calculated around it to the maximum possible distance.](image-url)
7. SYNTHESIS: GLOBAL MODELS
FOR THE ORIGIN OF MAGNETIC LINEATIONS

We review all alternative hypotheses for the origin of magnetic lineations on Mars and, by process of elimination, assess those that remain most likely.

Plate Tectonics. This is the original hypothesis of Connerney et al. (1999). Arbitrarily large (unconstrained) crustal extension and intrusion is allowed. An unmapped location for subduction is required to conserve area. There are strong geometrical constraints to motion for rigid lithosphere, including the fit of lineation arcs to great circles and bilateral symmetry of lineations. Neither of these constraints is satisfied for the martian magnetization. Finally, a near-constant crustal thickness should be produced by decompression melting, but gravity data indicates a strong gradient in crustal thickness in the southern highlands (Zuber et al., 2000). We conclude that earth-like plate-tectonics did not form the magnetic lineations.

Tharsis Tectonics. The load from Tharsis is known to have deformed the crust to great distances; could it also have led to the magnitude and direction of the magnetic lineations? Two prominent lineations can indeed be fit to great circles with centers near Tharsis, but the anomaly pattern as a whole does not follow Tharsis radials. Furthermore, strains from such loading, or “vertical tectonics,” are expected to be much smaller, in the range of a few percent; recall that the strain associated with the lineations is of order ten percent or more.

Spin Tectonics. This includes faulting due to the shift of the rotational bulge during spin-axis reorientation or polar wander and the relaxation of the rotational bulge due to tidal despinning. For the former, we expect N-S oriented extensional structures near the paleopole and only moderate strains (several %, e.g., Grimm and Solomon, 1986). For the latter, E-W oriented fractures can occur, but they are restricted to poleward of 36° from the equator (Pechmann and Melosh, 1979). The magnetic lineations of Mars fail the geometrical tests for both of these models.

Northern Crustal Depletion. Subduction or other crustal recycling is one hypothesis for formation of the global dichotomy. In this scenario, “back-arc” tensional stresses might be set up in the south, which would cause strong gradients in extension, intrusion, magnetization. The expected strain magnitude unknown, but it would likely be bigger than vertical or spin tectonics but less than plate tectonics: this is consistent with moderately large strains inferred for the magnetic lineations.

Southern Crustal Accretion. This is an alternative hypothesis for formation of the global dichotomy in which the a megaplume results in generation of a massive crustal plateau. This is consistent with the basaltic composition of the southern crust (Christiansen et al., 1999) and indeed is facilitated by a 2.6x gravity-scaled efficiency for decompression melting (McKenzie, 1999). Melt thicknesses of several tens of km result from several tens of km extension (see Grimm and Hess, 1997) at hotter Noachian mantle temperatures (Schubert et al., 1992). A southern megaplume could generate large, latitudinally varying crustal thickness (80 km at S. pole to 35 km at N. plains boundary; Zuber et al., 2000). Radial extension over the uplift causes
circumferential extension fractures (*Banerdt et al*, 1982); subsequent melt intrusion and solidification would form the magnetic lineations. This model is most consistent with all available data.

In either the northern-depletion or southern-accretion models, the center of mass (CM) becomes displaced from the center of figure (CF), resulting in a spin imbalance. Polar wander then brings the CM-CF offset into alignment with the spin axis, i.e., to a N-S configuration. The near-perfect N-S alignment of the CM-CF offset has not heretofore been explained.

Figure 9 (on following page) illustrates the northern-depletion and southern-accretion models with subsequent spin-axis realignment.

8. RELATION OF CRUSTAL MAGNETISM TO SURFACE GEOLOGY

The spatial distribution of crustal magnetization is statistically correlated to the locations of Noachian, southern-highlands valley networks. We hypothesize that at least some of these valley networks were carved by hydrothermal discharge over the cooling intrusions represented by the magnetization. To test this hypothesis, a numerical model for hydrothermal discharge on Mars was formulated and calibrated; a by-product is a better understanding of the physical controls on hydrothermal discharge as well as quantitative constraints on crustal permeability imposed by geochemistry. We found that the total discharge due to intrusions building that part of the southern highlands crust associated with valley networks is comparable to the discharge inferred from valley geometry, supporting the hypothesis. Details are given in the attached reprint.
Northern Crustal Depletion (Downwelling) model for formation of magnetic lineations through "back-arc" extension. Crust: green, mantle: red, lineations (in cross-section), red line; CM, center of mass; CF, center of figure.

Southern Crustal Accretion (upwelling) model for magnetic lineations through extension above megaplume.

Spin-axis reorientation due to loss of crust in north or gain of crust in south brings CM-CF offset into N-S alignment, also brings magnetic lineations into sublatitudinal orientation.
5. REFERENCES


Banerdt et al. (1982) JGR, 87, 9723.


Christiansen et al. (1999) AGU Fall Mtg., P12B-05.


Zuber et al. (2000) *Science*, 287, 1788
Controls on Martian hydrothermal systems: Application to valley network and magnetic anomaly formation

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1. Introduction

Hydrothermal circulation is an important part of many terrestrial igneous, metamorphic, and sedimentary environments and has profound geochemical and biological implications. On the Earth it accelerates the cooling of magmatic bodies in systems ranging from divergent plate boundaries to individual volcanoes and frequently produces discharge in the form of hot springs, geysers, and submarine vents. There is also evidence for the past existence of hydrothermal systems on Mars (see Farmer [1996] for a review). Valley networks are associated with structures that are able to force groundwater to the surface [Baker et al., 1992]; these include some of the younger volcanoes [Galick, 1998], ancient volcanoes, rifts, and impact craters [Tanaka et al., 1998]. Many more observations have led to groundwater discharge as the favored method of runoff production (see Baker et al. [1992], Baker [2001], and Carr [1996] for reviews) [Malin and Carr, 1999; Malin and Edgett, 2000]. Hydrothermal circulation in a magma chamber. The principal application of our model is to test the viability of hydrothermal circulation as the primary process responsible for the broad spatial correlation of Martian valley networks with magnetic anomalies. For host rock permeabilities as low as $10^{-17} \text{ m}^2$ and intrusion volumes as low as 50 km$^3$, the total discharge due to intrusions building that part of the southern highlands crust associated with magnetic anomalies spans a comparable range as the inferred discharge from the overlying valley networks. INDEX TERMS: 1832 Hydrology: Groundwater transport; 1860 Hydrology: Runoff and streamflow; 1545 Geomagnetism and Paleomagnetism: Spatial variations (all harmonics and anomalies); 5440 Planetology: Solid Surface Planets: Magnetic fields and magnetism; 5114 Physical Properties of Rocks: Permeability and porosity; KEYWORDS: Hydrothermal, groundwater, runoff, crustal magnetism, intrusions

2. Model

Two-dimensional, axisymmetric representations of hydrothermal circulation in a magma chamber and its host rock were modeled with the U.S. Geological Survey (USGS) code HYDROTHERM [Hayba and Ingebritsen, 1994, 1997] and its graphical user interface HPost (P. S. Hsich, USGS, unpublished material, 2000). HYDROTHERM can model temperatures from about 0$^\circ$C to 1200$^\circ$C and pressures from 0.5 to 10,000 bars and keeps track of both liquid...
and gaseous phases of pure water. It solves mass, momentum, and energy balance equations expressed in terms of dependent variables
pressure and enthalpy. The choice of enthalpy over temperature allows the thermodynamic state of the fluid to be specified uniquely
under two-phase conditions. Viscosity and density for a particular
temperature and pressure are obtained from a look-up table. [5]

The momentum balance equation is Darcy’s law, which for a single fluid phase $i$ is

$$v_i = -\frac{k_i}{\mu_i} (\nabla p + \rho_i g \nabla z),$$

where $v_i$ is the Darcy velocity, $k_i$ is intrinsic permeability, $\mu_i$ is
dynamic viscosity, $p_i$ is pressure, $\rho_i$ is density, $g$ is gravitational
acceleration, and $z$ is depth. The relative permeability $k_{rel}$ quantifies
the reduction of the flow of phase $i$ due to the presence of the other
phase. The mass balance (continuity) equation for phase $i$ is

$$\frac{\partial}{\partial t} (n \phi_i) + \nabla \cdot (\rho_i v_i) = 0,$$

where $n$ is porosity, $t$ is time, and $\phi_i$ is volumetric saturation ($S_{\text{water}} + S_{\text{steam}} = 1$, i.e., the medium is fully saturated). The continuity
equation for the entire system is the sum of the equations for each
phase. The energy-balance equation for the entire system is

$$\frac{\partial}{\partial t} \left[ (1 - n) h_i + n \sum S_i h_i \right] + \nabla \cdot \left[ \sum k_{rel} h_i v_i \right] - K_m \nabla T = 0,$$

where $h$ is enthalpy, $T$ is temperature, $K_m$ is medium thermal
conductivity, and subscript $r$ refers to rock matrix properties.

HYDROTHERM solves the equations by performing Newton-
Raphson iterations on an equivalent finite difference system (with
the horizontal dimension expressed in radial coordinates) until mass
and energy residuals fall below specified maximum values.

[6] The horizontal extent of the host rock is 20 times that of the
magma chamber and the vertical extent is 20 km, both of which are
sufficient to accommodate flow from all magma chambers studied.
The right vertical boundary is not intended to represent the limit of
a horizontally bounded water source, but to provide enough space
for a realistic response to a local temperature perturbation. As
much fluid flows through this boundary as is necessary to balance
the net flow through the top horizontal boundary. The suitability of
the chosen horizontal extent was tested by a baseline model
measuring 40 chamber radii across, which yielded almost the
identical discharge to the original model. Fluid is allowed to cross
the upper horizontal and right vertical boundaries, while tempera-
ture and pressure are fixed. Recharge from the surface is typically
<10% of discharge, indicating that the strength of discharge does
not depend on infiltration of runoff. Adaptive boundary conditions
to prevent infiltration for a dry Mars could therefore be neglected.
Sufficient recharge or discharge occurs through the right vertical
boundary to conserve the mass of the system.

[7] The initial pressure distribution is hydrostatic, with a surface
value of 1 bar. A small geothermal gradient (0.5°C/km), applied to
ensure stable decay of the surface discharge, does not otherwise
affect the model (the effects of substantial geothermal gradients
were explored, and the results are described below). The lowest
temperature HYDROTHERM can handle in this model is 10°C,
and this was used for the surface boundary condition. Both the
surface temperature and pressure were adopted for numerical
convenience, and they neither affect the results nor are intended
to represent Earth-like climatic conditions. The left vertical and
lower horizontal boundaries are no-flow, and the temperature and
pressure are free to vary. Grid spacing is ~100 m in both directions
near the magma chamber and at greater horizontal distances
increases logarithmically. [8] Present Martian conditions may include a permafrost layer
which must first be melted before discharge is produced. Guick
[1998] estimated that the time required to melt a 2 km thick layer
of permafrost was much less than the lifetime of the hydrothermal
system, making it unlikely that quantities integrated over the
lifetime of the system, such as total discharge, should be signifi-
cantly altered. During the late Noachian, when the hydrothermal
systems proposed here were active, the permafrost layer was likely
to be much thinner than 2 km, making melting times even shorter.
A HYDROTHERM simulation that quantifies the role of ice in our
models is described below.

[9] Our baseline model consists of a 50 km³ chamber emplaced
at a depth of 2 km below the surface into host rock of permeability
10⁻¹⁶ m². The dimensions of the chamber are expressed in terms of
its aspect ratio ($D/H$), defined here as the ratio of diameter to
height. The baseline chamber has $D/H = 2$. The volume of the
chamber is taken after that modeled by Hayba and Ingebritsen
[1997] and is also of similar size to magma chambers found under
mid-oceanic ridges [Burnett et al., 1989]. To test if our baseline
model is intended only as an example model from which we later
develop and does not necessarily represent any sort of
"ideal" parameter values.

[10] The chamber is emplaced instantaneously at a temperature
of 900°C. Because this approach ignores discharge produced during
supersolidus cooling and the finite intrusion process, it produces
relatively conservative results. The magma chamber is initially
impermeable, but as it cools through a brittle-ductile transition
(BDT) between 400° and 360°C, it is assumed to fracture and
become as permeable as the surrounding host rock [Hayba
and Ingebritsen, 1997]. A semilog form is adopted for this transition,
wherein the log of the permeability scales linearly with temperature.
Note that the "permeability" BDT may differ somewhat from the
classic "deformational" BDT [e.g., Kohlschedel et al., 1995]. All
models use a rock density of 2500 kg m⁻³, a thermal conductivity'
of 2.0 W m⁻¹ K⁻¹, and a porosity of 15%. Deviations from the
baseline model include host rock permeabilities of 10⁻¹³ and
10⁻¹⁵ m², porosity of 25%, magma chamber aspect ratio of 0.2
at depths of 8.5 and 15 km, and volumes of 100 and 2000 km³.

[11] Steam and water fluxes, while fundamental to discharge
calculations, may also be used to test hypotheses regarding the
possible geochronological alteration in the system. For each time step, a
measure of water-to-rock ratio for reactions above a particular
threshold may be calculated by measuring the total
mass of water that passes through the region warmer than the
threshold value and dividing by the volume of the region.

3. Results

[12] In the baseline model (magma chamber depth is 2 km,
volume is 50 km³, $D/H = 2$, and host rock permeability is 10⁻¹⁶ m²;
Figure 1) an initial peak in the surface discharge occurring at only a
few hundred years following magma emplacement is due to thermal
pressurization [Delaney, 1982]. Its peak value is not significantly
greater than discharges that occur later in the model. The extremely
short-lived nature of this effect, which is less extreme for higher
host rock permeabilities, contributes negligibly to the total, time-
integrated mass of water produced by the system (total discharge),
which we assume to be of primary importance in valley erosion.
Additionally, thermal pressurization may be weaker in the more
realistic case of a finite duration intrusion process.

[13] The broad peak at 45 kyr is due to thermal convection
of groundwater. At first, the chamber is supercritical and imperme-
able, allowing convection in the surrounding host rock only.
Surface discharge peaks when the magma chamber has cooled
sufficiently to become permeable and admit advection. When only
3.1. Permeability

For simplicity, permeability was homogeneous in all models. Gulick [1998] used a value of $10^{-11}$ m$^2$ for hydrothermal systems on Mars inferred from young, near-surface Hawaiian volcanics. Ingebritsen and Sanford [1998], however, report that permeabilities in the east rift zone at Kilauea, while high near the surface ($10^{-10} - 10^{-11}$ m$^2$), are much lower at depths of just 1–2 km ($10^{-16} - 10^{-15}$ m$^2$) in rock of the same composition. Manning and Ingebritsen [1999] estimate values of between $10^{-17}$ and $10^{-14}$ m$^2$ for the mean continental permeability between 1 and 10 km depth. We modeled permeabilities from $10^{-17}$ to $10^{-15}$ m$^2$ (Figure 2). The upper limit is for computational convenience; we demonstrate below that results for higher permeabilities can be extrapolated from this range.

Discharge from the low-permeability ($10^{-17}$ m$^2$) model has a characteristically large thermal pressurization peak followed by a weak main peak. Conduction is the dominant mode of heat transfer in this model, and the spatially integrated surface discharge at any given time (henceforth called “instantaneous discharge”) is only about a tenth that of the baseline model. In the baseline model, convection and conduction play comparative roles in heat transport, while in the high-permeability ($10^{-15}$ m$^2$) model, convection is dominant. Discharge in all three models lasts between 2500 and 3000 kyr, but rate of decay of discharge is, in general, proportional to permeability, with the $10^{-17}$ m$^2$ model decaying to a tenth of its peak value in 200 kyr, the baseline model in 175 kyr, and the $10^{-15}$ m$^2$ model in 135 years.

An important feature of the $k = 10^{-15}$ m$^2$ model is the presence of a steam-dominated zone above the magma chamber during the first several thousand years. This phenomenon may have implications for chemical alteration, as described in the discussion below.

The relationship of discharge to permeability was investigated using the high-permeability results of Gulick [1998] as a starting point. She modeled a $10^{-11}$ m$^2$ hydrothermal system which included the magma chamber implicitly through a heat flux boundary condition based on the analytical solution to the conductive heat flow through the wall of an infinitely long cylinder. This implies that no heat was lost through the roof and floor of the

Figure 1. Surficial discharge from the baseline model (magma chamber depth 2 km, volume 50 km$^3$, aspect ratio 2, host rock permeability $10^{-13}$ m$^2$). The regions separated by solid vertical lines in the enlarged portion of the figure denote the different phases present within in the model. The magma chamber becomes permeable over the interval 0.7–57 kyr. The noise on the curve is due mainly to discretization effects in the presence of steam, which has much greater velocity than the liquid phase.

Figure 2. Surficial discharge from a 2 km deep magma chamber of volume 50 km$^3$ for host rock permeabilities of $10^{-17}$, $10^{-16}$, and $10^{-15}$ m$^2$. The total discharge of the $k = 10^{-16}$ m$^2$ model, in kilograms, appears above its curve. The numbers above the other curves denote the fraction of this discharge that their corresponding models produced.
more efficient numerically and allows us to extend the range of
chamber and that no BDT was encountered on cooling. By
this figure, with an assumed water density of 1000 kg m$^{-3}$.
by Gulick was converted from volume to mass for the purpose of
three full geometry models (from Figure 2). The value obtained
discharge obtained by the single permeability modeled by Gulick
these data, with its large permeability asymptote, is shown. The
geometry similar to those of

discharge and permeability could be represented approximately by

\[ D = a - be^{-CK}, \]

where $D$ and $K$ are the base 10 logarithms of total discharge and
permeability, respectively, and $a$, $b$, and $c$ are positive constants
(with best fit values of 38, 0.089, and 0.12, respectively). As $K \to \infty$,
$D \to a$, meaning that as permeability tends to very high values,
discharge increases asymptotically toward a finite maximum value.
This is not surprising, since the forces driving the flow are limited
by the amount of heat contained in the magma chamber. For the
range of permeabilities covered with our own, more complete
geometry ($K < -15$), $D$ is approximately linear in $K$. Note that
the total discharges (whose relative magnitudes are given above each
curve in Figure 2) exhibit this linear behavior.

For our range of permeabilities, total discharge is approxi-
mately the same for both geometries. At higher magma chamber
aspect ratios, however, it is no longer reasonable to assume that
heat is lost through the chamber walls alone, and the total
discharges for the two geometries are expected to diverge. Instan-
taneous discharge is not the same for both geometries, even with
$D/H = 2$. Our geometry produces a greater maximum than the
simpler geometry ($\sim 1.75$ times as high for $K = 10^{-16}$ m$^2$) but drops
more rapidly thereafter.

Magma chamber cooling times, defined here as the average
time taken for chamber nodes to cool below a specified threshold
temperature, are of interest in these models. For all three perme-
abilities the cooling time for a threshold of 450°C (i.e., half the
emplacement temperature) is $\sim 30$ kyr. While an increase in host
rock permeability increases the velocity of the flow alongside the
still impermeable magma chamber, it does not significantly
enhance cooling [see Norton and Knight, 1977]. Once the chamber
becomes permeable, however, cooling progresses at different rates
in each model. The time taken for the chamber to cool to 250°C
was 67, 61, and 40 kys for permeabilities of $10^{-17}$, $10^{-16}$, and
$10^{-15}$ m$^2$, respectively.

3.2. Volume

Head and Wilson [1994] suggest that Martian magma
reservoir volumes could be as great as 2000 km$^3$ and that chamber
depths are most likely to range from 8 to 12 km. We ran models of
$D/H = 2$ magma chambers with volumes of 100 and 2000 km$^3$ at a
depth of 8.5 km and with host rock permeability of $10^{-16}$ m$^2$
(Figure 4). The 100 km$^3$ chamber produced $\sim 2.5$ times as much
discharge as the 50 km$^3$ chamber, while the jump from 100 to 2000
km$^3$ resulted in a factor of 73 increase. This approximately linear
relationship is reflected in the magnitude of the instantaneous
discharge, whose peak value has proportional increases. This,
coupled with an increase in cooling time with volume (and there-
fore an increase in the discharge lifetime), explains the observed
total discharge increase.

Although not obvious on the semilog plot of Figure 4, the
discharge produced by the 2000 km$^3$ chamber drops off at $\sim 800$
kyr, which is in agreement with values obtained by Cathles et al.
[1997] for a 2500 km$^3$ chamber with host rock permeabilities
ranging from $4 \times 10^{-17}$ to $10^{-18}$ m$^2$.

3.3. Depth

Models with 50 km$^3$ magma chambers at depths of 8.5 and
15 km and with host rock permeability of $10^{-16}$ m$^2$ were run
(Figure 5). Total discharge decreases by a factor of $\sim 1.5$ with each
6.5 km increase in depth. Similar depth dependence exists for the
same three chamber depths at host rock permeabilities of $10^{-17}$
and $10^{-15}$ m$^2$. A summary of the total discharge of all nine depth and
permeability combinations is depicted in Figure 6. The very small
discharge lifetimes for the 1000 km$^3$ model appears above its curve. The numbers above the
other curves denote the fraction of the discharge that their
 correspondent models produced.
corresponding models produced.

other curves denote the fraction of this discharge that their deep chamber appears above its curve. The numbers above the rock permeability of $10^{-16}$ m$^2$. The total discharge of the 2 km depths of 2, 8.5, and 15 km, all with aspect ratio 2, and with host figure 5. Surficial discharge from 50 km$^3$ magma chambers at 3.4. Aspect Ratio

[21] Magma chambers with $D/H = 0.2$, 2, and 20 were modeled (figure 7). These correspond to chamber radii of 1.2, 2.5, and 5.4 km and chamber heights of 12, 2.5, and 0.54 km. Chamber depth in all cases is 2 km, and chamber volume is 50 km$^3$. The horizontal extent was fixed at 50 km for all three models, rather than scaled with the chamber radius, so that the influence of the right vertical boundary was the same in each model. The model discharge of these models is controlled by the cooling time of the magma chamber, and therefore its surface area-volume ratio $A/V$ (1.89, 1.59, and 4.06 km$^{-1}$ for the three aspect ratios, respectively). There is an approximately linear inverse relationship between $A/V$ and total discharge; that is, high $A/V$ implies low total discharge and vice versa. The effects of aspect ratio are observed more directly when a substantial geothermal gradient is applied (see section 3.5).

Figure 6. Total discharge from 50 km$^3$ magma chambers of aspect ratio 2 at depths of 2, 8.5, and 15 km and host rock permeabilities of $10^{-17}$, $10^{-16}$, and $10^{-15}$ m$^2$.

travel farther in order to expel heat from the system and is therefore less efficient at cooling the magma chamber.

Figure 7. Surficial discharge from 50 km$^3$ magma chambers with $D/H = 0.2$, 2, and 20, all at a depth of 2 km, and with host rock permeability of $10^{-16}$ m$^2$. The total discharge at $D/H = 2$ appears above its curve. The numbers above the other curves denote the fraction of this discharge that their corresponding models produced.

In all three models the flow pattern consists of a single, clockwise rotating convection cell alongside the chamber. For $D/H = 20$ this pattern does not give way to a series of cells above the chamber roof, as one might expect for flow between two infinite horizontal surfaces at different temperatures.

3.5. Geothermal Gradient

[25] Models with magma chamber aspect ratios of 0.2, 2, and 20 were run with an initial host rock geothermal gradient of 20°C/km, perhaps representative of early Mars [Schubert et al., 1992]. The presence of a geothermal gradient significantly affects the hydrothermal discharge. The flat, sill-like chamber ($D/H = 20$) produces the greatest total discharge because of its large horizontal exposure. The discharge dissipates more rapidly, however, because the chamber, having the greatest $A/V$ and being oriented perpendicular to the main flow direction, cools rapidly. Conversely, the tall, pipe-like chamber ($D/H = 0.2$), being immersed in warmer temperatures at depth, cools more gradually. Its geometry produces the largest convection cell and offers minimum obstruction to flow, resulting in the greatest total discharge despite its low peak value. Overall, tenfold variations in aspect ratio produce changes in discharge of less than a factor of 3.

[26] The effect of geothermal gradient is also observed in cooling times. The times taken for magma chambers to cool to half their emplacement temperature are, in order of increasing aspect ratio, 25, 45, and 4.0 kyr, respectively. In the absence of a significant geothermal gradient, the same chambers cool in 16, 28, and 3.5 kyr, respectively. An 8.5 km deep chamber in a geothermal gradient of 20°C/km cools to half its initial temperature in 52 kyr (as opposed to 30 kyr for the model with negligible geotherm). It also produces ~7 times as much discharge (with a peak 3.5 times higher) as the identical model with a negligibly small geothermal gradient.

[27] It should be noted that for permeabilities $>10^{-14}$ m$^2$, the Rayleigh number of a plane porous medium [e.g., Turcotte and Schubert, 1982] at 20°C/km indicates that weak convection may occur in the absence of a magma chamber. This phenomenon is observed in our high geothermal gradient models when no magma chamber is emplaced. Free convection in the terrestrial crust has been invoked by Raffensperger and Garven [1995] to explain the location of uranium ore deposits in sedimentary basins in Canada and Australia. Travis et al. [2001] showed that free convection may be capable of melting significant volumes of subsurface ice in the
Martian crust. However, all of these models, including our own, have homogeneous permeability in the convecting zone: realistic vertical heterogeneity in planetary crusts will both inhibit the development of large-scale crustal convection and decrease the efficiency of heat transfer. Although background geothermal gradients may affect cooling times of intrusions, we view their contribution to hydrothermal circulation as doubtful.

3.6. Ice

Gulick [1998] suggested that a 2 km thick subsurface permafrost layer above a 50 km³ magma chamber would melt in a few tens of thousands of years. This is of the same order as the time taken for discharge to peak in our own models, indicating that ice could significantly reduce total discharge. We thus ran the baseline model with a 1 km thick layer of subsurface ice (Figure 8), modeled after Bonacina et al. [1973], who approximated the melting process as a cooling period over a small finite temperature interval $\Delta T$. During melting the material is assigned an augmented specific heat given by

$$C_T = C_S + \frac{L}{\Delta T},$$

where subscripts $S$ and $L$ refer to the solid and liquid phases, respectively, and $L$ is the latent heat of fusion. We used $C_S = 1000 \text{ J kg}^{-1} \text{ K}^{-1}$, $L = 3.34 \times 10^7 \text{ J kg}^{-1}$, and $T = 5^\circ\text{C}$, giving $C_L = 6.78 \times 10^5 \text{ J kg}^{-1} \text{ K}^{-1}$.

The hydrothermal system took 52 kyr to melt a hole in the ice. The strong upward convection associated with increasing magma chamber permeability did not noticeably increasing melting rate. The total discharge produced by the model was $2.95 \times 10^{11} \text{ kg}$, about a quarter of that of the baseline model. This relatively severe reduction (which is expected to be worse for deeper chambers and lower host rock permeabilities) places an upper bound on permafrost thickness during valley formation on Mars. For valleys to form through sapping processes alone, the permafrost must be melted through, or aquifers carrying groundwater beneath the permafrost must intersect the surface. In either case, the permafrost can be no thicker than a few hundred meters.

3.7. Water Table

Surface discharge will occur only if convection is strong enough to elevate groundwater from the initial water table depth to the surface. We compared with hydrostatic conditions the vertically integrated product of density, gravitational acceleration, and depth in order to estimate the height a water table may attain through thermal-convective expansion. Integrations were performed over all vertical columns of finite difference blocks at all times and for a

Figure 9. Discharge as a function of water table depth, normalized to the zero depth water table value.
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Table 1. Water/Rock Ratios by Mass Calculated for 8.5 km Deep Chambers of the Indicated Volumes and Host Rock Permeabilities k

<table>
<thead>
<tr>
<th>Threshold Temperature, °C</th>
<th>k = 10^{-15} m^2</th>
<th>k = 10^{-16} m^2</th>
<th>k = 10^{-17} m^2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Volume = 50 km^3</td>
<td>Volume = 50 km^3</td>
<td>Volume = 2000 km^3</td>
<td></td>
</tr>
<tr>
<td>150</td>
<td>0.12</td>
<td>0.025</td>
<td>0.067</td>
</tr>
<tr>
<td>200</td>
<td>0.18</td>
<td>0.037</td>
<td>0.10</td>
</tr>
<tr>
<td>250</td>
<td>0.28</td>
<td>0.054</td>
<td>0.15</td>
</tr>
</tbody>
</table>

range of water table depths. The discharge was calculated from those parts of the water table that intersected the surface and was compared to zero depth water table models. The results (Figure 9) show that models with a water table 50 m deep produce only 10% of their zero depth water table equivalents, while models with a 100 m deep table produce only 2%.

4. Discussion
4.1. Implications for Hydrothermal Alteration

[31] Water/rock ratio (W/R) can be an important influence on the mineralogy of alteration products in hydrothermal systems. Martian hydrothermal alteration is thought to occur at low W/R (<10 by mass) [Griffith and Shock, 1997; Newsom et al., 1999], in which case its effect is small, and the initial mineral composition is the primary influence on the alteration assemblage [Griffith and Shock, 1997].

[32] HYDROTHERM mass flux results allow the W/R of a hydrothermal model to be calculated. At each time step, regions of the model above a specific threshold temperature are identified, and the total flux passing through them is calculated. On the basis of these data, a W/R (by mass) is calculated for each finite difference block in the model that is at some time above the threshold temperature. The average W/R values for the three reaction temperatures modeled by Griffith and Shock [1997], and for various model dimensions, are shown in Table 1.

[33] The large velocities in the $k = 10^{-15}$ m^2 model give the highest W/R for all threshold temperatures, ~5 times that of the $k = 10^{-16}$ m^2 model with a chamber of the same volume. Negligible geothermal gradient and low host rock permeability in our models are both factors that contribute to small W/R. Typical Martian geotherms (~50°C/km) will lead to larger W/R, so these estimates can be viewed as lower bounds. The presence of a significant geotherm may be expected to increase discharge by a factor of about 3 or 4 (as observed in the model with an 8.5 km deep chamber with geothermal gradient), and since W/R is directly proportional to the discharge flowing through the alteration area, it may experience a similar increase. This would still leave most values listed in Table 1 below unity.

[34] A primary goal of the work by Griffith and Shock [1997] was to calculate the amount of water bound to the rock during alteration. In a model based on the Shergotty SNC meteorite, ~8% of the final mineral assemblage (by weight) was water, implying that water/rock ratios >0.08 would be necessary to sustain hydrothermal circulation. Assuming the Shergotty composition is sufficiently generic to be applied to our own models, a W/R limit of 0.08 implies a minimum permeability of $10^{-16}$ m^2 for a 150°C reaction in host rock surrounding a 2000 km^3 chamber (Table 1).

[35] Newsom et al. [1999] suggest that the relative abundances of mobile elements such as sulfur, chlorine, sodium, and potassium may be explained by the presence of a mixture of neutral-chloride and acid-sulfate fluids during soil formation. Production of the latter fluid requires vapor transport [Rye et al., 1992], and Ingebritsen and Sorensen [1988] discuss situations in which vapor-dominated zones may occur. Their models require combinations of low-permeability barriers and in some instances topographic gradients to sustain vapor-dominated zones, which develop in the shallow subsurface only. These specialized structures have not been included in our generalized models; nonetheless, our highest permeability model ($10^{-15}$ m^2) with the shallowest chamber (2 km) does produce a short-lived (few thousand years) two-phase zone between magma chamber and surface. Steam develops here because of a combination of low pressures (which drop to a

Figure 10. Overlay of the absolute values of the 200 km vertical magnetic anomalies [Purucker et al., 2000] and the valley networks [Kieffer, 1981] (cylindrical projection). Valleys were not mapped for latitudes below 60°S. The anomalies range from 0 to 670 nT.
Table 2. Statistical Data for the Correlation Between Valley Networks and Magnetic Anomalies

<table>
<thead>
<tr>
<th>Magnetic Threshold, nT</th>
<th>p₀</th>
<th>p</th>
<th>(p - p₀)p</th>
<th>dₜ = 2°</th>
<th>dₜ = 3°</th>
<th>dₜ = 4°</th>
<th>dₜ = 6°</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.896</td>
<td>0.962</td>
<td>0.069</td>
<td>0.0785</td>
<td>0.182</td>
<td>0.494</td>
<td>0.464</td>
</tr>
<tr>
<td>2</td>
<td>0.827</td>
<td>0.928</td>
<td>0.109</td>
<td>0.0204</td>
<td>0.102</td>
<td>0.210</td>
<td>0.654</td>
</tr>
<tr>
<td>5</td>
<td>0.669</td>
<td>0.877</td>
<td>0.191</td>
<td>0.0028</td>
<td>0.0480</td>
<td>0.172</td>
<td>0.268</td>
</tr>
<tr>
<td>10</td>
<td>0.469</td>
<td>0.658</td>
<td>0.287</td>
<td>0.0023</td>
<td>0.0345</td>
<td>0.160</td>
<td>0.434</td>
</tr>
<tr>
<td>20</td>
<td>0.295</td>
<td>0.465</td>
<td>0.366</td>
<td>0.0045</td>
<td>0.0590</td>
<td>0.330</td>
<td>0.612</td>
</tr>
<tr>
<td>50</td>
<td>0.126</td>
<td>0.201</td>
<td>0.373</td>
<td>0.0068</td>
<td>0.0303</td>
<td>0.350</td>
<td>0.612</td>
</tr>
<tr>
<td>100</td>
<td>0.0521</td>
<td>0.0818</td>
<td>0.363</td>
<td>0.0200</td>
<td>0.450</td>
<td>0.575</td>
<td>1.000</td>
</tr>
</tbody>
</table>

The area containing magnetic anomalies falls off sharply at higher thresholds, allowing a greater probability of chance correlation. At lower thresholds the entire map is considered anomalous, and so again it is easier to produce randomly the observed correlation. Over length scales at which the magnetic anomalies can be described as strongly coherent (dₜ < 4 degrees), the probability of a chance correlation is relatively small (q < 0.16).

A genetic correlation requires that the valley networks and magnetic anomalies are the same age. A significant majority of valleys are Noachian (70–92%) [Scott and Dohm, 1992; Carr, 1996]; many of those that are younger are not in the southern highlands and so are excluded from our survey. The magnetic anomalies, because of their inferred deep-crustal origin (see proposition 1 below) without surface manifestation, must also be very old. However, even the oldest valley networks individually preserve only some part of the Noachian that was not subsequently locally erased, whereas the magnetic anomalies probably reflect a greater span of crustal history. In other words, some valley networks that were associated with magnetic anomalies may have been resurfaced, whereas other valley networks may have formed subsequent to emplacement of the magnetic anomalies by hydrothermal or other processes. The normalized excess of correlated valleys and magnetic anomalies (p - p₀)p₀ (Table 2) may be taken as representative of the fraction of valleys for which a genetic correlation may be inferred, between about one quarter and one third.

The role of hydrothermal circulation in the relationship between valley networks and magnetic anomalies may now be described in terms of a central hypothesis composed of two main propositions, defined and discussed in the following sections.
al., 1979]. Therefore deep crustal magnetization of Mars is reasonable.

[44] The mineral composition of the magnetized material producing the magnetic anomalies is likely to contain magnetite or hematite as the primary magnetic carrier. Although magnetite is generally favored, there is much support for hematite. E.g. Connerney et al., 1999]. Keleteteke et al., (2000) show that for an applied magnetic field of 0.1 mT (about twice the strength of the Earth's present geomagnetic field), multidomain hematite reaches maximum TRM saturation, whereas magnetite reaches only a few percent thereof.

4.2.2. Proposition 2. [45] The second proposition is that hydrothermal discharge attending crustal formation processes in the southern hemisphere of Mars was sufficient to provide the water necessary to carve the valley networks. We assume that where instantaneous discharges predicted by our models are too small to do significant geomorphic work, topographical variations and near-surface heterogeneities in the host rock permeability (especially in the horizontal dimensions) are sufficient to concentrate discharge to the required levels. Hydrothermal systems on Earth exhibit such discretized concentration of outflow, as evidenced by the presence of geysers and springs rather than diffuse outflow everywhere above the magmatic intrusion. The Martian valley networks are characterized by low drainage densities, implying again that crustal heterogeneities may localize discharge.

[46] Testing this proposition requires knowledge of the amount of water necessary to erode the valley networks, the total amount of water available to hydrothermal systems, and the actual hydrothermal discharge produced. An estimate of the required water volume for valley erosion may be made from values of areal coverage, drainage density, valley cross section, and sediment-to-water ratio. Using a map of drainage densities [Carr and Chuang, 1997], we estimate the total area covered by valley networks to be about 1.4 × 10^17 km^2. Since we estimated earlier that only one quarter of the valley networks may preserve direct interaction with hydrothermal systems, we use a reduced effective area of 3.5 × 10^16 km^2. Carr and Chuang [1997] calculated a globally averaged drainage density of 0.0032 km^-1. The product of effective area and drainage density, multiplied by typical valley width and height (5 km and 150 m, respectively), results in 8.6 × 10^16 m^3 of removed material. Sediment-to-water ratios may range from 1.4 to 1:1000 [Gulick, 1998, and references therein], implying that the volume of water required to erode the valleys was between 3.5 × 10^16 and 8.6 × 10^16 m^3, equivalent to a global water layer between 0.2 and 60 m deep. These values are well below the estimate of hundreds of meters for the global crustal inventory, indicating that discharge from the valleys had a negligible impact on the global water budget. They further imply that discharged water need not have been recharged to the crust.

[47] The total hydrothermal discharge produced is computed as follows. First, we assume that each magma chamber contributing to crustal formation intrudes into steady ambient temperature and pressure conditions; this is most likely if intrusions that formed the southern highlands moved between different loci rather than spreading from a single location [Grimm, 2000]. In this way, the discharge contribution from a single intrusion is just that of its equivalent HYDROTHERM model. We assume further that the crust covering the area occupied by valley networks (as calculated above) is eventually built up to a depth of 20 km by magma chambers packed side by side and one on top of the other. We sum the total discharges from each intrusion to calculate the total mass of surface water produced. We consider only those intrusions that contribute to magnetic anomalies, discarding other intrusive events. Determining the relative contribution of individual intrusions is not possible, but the probability p0 may be used as an indicator of the appropriate fraction to be considered. Its value for a magnetic threshold of 10 nT, i.e., 0.469 (Table 2), is applied to all of our results.

Figure 11. Summary of global discharge production (by mass, in kg) for various model parameters. The two dashed lines denote the bounds on total required discharge. Represented are discharge values for different magma chamber aspect ratios, for host rock permeabilities, for magma chamber volumes, and for the baseline model with different assumed water table depths. Note that the baseline model (D/H = 2, k = 10^-16 m^2, chamber volume = 50 km^3, water table depth = 0 m) appears in each vertical group of bars.

[48] A summary of discharge production over the entire region of interest is shown for different models in Figure 11. The two thick dashed lines denote the bounds on total required discharge. Represented are discharge values for different magma chamber aspect ratios (all with a volume of 50 km^3, host rock permeability of 10^-16 m^2), host rock permeabilities (all chambers have D/H = 2 and a volume of 50 km^3), magma chamber volumes (all chambers have D/H = 2, host rock permeability is 10^-16 m^2), and values for the baseline model with different assumed water table depths.

[49] All models meet or exceed the discharge production requirements, although it should be noted that if the k = 10^-15 m^2 model is assumed to have a water table of 100 m, its production will fall below the minimum required value. Other factors such as evaporation may further reduce the discharge available to carve valleys. The present evaporation mass flux is likely in the region of 4 × 10^-8 kg m^-2 s^-1 [Ingersoll, 1974], which would reduce the effective discharge by as much as an order of magnitude or more. Ice may also reduce discharge (section 3.6). A 1 km thick layer of subsurface ice in our baseline model causes a 75% drop in total discharge. In models with other parameter values (e.g. greater magma chamber depth and smaller host rock permeability), ice may inhibit discharge more severely, if not completely.

5. Conclusions

[50] In numerical models of Martian hydrothermal systems we explored the control on surface discharge of magma chamber depth, volume, aspect ratio, and host rock permeability and porosity. Discharge has an approximately linear relationship to magma chamber volume and host rock permeability (in the range 10^-17-10^-15 m^2). The influences of depth and aspect ratio are weaker, and that of porosity is negligible.

[51] Some geochemical aspects of Martian hydrothermal systems were considered by calculating water-rock ratios in our numerical models at various reaction temperatures. Ratios tend to be low but sufficiently large at mean permeabilities >10^-16 m^2 for
groundwater flow to be sustained, consistent with the expected storage of water in alteration assemblages. The presence of a short-lived vapor-dominated zone in our model with high host rock permeability ($10^{-15}$ m$^2$) and shallow chamber depth (2 km) suggests that hydrothermal alteration processes may be responsible for the observed relative abundances of certain salts in the Martian soil, although other forms of alteration should not be excluded.

[51] Crustal formation processes which formed the magnetic anomalies observed on Mars today may have been attended by hydrothermal circulation that also provided surface water for valley network erosion. This idea is in agreement with the observed spatial correlation between magnetic anomalies and valleys and was tested further within the framework of a central hypothesis made up of two propositions. The first is that the magnetic anomalies formed as intruded crust and that the acquisition of thermoremanence occurred at relatively great depth. The second is that hydrothermal discharge attending global crustal formation processes is sufficient to provide the water necessary to carve the planet's valley networks.

[53] We tested this hypothesis using the numerical models described above, assuming that the crust was formed by the heterogeneously spaced and timed intrusion of multiple magma chambers, each of which produced the discharge predicted by an individual numerical model. We determined that many model configurations in the explored portion of the parameter space were capable of producing sufficient water to erode those valley networks statistically related to hydrothermal circulation. In particular, modest crustal permeabilities of $10^{-10} - 10^{-15}$ m$^2$ can produce the discharge required to carve valleys and satisfy geochemical constraints, even in the presence of mitigating factors such as evaporation and finite water table depth.

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References
Hayba, D. O., and S. E. Ingebritsen, The computer model HYDRO-