Thermal demagnetization of Martian upper crust by magma intrusion

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[1] The absent or weak magnetic field above large Martian volcanoes may provide constraints on their formation and the carrier of magnetization. We consider the ability of magma intrusions to thermally demagnetize the shallow crust beneath volcanoes through heat conduction and hydrothermal circulation. If magnetization is dominated by magnetite, the volume of crust that is thermally demagnetized is similar to the volume of crust emplaced since the dynamo field disappeared. If magnetization is dominated by pyrrhotite, the volume of crust that is demagnetized is typically more than twice the volume of magma intruded. Hydrothermal circulation contributes negligible additional thermal demagnetization, beyond that from heat conduction alone, for permeabilities less than \(10^{-15}\) m\(^2\). Citation: Ogawa, Y., and M. Manga (2007), Thermal demagnetization of Martian upper crust by magma intrusion, Geophys. Res. Lett., 34, L16302, doi:10.1029/2007GL030565.

1. Introduction

[2] The history, location and volume of magmatic activity reflect the thermal evolution of a planet’s interior. On Mars, extrusive magmatic history can be assessed by counting craters on regions resurfaced by lava flows or products of explosive eruptions. Probing magma intruded below the surface must rely on remote, satellite measurements. Gravity measurements can be used to identify and estimate the volume of (now solid) subsurface magma bodies [e.g., Kiefer, 2004]. Crustal magnetization, in particular the weak or absent magnetic field above Martian volcanoes and volcanic plateaus, may provide a second constraint on the history and volume of subsurface magma bodies.

[3] The magnetic signature of volcanoes records information about their structure and evolution provided the history of the magnetic field and the origins of magnetized and demagnetized crust are known. The present Martian magnetic field is produced by crustal magnetization. The strength, carrier, and depth of magnetization are not well constrained by satellite measurements of the magnetic field [Connerney et al., 2004]. The depth distribution of magnetization is difficult to determine but probably confined to the upper 50 km or less [e.g., Nimmo and Gilmore, 2001; Arkani-Hamed, 2002]. On the other hand, the history of its acquisition is slightly better constrained. The strongest fields exist above the old southern highlands [Connerney et al., 1999]. The large Hellas and Argyre impact basins are virtually devoid of crustal fields, suggesting that the impacts that made these basins removed any pre-existing magnetization, and the dynamo that generated the early Martian field had ceased so that any melt created by the impacts did not acquire a magnetization [Acuna et al., 1999]. The near-absence of magnetic field above the Tharsis region suggests that the dynamo did not restart since the formation of the bulk of Tharsis plateau [Arkani-Hamed, 2004; Johnson and Phillips, 2005].

[4] On Earth, the magnetic field above volcanoes has been used to estimate their ages, construction processes, and volume of subsurface magmas [e.g., Maia et al., 2005]. An important difference between the two planets is that on Earth magmatism always occurs in the presence of a core-generated field. In contrast, most Martian volcanoes formed after the dynamo ceased [e.g., Arkani-Hamed, 2004] with the exception, perhaps, of Hadriaca Patera [Lillis et al., 2006]. Therefore, the magnetic signature of Martian volcanoes should show signs of demagnetization, and magnetic anomalies due to remagnetization should be absent or very weak. Hydrothermal circulation is potentially a dominant cause of demagnetization at Martian volcanoes [e.g., Nimmo, 2000; Harrison and Grimm, 2002; Smrekar et al., 2004; Solomon et al., 2005]. Away from volcanoes, impacts [Mohit and Arkani-Hamed, 2004] and low-temperature hydration [e.g., Briel and Petersen, 1983] may instead dominate demagnetization of the shallow crust.

[5] Here we reevaluate the ability of subsurface magmatic activity to thermally demagnetize the Martian crust through hydrothermal circulation and heat conduction and consider the implications for the magnetic carrier and magmatic history. We determine the amount of thermal demagnetization that is possible as a function of crust permeability and depth of magma emplacement. We also quantify the temperatures and volumes of water involved in the hydrothermal circulation.

2. Numerical Model

[6] When magma is intruded into the crust, heat transferred to groundwater causes hydrothermal convection. At depths less than 3–4 km water may boil and at greater depths becomes a supercritical fluid. To simulate such 2-phase gas-liquid hydrothermal convection, we use the code HYDROTHERM developed by Hayba and Ingebritsen [1997], which also models supercritical fluids.

[7] We focus on hydrothermal convection induced by the intrusion and cooling of sills. We assume instantaneous emplacement of a 1140°C sill that is 4 km long and 500 m thick, somewhat thicker than the O(100) m dike widths [Ernst et al., 2001] and ~150 m sill thicknesses [Wilson and Mouginis-Mark, 2003] estimated on Mars. We neglect the latent heat of crystallization, which would extend the
cooling time of the sill by up to 30% but not affect temperatures driving circulation. The 2D Cartesian computational domain extends from the surface to a depth of 13.5 km and horizontally 18 km beyond the end of the sill. The upper surface is an isothermal, no-flow boundary with a temperature of 0.5°C, and pressure of 0.1 MPa. The bottom surface is a no-flow, fixed heat flux boundary. Thermal conductivity is 1.6 W/mK [Ingebritsen et al., 1994]. Gravity is 3.71 m/s². We assume an open boundary on the vertical boundaries of the computational domain with a hydrostatic pressure gradient and fixed geothermal temperature gradient.

Rather than choosing a preferred model for subsurface properties [e.g., Clifford and Parker, 2001], we consider simpler models and vary the most uncertain parameters in order to assess the sensitivity of our conclusions to model properties. We vary four parameters: permeability \( k \) of the aquifer, from \( 10^{-18} \) m² to \( 10^{-13.5} \) m²; aquifer porosity from 5 to 30%; depth \( D \) of the sill from 1 to 10 km; temperature gradient is 15, 25 or 37.5°C/km, corresponding to heat flows between 0.024 and 0.060 W/m², a range covering that estimated for the first approximately Ga of Mars [e.g., Williams and Nimmo, 2004]. The sill porosity is 0.05 and its permeability is \( 10^{-18} \) m². We neglect the temperature- and pressure-dependence of all variables, except for the properties of water. To assess the effects of hydrothermal, possibly multiphase, convection we compare results with the case in which all heat transfer occurs by thermal conduction alone.

3. Hydrothermal Convection Around a Sill

To estimate the potential of hydrothermal circulation to demagnetize the upper Martian crust, we track the evolution of temperature and fluid flow in response to the

Figure 1. (a) Highest temperature ever reached \( (T_{\text{max}}) \) after intrusion. Only part of aquifer is shown. Contour interval is 50°C. (b) Vertical distribution at distance 0 of temperature and its evolution after intrusion. In both Figure 1a and Figure 1b, \( k = 10^{-14.5} \) m², \( D = 3 \) km, and \( dT/dz = 37.5°C/km \). Asymmetric heating is dominated by thermal convection and vaporization. The area extending several hundred meters above the sill is a steam layer.

Figure 2. (a) Area ratio \( (A_{\text{TM}}) \) as a function of permeability \( k \) (solid lines) and porosity (various symbols). (b) Nusselt number \( (Nu) \) and (c) distribution of four phases (water, steam, 2-phase mixture, and supercritical fluid) 5 ka after intrusion, as a function of permeability \( k \).
intrusion of the sill. We assume that thermal demagnetization occurs within a zone that has experienced a temperature above the Curie temperature ($T_c$), though demagnetization may begin at the blocking temperature, as much as 100°C lower. We ignore the potential for reacquisition of magnetization. We thus monitor the highest temperature ever reached at each point in the subsurface and compare this temperature with $T_c$ for various minerals that have been proposed to dominate magnetization: pyrrhotite ($T_c \approx 320$°C), magnetite ($T_c \approx 580$°C), and hematite ($T_c \approx 670$°C) [e.g., Collinson, 1997; Dunlop and Arkani-Hamed, 2005].

3.1. Effect of Permeability

Figure 3. (a) Distribution of water-rock ratio ($W/R$); $k = 10^{-14.5}$ m$^2$, $D = 3$ km, $dT/dz = 37.5$°C/km, the same case as Figure 1. The flow of water above 250°C is integrated for 10 ka. Vertical asymmetry is caused by convective heating and vaporization. (b) Variation of the average water-rock ratio ($W/R_{th}$) as a function of $k$. $T_{th}$ represents the threshold temperature above which the water is measured. $W/R_{th}$ increases with increasing $k$, reflecting the strength of convection.

The area extending several hundred meters above the sill remains almost isothermal ($\sim 300$°C) because it is a layer of steam and temperature is buffered by the latent heat of vaporization.

In order to characterize the amount of crust that could undergo thermal demagnetization, we define a dimensionless quantity termed the area ratio, $A_{Tc}$ (a 2D equivalent of a volume ratio: the cross-section area that experiences temperatures greater than $T_c$ divided by the cross-section area of the sill). Figure 2a shows $A_{Tc}$ for various $T_c$ as a function of crust permeability, $k$, where $D$ and $dT/dz$ are fixed at 3 km and 37.5°C/km, respectively. The value for the conduction-only case (no flow permitted) is shown for reference. $A_{320}$ increases slightly as $k$ increases, reaches a maximum for $k \approx 10^{-16}$ m$^2$, and finally decreases significantly with further increases in $k$. The same pattern occurs for other values of $T_c$, for example $A_{580}$, although $A_{580}$ has a maximum for $k$ around $10^{-14.5}$ m$^2$. $A_{670}$ always stays smaller than $A_{320}$. The maximum of $A_{Tc}$ is $\sim 2.3$ (at $k \approx 10^{-15.75}$ m$^2$ for $T_c = 320$°C). Figure 2a shows that aquifer porosity has little effect on our conclusions.

Figure 4. The area ratio ($A_{Tc}$) as a function of intrusion depth $D$ for (a) $dT/dz = 15$°C/km and (b) $dT/dz = 25$°C/km. Each inset shows contours of $T_{max}$ after 20 ka, with a contour interval of 50°C, for the case with deepest sill ($D = 10$ km) and $k = 10^{-15}$ m$^2$. [10] Figure 1a shows isotherms of the highest temperature ever reached at each point in the subsurface and compare this temperature with $T_c$ for various minerals that have been proposed to dominate magnetization: pyrrhotite ($T_c \approx 320$°C), magnetite ($\sim 580$°C), and hematite ($\sim 670$°C) [e.g., Collinson, 1997; Dunlop and Arkani-Hamed, 2005].
A limit on $A_{tc}$ can be understood from the effects of convection and phase changes on the temperature distribution around the sill. When $k$ exceeds $10^{-15} \text{ m}^2$, strong convection develops and can be quantified by changes in the Nusselt number, $Nu$, a dimensionless number defined as the ratio of total heat transferred divided by the heat that would be transferred by conduction alone. Figure 2b shows that for $k > 10^{-12} \text{ m}^2$, vigorous convection enhances vertical heat transfer and thus cools the sill more rapidly and heats the aquifer more effectively. This heat, however, is distributed over a greater area and hence $A_{tc}$ decreases. A runaway heating of the aquifer is avoided because of a trade off between convective heating of the aquifer and cooling of the sill, which limits $A_{tc}$. The decrease of $A_{320C}$ with increasing $k$ over the range $10^{-16} \text{ m}^2 < k < 10^{-15} \text{ m}^2$ is caused by the phase change (vaporization) rather than convection itself because of the large amount of energy required to cause the phase change. Figure 2c shows the percentage of each phase of water 5 ka after intrusion of the sill. As $k$ increases, more water vaporizes thus consuming more heat and transporting it across the zone with vapor as latent heat. We also assess the effect of sill size. Calculations with a sill 5 times as wide or half as thick give similar $A_{tc}$.

While we focus on thermal demagnetization, magnetic minerals may also lose (or acquire) magnetization if they are chemically altered, for example, through oxidation promoted by water circulation. The ratio of the water mass that passes through the rock to the mass of the rock is termed the water-rock ratio, denoted $W/R$. Without knowing or modeling the composition and oxidation state of the water we are unable to say anything more quantitative about the potential for demagnetization by chemical alteration.

Figure 3a shows contours of the water-rock ratio defined for each point in the aquifer ($W/R$) for the same case shown in Figure 1. Figure 3b shows the effect of permeability $k$ on a value of $W/R$ averaged over the region that experiences temperatures greater than 250°C. Here we consider two ways of characterizing the water circulation: first we measure only the water above $T_{sh} = 250^\circ\text{C}$ that passes through a given parcel of rock, second we monitor all the water that circulates through the rock. In both cases, $W/R$ increases with increasing $k$, reflecting the strength of thermal convection. The average $W/R$ exceeds 10 for $k > 10^{-14} \text{ m}^2$. For $10^{-16} \text{ m}^2 < k < 10^{-15} \text{ m}^2$, the similarity of both $W/R$ ratios indicates that most mass transfer occurs as steam, consistent with active vaporization for these permeabilities causing the decrease of $A_{320C}$.

### 3.2. Effect of Intrusion Depth and Thermal Gradient

Figure 4 shows $A_{tc}$ as a function of intrusion depth for a range of $k$. The background temperature gradient $dT/dz$ is fixed at (a) 15°C/km and (b) 25°C/km, respectively. Figure 4 shows that deeper intrusion causes a more extensive area to be heated, and hence increases $A_{320C}$. The area heated increases as $dT/dz$ increases. This trend is also true for the conduction-only case and can be understood from the initial temperature difference between the sill and surrounding rocks: if the intrusion is deep and/or $dT/dz$ is high, the temperature of the crust where the sill intrudes is warmer and it is thus easier to raise the temperature to $T_c$. Moreover, for deep intrusion, at high temperatures a supercritical fluid exists and boiling does not occur. Consequently, less heat is consumed in the phase transition and more heat is transported away from the sill as sensible heat.

### 4. Discussion and Conclusions

The near absence of magnetic field above most Martian volcanoes places constraints on the dominant carrier of remanent magnetization and/or the intrusive history beneath the volcano. We note, however, that caution should be exercised in translating field strength at high altitudes to magnetization contrasts. We find that for $k < 10^{-15} \text{ m}^2$, convection has little effect on demagnetization for all carriers of magnetization. If magnetization is dominated by magnetite [e.g., Brachfeld and Hammer, 2006], because $A_{tc} \sim 1$, the volume of crust that is thermally demagnetized is similar to the volume of crust composed of sills emplaced since the field disappeared for all $k$. For magnetite, this conclusion should not be strongly sensitive to model approximations or uncertainties in the temperature at which thermal demagnetization occurs. In this case, the emplacement of most of the magnetized (upper) crust must postdate the termination of the dynamo. Alternatively, if magnetization is dominated by pyrrhotite, the magnetized thickness of crust is $< 20 \text{ km}$ for plausible temperature gradients $> 15 \text{ K/km}$; $A_{tc}$ is large enough that for $k < 10^{-14} \text{ m}^2$ intrusion of $\sim 5 \text{ km}$ of magma, inferred from gravity studies [e.g., Kiefer, 2004], can demagnetize much (>50%) of the magnetized crust. If demagnetization is chemical and requires abundant water, a plausible $k > 10^{-16} \text{ m}^2$ [e.g., Manga, 2004] is required to get significant amounts of water to circulate through the crust. A final possibility is that the magnetization at volcanoes is not small but instead the coherence scale is small enough that the magnetic field, even at 170 km [Lillis et al., 2006], is very weak.

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