

Controls on Martian hydrothermal systems: Application to valley network and magnetic anomaly formation

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[1] Models of hydrothermal groundwater circulation can quantify limits to the role of hydrothermal activity in Martian crustal processes. We present here the results of numerical simulations of convection in a porous medium due to the presence of a hot intruded magma chamber. The parameter space includes magma chamber depth, volume, aspect ratio, and host rock permeability and porosity. A primary goal of the models is the computation of surface discharge. Discharge increases approximately linearly with chamber volume, decreases weakly with depth (at low geothermal gradients), and is maximized for equant-shaped chambers. Discharge increases linearly with permeability until limited by the energy available from the intrusion. Changes in the average porosity are balanced by changes in flow velocity and therefore have little effect. Water/rock ratios of ~ 0.1 , obtained by other workers from models based on the mineralogy of the Shergotty meteorite, imply minimum permeabilities of 10^{-16} m² during hydrothermal alteration. If substantial vapor volumes are required for soil alteration, the permeability must exceed 10^{-15} m². 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} m² and intrusion volumes as low as 50 km³, 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

1. Introduction

[2] 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 [*Gulick*, 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 may also have altered the Martian crust and further produced weathering products and soil [*Griffith and Shock*, 1997; *Newsom et al.*, 1999]. The greatest consequence of hydrothermal activity on Mars may be its ability to sustain life [*Shock*, 1996; *Farmer*, 1996].

[3] We present here numerical models of the thermal convection of groundwater in a porous host rock due to the presence of an intruded magma chamber. An extensive portion of the available parameter space is explored in order to quantify the effects that magma chamber volume, depth, aspect ratio, and host rock permeability and porosity have on surface discharge. The results of this general approach are then applied to some specific issues. First, we use geochemical constraints to bound the permeability of the Martian crust. Second, we test the hypothesis that hydrothermal circulation can explain the putative correlation observed between valley networks and magnetic anomalies [*Jakosky and Phillips*, 2001]. We suggest that large amounts of water were circulated throughout the southern-highlands crust due to magmatic intrusion and that the portion of this water discharged to the surface can quantitatively account for the valley networks preserved since the end of crustal formation.

2. Model

[4] 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 HTpost (P. S. Hsieh, USGS, unpublished material, 2000). HYDROTHERM can model temperatures from about 0° to 1200°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

$$\mathbf{v}_i = -\frac{k_{\text{rel}}^i k}{\mu_i} (\nabla p + \rho_i g \nabla z), \quad (1)$$

where \mathbf{v}_i is the Darcy velocity, k is intrinsic permeability, μ_i is dynamic viscosity, p is pressure, ρ_i is density, g is gravitational acceleration, and z is depth. The relative permeability k_{rel}^i 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

$$\frac{\partial}{\partial t} (n S_i \rho_i) + \nabla \cdot (\rho_i \mathbf{v}_i) = 0, \quad (2)$$

where n is porosity, t is time, and S_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

$$\frac{\partial}{\partial t} \left[(1-n) \rho_r h_r + n \sum_i S_i \rho_i h_i \right] + \nabla \cdot \left[\sum_i k_{r,i} \rho_i h_i \mathbf{v}_i \right] - \nabla \cdot K_m \nabla T = 0 \quad (3)$$

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 temperature 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. *Gulick* [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 significantly 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]. Note that our baseline model is intended only as an example model from which we later deviate extensively 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., *Kohlstedt 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 1%. Deviations from the baseline model include host rock permeabilities of 10⁻¹⁷ and 10⁻¹⁵ m², porosity of 25%, magma chamber aspect ratios of 0.2 and 20, 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 geochemical alteration in the system. For each time step, a measure of water-to-rock ratio for reactions above a particular temperature 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 impermeable, 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

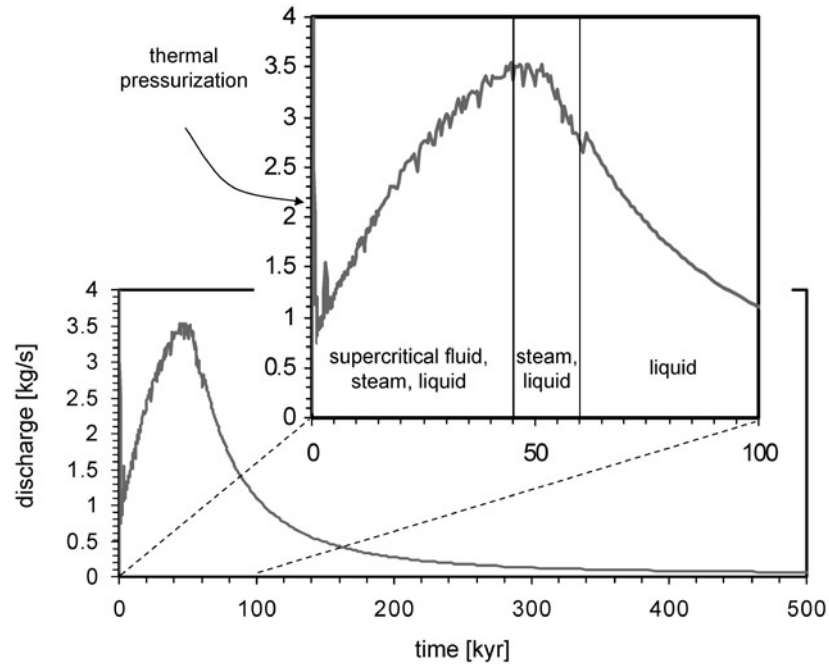


Figure 1. Surficial discharge from the baseline model (magma chamber depth 2 km, volume 50 km^3 , aspect ratio 2, host rock permeability 10^{-16} m^2). The regions separated by solid vertical lines in the enlarged portion of the figure denote the different phases present within the model. The magma chamber 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.

the liquid phase remains, the exponential decay of discharge becomes much smoother and dies out at about 2.8×10^6 years.

3.1. Permeability

[14] 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^{-9} 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.

[15] 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.

[16] An important feature of the $k = 10^{-15} \text{ m}^2$ 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.

[17] 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

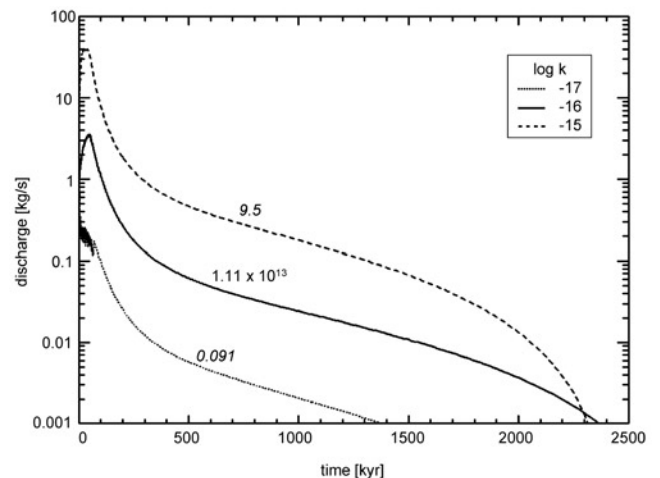


Figure 2. Surficial discharge from a 2 km deep magma chamber of volume 50 km^3 for host rock permeabilities of 10^{-17} , 10^{-16} , and 10^{-15} m^2 . The total discharge of the $k = 10^{-16} \text{ m}^2$ model, in kilograms, appears above its curve. The numbers above the other curves denote the fraction of this discharge that their corresponding models produced.

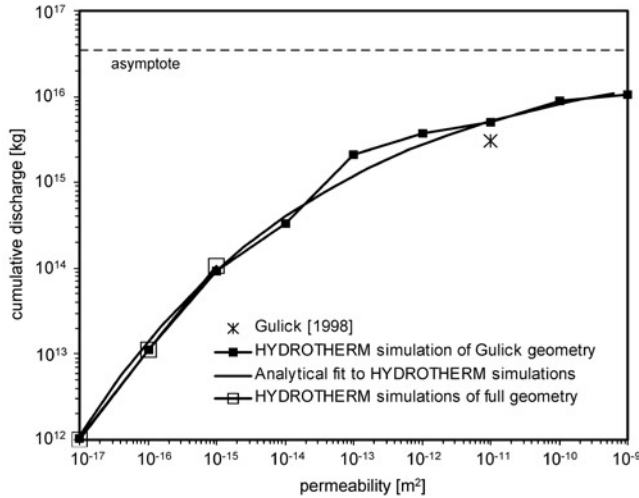


Figure 3. Discharge results from HYDROTHERM models of a geometry similar to those of *Gulick* [1998]. An analytic fit to these data, with its large permeability asymptote, is shown. The discharge obtained by the single permeability modeled by *Gulick* is provided for comparison, as are the total discharges of our three full geometry models (from Figure 2). The value obtained by *Gulick* was converted from volume to mass for the purpose of this figure, with an assumed water density of 1000 kg m^{-3} .

chamber and that no BDT was encountered on cooling. By adopting these limitations in HYDROTHERM, the problem is more efficient numerically and allows us to extend the range of assumed permeability (Figure 3) and to compare the results directly to those of *Gulick* [1998]. Our results agreed fairly well, although a small difference resulted owing to our more realistic magma chamber heat flux and possibly other factors such as numerical solution. Overall, we found that the relationship between total discharge and permeability could be represented approximately by

$$D = a - be^{-cK}, \quad (4)$$

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 $K \rightarrow \infty$, $D \rightarrow a$, 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.

[18] For our range of permeabilities, total discharge is approximately 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. Instantaneous discharge is not the same for both geometries, even with $D/H = 2$. Our geometry produces a greater maximum than the simpler geometry (~ 1.75 times as high for $k = 10^{-16} \text{ m}^2$) but drops more rapidly thereafter.

[19] 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 permeabilities the cooling time for a threshold of 450°C (i.e., half the emplacement temperature) is ~ 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 as different rates in each model. The time taken for the chamber to cool to 250°C was 67, 61, and 40 kyrs for permeabilities of 10^{-17} , 10^{-16} , and 10^{-15} m^2 , respectively.

3.2. Volume

[20] *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 ~ 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 therefore an increase in the discharge lifetime), explains the observed total discharge increase.

[21] Although not obvious on the semilog plot of Figure 4, the discharge produced by the 2000 km^3 chamber drops off at ~ 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 4×10^{-17} to 10^{-16} m^2 .

3.3. Depth

[22] 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 ~ 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 geothermal gradient in these models makes magma chamber cooling times only marginally sensitive to changes in depth. All three chambers cool to half their emplacement temperature (450°C) in ~ 30 kyr. The 2 km deep chamber cools to 250°C in 61 kyr, while the 15 km chamber takes 71 kyr. Flow from deeper chambers must

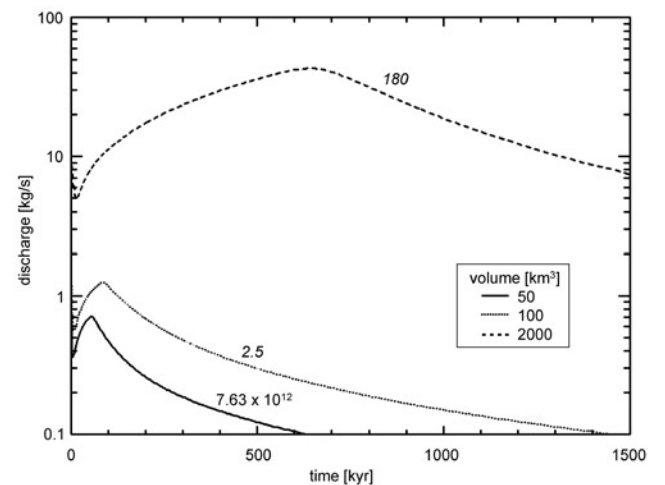


Figure 4. Surficial discharge from magma chambers of volume 50, 100, and 2000 km^3 , all at a depth of 8.5 km, an aspect ratio of 2, and with host rock permeability of 10^{-16} m^2 . The total discharge of the 50 km^3 model appears above its curve. The numbers above the other curves denote the fraction of this discharge that their corresponding models produced.

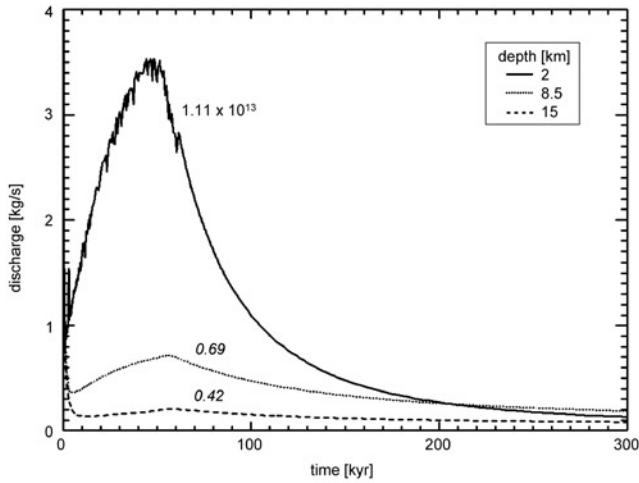


Figure 5. Surficial discharge from 50 km³ magma chambers at depths of 2, 8.5, and 15 km, all with aspect ratio 2, and with host rock permeability of 10⁻¹⁶ m². The total discharge of the 2 km deep chamber appears above its curve. The numbers above the other curves denote the fraction of this discharge that their corresponding models produced.

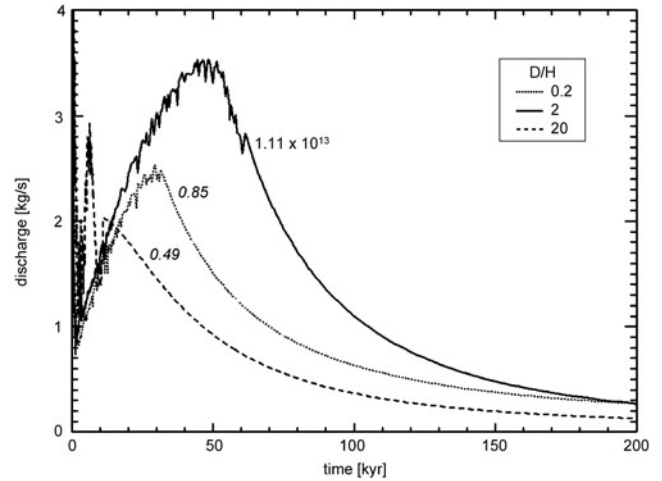


Figure 7. Surficial discharge from 50 km³ magma chambers with $D/H = 0.2, 2,$ and $20,$ all at a depth of 2 km, and with host rock permeability of 10⁻¹⁶ m². The total discharge at $D/H = 2$ appears above its curve. The numbers above the other curves denote the fraction of this discharge that their corresponding models produced.

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

3.4. Aspect Ratio

[23] 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 km³. 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 total 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).

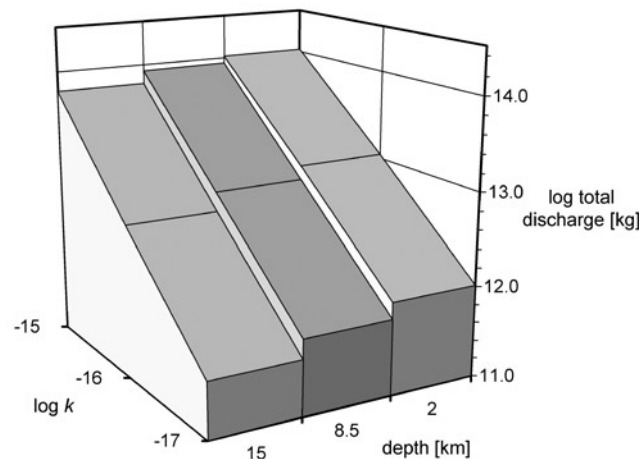


Figure 6. Total discharge from 50 km³ magma chambers of aspect ratio 2 at depths of 2, 8.5, and 15 km and host rock permeabilities of 10⁻¹⁷, 10⁻¹⁶, and 10⁻¹⁵ m².

[24] 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 peak 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 >10⁻¹⁶ m², 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

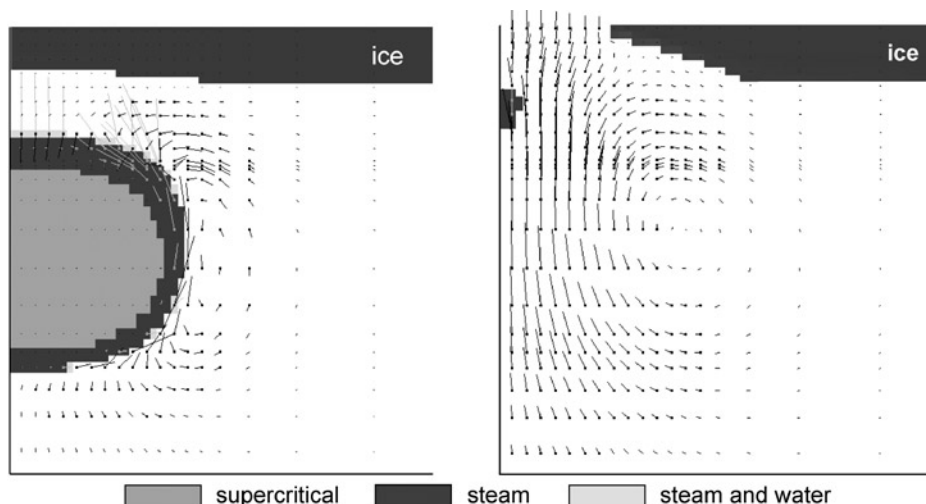


Figure 8. Baseline model with an initially 1 km thick subsurface ice layer shown at (left) 15 kyr and (right) 59 kyr. Both horizontal and vertical dimensions in each image are approximately 5 km. Water and steam velocities are shown by black and gray streamers, respectively. Only every other row and column are shown. Shaded regions (apart from the ice) represent different fluid phases as described in the key. At 59 kyr, only a small remnant of steam remains, and the magma chamber has become as permeable as the host rock.

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

[28] *Gulick* [1998] suggested that a 2 km thick subsurface permafrost layer above a 50 km^3 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 ΔT . During melting the material is assigned an augmented specific heat given by

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

where subscripts S and L refer to the solid and liquid phases, respectively, and L is the latent heat of fusion. We used $c_L = 1000 \text{ J kg}^{-1} \text{ K}^{-1}$, $L = 3.34 \times 10^5 \text{ J kg}^{-1}$, and $T = 5^\circ\text{C}$, giving $c_L = 6.78 \times 10^4 \text{ J kg}^{-1} \text{ K}^{-1}$.

[29] 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 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

[30] 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

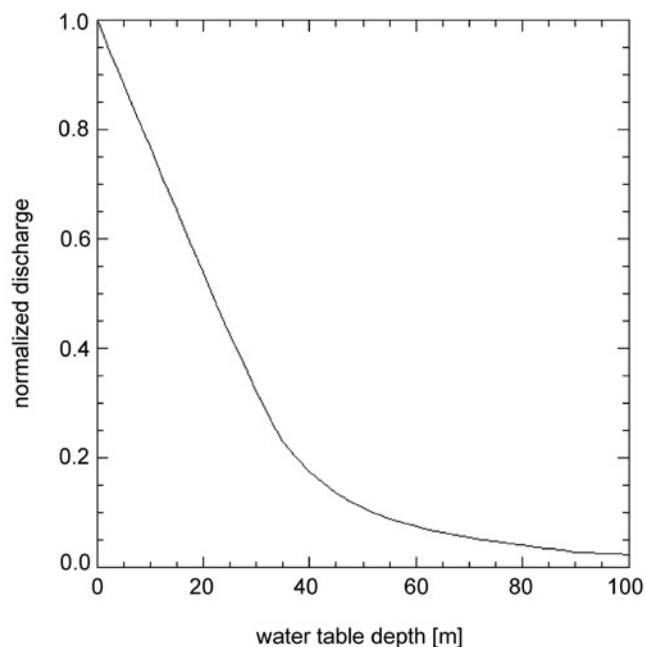


Figure 9. Discharge as a function of water table depth, normalized to the zero depth water table value.

Table 1. Water/Rock Ratios by Mass Calculated for 8.5 km Deep Chambers of the Indicated Volumes and Host Rock Permeabilities k

Threshold Temperature, °C	$k = 10^{-15} \text{ m}^2$ Volume = 50 km ³	$k = 10^{-16} \text{ m}^2$ Volume = 50 km ³	$k = 10^{-16} \text{ m}^2$ Volume = 2000 km ³
150	0.12	0.025	0.067
200	0.18	0.037	0.10
250	0.28	0.054	0.15

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 $k = 10^{-15} \text{ m}^2$ model give the highest W/R for all threshold temperatures, ~ 5 times that of the $k = 10^{-16} \text{ m}^2$ 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 ($>50^\circ\text{C}/\text{km}$) 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, $\sim 8\%$ 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 km³ 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 Sorey [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

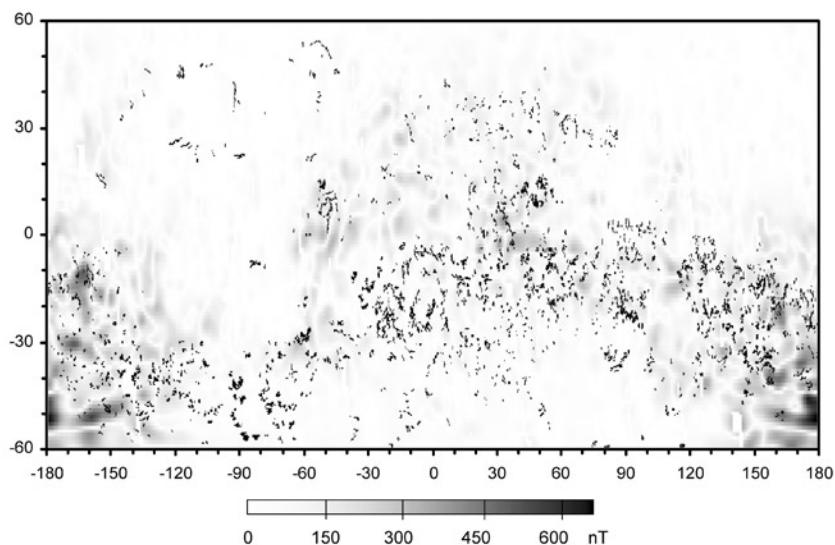


Figure 10. Overlay of the absolute values of the 200 km vertical magnetic anomalies [Purucker et al., 2000] and the valley networks [Kieffer, 1981] (cylindrical projection). Valleys were not mapped for latitudes below 60°S . The anomalies range from 0 to 670 nT.

Table 2. Statistical Data for the Correlation Between Valley Networks and Magnetic Anomalies

Magnetic Threshold, nT	p_0	p	$(p - p_0)/p$	Q			
				$d_m = 2^\circ$	$d_m = 3^\circ$	$d_m = 4^\circ$	$d_m = 6^\circ$
1	0.896	0.962	0.069	0.0785	0.182	0.494	0.464
2	0.827	0.928	0.109	0.0204	0.102	0.210	0.654
5	0.669	0.827	0.191	0.00228	0.0480	0.172	0.268
10	0.469	0.658	0.287	0.00223	0.0345	0.160	0.434
20	0.295	0.465	0.366	0.00459	0.0590	0.166	0.343
50	0.126	0.201	0.373	0.0686	0.303	0.330	0.612
100	0.0521	0.0818	0.363	0.260	0.450	0.575	1

minimum of about two thirds hydrostatic pressure) and high temperatures. At lower permeabilities, fluid rising from the chamber does not transport enough heat upward to elevate temperatures to the required levels for steam production.

[36] If substantial quantities of vapor are required for soil alteration, then the abundance of steam in our high-permeability model may point to 10^{-15} m^2 as a lower bound to permeability, a somewhat tighter constraint than that imposed by W/R ratio alone. Alternatively, if volcanic aerosols [Newsom *et al.*, 1999] or basalt palagonitization [McSween and Keil, 2000] produce sufficient soil alteration, then no such limit is necessary.

4.2. Relationship of Hydrothermal Circulation to Valley Networks and Magnetic Anomalies

[37] The apparent spatial correlation between valley networks and magnetic anomalies suggests a link between processes involving the acquisition of thermoremnance in a cooling intrusion and processes involving the production of surface water for the creation of fluvial channels, i.e., hydrothermal circulation. An overlay of maps (Figure 10) showing the valley networks [Kieffer, 1981] and the vertical component of the magnetic anomalies as measured by the Mars Global Surveyor magnetometer [Acuña *et al.*, 1999] inferred at a constant altitude of $\sim 200 \text{ km}$ [Purucker *et al.*, 2000] visually suggests that the two are correlated [Jakosky and Phillips, 2001].

[38] A statistical analysis of the relative positions of valley networks and magnetic anomalies quantifies the correlation. A binomial test was performed, with success defined as the occurrence of a valley network and a magnetic anomaly (with vertical component above a specified threshold) in the same data bin. We consider only the southern highlands in these calculations. The valley network map [Kieffer, 1981] contains the coordinates of the 0.25 by 0.25 degree bins that contain valleys; the number and length of the valleys are not quantified. The magnetic anomaly map [Purucker *et al.*, 2000] is binned at 1 by 1 degree and was regridded at 0.25 degree to match the valley grid. The number of degrees of freedom must, however, be adjusted to match the true spatial resolution, which is limited by the magnetic data. The total number of 0.25 degree bins is therefore multiplied by the quantity

$$\varepsilon = \left(\frac{0.25}{d_m} \right)^2,$$

where d_m is the length scale in degrees for magnetic resolution. We determined that the variogram for the 200 km magnetic field reaches half of its sill (asymptotic) value in 200 km and 90% of the sill value in 400 km, in agreement with the rule of thumb for potential fields that the spatial resolution is approximately equal to measurement altitude. Therefore $d_m \sim 200 \text{ km}$ (~ 3 degrees) may be most appropriate, but we consider a range from 2 to 6 degrees.

[39] Relevant probabilities may be obtained using the total number N of 0.25 by 0.25 degree bins in the region of interest, the number n of these bins containing valleys, the number m of bins with magnetic anomalies above a specified threshold, and the

number C of correlations. The observed correlation probability is then $p = C/n$, while the expected correlation probability for a valley placed randomly in the region of interest is $p_0 = m/N$. The probability q of obtaining C or more correlations if the n valleys are placed randomly in the region of interest can be obtained using the cumulative binomial distribution with εC , εn , and p_0 as the minimum number of successes, the number of independent trials, and the probability of a single success, respectively. A low value of q implies that the relative distribution of valleys and magnetic anomalies on Mars is not what one would expect from chance.

[40] The results for a range of magnetic anomaly thresholds (Table 2) indicate that minimum chance probabilities occur near a 10 nT threshold for the vertical magnetic field at 200 km altitude. 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 ($d_m < 4$ degrees), the probability of a chance correlation is relatively small ($q < 0.16$).

[41] 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 $(p - p_0)/p_0$ (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.

[42] 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.

4.2.1. Proposition 1. [43] The first proposition is that the magnetic anomalies formed as intruded crust and that the acquisition of thermal remnant magnetization (TRM) occurred at relatively great depth. The strong observed magnetizations in the Martian crust of 20–40 A m² [Connerney *et al.*, 1999; Grimm, 2000] imply magnetization depths of up to a few tens of kilometers. The presence of magnetic anomalies in Arabia Terra, which has been strongly resurfaced [McGill, 2000; Hynes and Phillips, 2001], also points to a deep origin. On Earth, Layer 3 gabbros of the oceanic crust are magnetized to the extent that they contribute between 25 and 75% of the observed marine anomalies [Pariso and Johnson, 1993], while the underlying mantle is unmagnetized [Wasilewski *et*

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]. *Kletetschka 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 heterogeneity 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 1.4×10^7 km². 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 3.6×10^6 km². *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 8.6×10^3 km³ 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 3.5×10^4 and 8.6×10^6 km³, 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 p_0 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.

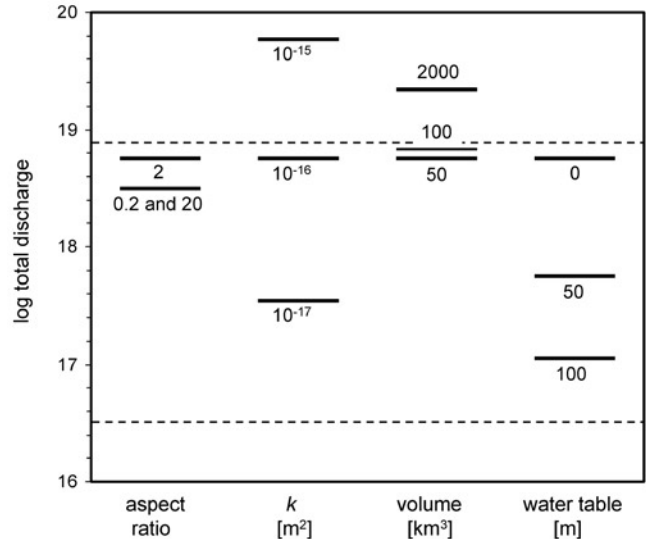


Figure 11. Summary of global discharge production (by mass, in kg) for various model parameters. The two dashed lines indicate the bounds on the mass of water required to erode the valley networks. Represented are discharge values for different magma chamber aspect ratios, for host rock permeabilities, for magma chamber volumes, and for the baseline model with different assumed water table depths. Note that the baseline model ($D/H = 2$, $k = 10^{-16}$ m², chamber volume = 50 km³, 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 km³, host rock permeability of 10^{-16} m²), host rock permeabilities (all chambers have $D/H = 2$ and a volume of 50 km³), magma chamber volumes (all chambers have $D/H = 2$; host rock permeability is 10^{-16} m²), 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 $k = 10^{-17}$ m² 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} m²). 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 $>10^{-16}$ m² 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} m²) 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.

[52] 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 thermoremnance 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 its 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^{-16} – 10^{-15} m² 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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