Effects of planetary thermal structure on the ascent and cooling of magma on Venus

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Abstract

Magellan radar images of the surface of Venus show a spatially broad distribution of volcanic features. Models of magmatic ascent processes to planetary surfaces indicate that the thermal structure of the interior significantly influences the rate of magmatic cooling and thus the amount of magma that can be transported to the surface before solidification. In order to understand which aspects of planetary thermal structure have the greatest influence on the cooling of buoyantly ascending magma, we have constructed magma cooling profiles for a plutonic ascent mechanism, and evaluated the profiles for variations in the surface and mantle temperature, surface temperature gradient, and thermal gradient curvature. Results show that, for a wide variety of thermal conditions, smaller and slower magma bodies are capable of reaching the surface on Venus compared to Earth, primarily due to the higher surface temperature of Venus. Little to no effect on the cooling and transport of magma are found to result from elevated mantle temperatures, elevation-dependent surface temperature variations, or details of the thermal gradient curvature. The enhanced tendency of magma to reach the surface on Venus may provide at least a partial explanation for the extensive spatial distribution of observed volcanism on the surface.

1. Introduction

The Magellan images of Venus have revealed a broad spatial distribution of volcanic landforms (Head et al., 1991, 1992). Previous work in modeling the ascent of magma on both Venus and Earth (Marsh, 1978, 1982; Marsh and Kantha, 1978; Sakimoto et al., 1992) has indicated that the planetary thermal structure significantly influences magmatic cooling rates and thus the amount of magma that can be transported to the surface before solidification. Since thermal structure may play a prominent role in controlling the distribution and magnitude of planetary volcanism, and given that the thermal structure of Venus is likely to be different from that of Earth (Kaula and Phillips, 1981; Phillips, 1981; Kaula, 1983), it is worthwhile to investigate the effects of different aspects of planetary thermal structure on magma ascent in more detail. In order to understand which aspects of the thermal structure have the greatest influence on the cooling of ascending magma, we have constructed magma cooling profiles for a plutonic buoyant ascent mechanism, and evaluated the profiles for variations in the surface temperature, surface and mantle temperature gradient, and thermal gradient curvature with depth. The cooling curves were then compared to constrain magma ascent velocities, source depths, and body sizes for Venus relative to Earth.
2. Planetary thermal structure

Because of the wide variety of possible thermal structure parameters—and thus variations—it is desirable that the expression for the thermal structure be flexible as well as mathematically convenient. Here, the planetary thermal structure is modeled as:

\[ \frac{T}{T_0} = 1 - \tau \left( 1 - \frac{z}{z_0} \right)^n \]  

which gives dimensionless temperature as a function of dimensionless depth where \( T \) is the temperature, \( T_0 \) is the source depth temperature, \( z \) is the depth, \( z_0 \) is the source depth, and \( n \) is a constant controlling thermal gradient curvature with depth. A dimensionless surface temperature adjustment is accomplished with:

\[ \tau = 1 - \frac{T_s}{T_0} \]  

where \( T_s \) is the surface temperature. Equation (1) is used both for its mathematical simplicity as well as its fit to the thermal gradients predicted by cooling half-space models (e.g., Sclater et al., 1980) that describe, to first order, the cooling with age of the terrestrial oceanic lithosphere. Figure 1 shows several curves produced from Eq. (1) by varying one or more of the following: surface temperature, mantle temperature, magma source depth, and thermal gradient curvature. The surface temperature gradient for any particular thermal structure is found by calculating \( dT/dz \) as \( z \) approaches zero. For otherwise similar thermal structures, the high surface temperature of Venus allows—but does not require—a shallower minimum magma source depth than on Earth.

3. Composition

The morphological resemblance of the venusian plains, volcanic features and structures to basaltic plains, features and structures on Earth (Campbell et al., 1989; Head et al., 1992), and the Soviet Venera and Vega lander bulk density and elemental abundance data (Surkov et al., 1983, 1984, 1987) imply a basaltic composition for the majority of the surface of Venus. Accordingly, a basaltic (olivine tholeiite) magmatic composition is used here. For ease of comparison, the magmatic compositions and thus the solidus and liquidus of the ascending magma are assumed to be the same on both Venus and Earth. Also, because of a probable dry Venus lithosphere (Kaula, 1990; Donahue and Hodges, 1992), a dry olivine tholeiite (quartz eclogite at high pressures) solidus and liquidus from the experimental data of Lambert and Wyllie (1972) is assumed for a magmatic composition when comparing magmatic cooling curves for Earth and Venus.

4. Plutonic ascent

The pluton ascent model is described in detail in previous publications (Marsh, 1978, 1982; Marsh and Kantha, 1978), and is essentially a low Reynolds number, high Peclet number problem of heat transfer through a thin thermal boundary layer around a sphere. The model is constructed for maximum possible heat transfer,
and so yields an upper bound on magmatic cooling and thus a lower bound on magma ascent velocity. The thermal effects of crystallization during ascent are discussed in Marsh (1978), and are generally small compared to the thermal structure effects in this analysis and thus are not treated here. The model assumes constant velocity buoyant ascent, body-averaged magma temperatures and properties, and an initially crystal-free magma. The resulting plutonic cooling curves, which are dominated by the convective cooling terms and strongly influenced by the planetary thermal structure, are expressed mathematically by:

$$\frac{T}{T_0} = \frac{J}{J + \gamma} \left(1 - e^{-(J + \gamma)t_0}\right) + e^{-\frac{\gamma}{J + \gamma} \left(\frac{\tau}{(J + \gamma)t_0}\right)^n} \cdot \sum_{p=0}^{n} \frac{(-1)^p n! \left[(J + \gamma)t\right]^{n-p}}{(n-p)!}$$

(3)

where $J = 3 N u \kappa / a^2$, $\gamma = \alpha g V / C_p$, $t$ is time, $t_0$ is total ascent time, $N u = 0.8 \ Pe^{1/2}$, $Pe = V a / \kappa$, $Pe$ is the Peclet number, $V$ is the velocity of magmatic ascent, $a$ is the body radius, $\kappa (= 1 \times 10^{-6} \ m^2 \ s^{-1})$ is the thermal diffusivity, $\alpha (= 6 \times 10^{-5} \ K^{-1})$ is the coefficient of thermal expansion, $g$ is the gravitational acceleration, and $C_p (= 1250 \ J \ kg^{-1} \ K^{-1})$ is specific heat capacity. The parameter $T$ is the mean magma temperature, $T_0$ is the magma temperature in the source region, and $n$ is a constant that defines the shape of the planetary thermal gradient (Eq. 1). Equation (3) reduces to the expression for the thermal gradient (Eq. 1) for an infinitely slow ascent (dimensionless ascent time $J t_o = \infty$), and to the adiabatic curve $T / T_0 = e^{-\gamma t}$ for an infinitely fast ascent ($J t_o = 0$). Typical plots of the resulting cooling curves for terrestrial and Venus conditions, contoured in $J t_o$ values, are illustrated in Fig. 2. In order for the magma to reach the surface unsolidified, the cooling curve must not cross the solidus before it reaches the surface. The allowable $J t_o$ values obtained from the cooling
curve plots for Venus and Earth can be directly compared with:

\[
\left(\frac{V_E}{V_V}\right) \left(\frac{a_E}{a_V}\right) = \left(\frac{J_{t0}V}{J_{t0}E}\right) \left(\frac{z_0E}{z_0V}\right)
\]  

(4)

to obtain relative minimum magma ascent velocities, source depths, and body sizes where \( V \) is the minimum magma ascent velocity, \( a \) is the minimum magma body radius, \( z_0 \) is the minimum magma source depth, \( J_{t0} \) is the dimensionless ascent parameter determined from the cooling curves, and the subscripts 'E' and 'V' indicate Earth and Venus values, respectively.

5. Results

Cooling curves for the plutonic ascent mechanism model were calculated for a variety of Venus surface temperature gradients, thermal gradient curvatures, and the variation of surface temperature with elevation on Venus.

In general, for the pluton ascent mechanism model presented here, the influence of the planetary surface temperature dominates the effects of variations in surface temperature gradient, thermal gradient curvature, and mantle temperature. The higher surface temperature of Venus, for otherwise similar planetary thermal structures, allows considerably smaller minimum possible magma body sizes and slower ascent velocities than would be possible on Earth for a reasonable range of source depths and venusian surface thermal gradients (10–25 K/km; Zuber, 1987; Grimm and Solomon, 1988; Zuber and Parmentier, 1990). The results for the pluton ascent model for variations in surface temperature gradient and two different thermal gradient curvatures are shown in Fig. 3. For example, for the same gradient curvature \((n=2)\) and surface thermal gradient as on Earth, a pluton on Venus could have a minimum source depth of about two-thirds of Earth’s, an ascent velocity of about one-third of Earth’s, and a body radius about two-thirds of Earth’s, and still reach the surface unsolidified. The results shown are for a dry basaltic (olivine tholeiite) composition (Lambert and Wyllie, 1972). The surface temperature effect is greater for more primitive magma compositions, but may be smaller for magmas of higher crystallinity. For example, for the case above of the same venusian gradient curvature \((n=2)\) and surface thermal gradient as on Earth, a venusian peridotitic pluton, with solidus and liquidus taken from Takahashi and Kushiro (1983), could have an ascent velocity of about one-fourth (instead of one-third) of Earth’s, and a body radius about three-fifths (instead of two-thirds) of Earth’s, and still reach the surface unsolidified. Normally, peridotite plutons would not be buoyant,
but the example serves as a compositional end member of the enhanced venusian magma transport possibilities.

The thermal and mechanical effects of crystallization have been neglected in this model and eruption may occur as long as the magma reaches the surface unsolidified. A rough estimate of the enhanced venusian magma transport effect of crystallization on eruption probability may be obtained by noting that, for eruptions of basaltic magma below the 55% crystal content "eruptability limit" (Marsh, 1981), the effect is reduced by 5–10%, and for eruptions near the liquidus, 10–20%.

A higher venusian surface thermal gradient (relative to Earth) also enhances magma transport possibilities, but to a much lesser degree than the surface temperature effect. Similarly, for higher values of thermal gradient curvature with depth \((n > 2\) in Eq. 1), the minimum possible as-
cent velocities and body sizes are also considerably less than those of Earth for plutonic ascent. The effects of small variations in surface or mantle temperatures are not significant. For the thermal structures investigated, the effect on the model of a postulated 100 K hotter Venus mantle (Phillips and Malin, 1983; Stevenson et al., 1983; Sotin and Parmentier, 1989) is simply to decrease the minimum possible source depth on Venus by approximately 10% (on average) of the total. The variation of surface temperature with elevation on Venus is approximately 8 K km\(^{-1}\) and decreases with increasing elevation (740–660 K; Seiff, 1983). Figure 4 shows the effect of these temperature variations on magma ascent. The solutions are nearly identical to those for higher mantle temperatures; and slightly greater than the effect of elevated venusian surface thermal gradients.

6. Discussion

For all of the thermal structures considered for plutonic magma ascent in this model, the minimum magmatic ascent velocity, the minimum body size, and the minimum source depth are significantly smaller on Venus than on Earth. This enhanced range of conditions favoring magma transport to the surface is due primarily to the high surface temperature on Venus; elevation-dependent surface temperatures, higher mantle temperatures, and/or higher gradient curvatures have little additional significant effect. Due to the larger difference between the solidus and liquidus temperatures, more primitive magma compositions than the basaltic olivine tholeiite that we assumed in this analysis experience more favorable transport conditions than the silicic compositions. The same $\Delta T$ type of argument indicates that, according to this analysis, more crystalline magmas should reach the surface more easily on Venus than on Earth. This result neglects the (possibly significant) considerations of latent heat of crystallization and considerable viscosity increases due to the addition of crystals in the end stages of cooling. However, considering only the thermal aspects of magma cooling and transport, we might expect more magmas on Venus capable of erupting or reaching the near-surface close to the crystallinity eruption limit (approximately 55% crystals; Marsh, 1981) than is found on Earth. The above-mentioned neglect of most of the thermal effects of crystallization, as well as the initial assumption of a constant velocity of ascent, may make this analysis more useful for considering the transport of magma from the base of the thermal lithosphere to a high-level magma chamber than it is for near-surface (the last 10–20% of the ascent distance) analyses (Marsh and Kantha, 1978). However, as long as the assumptions of constant velocity buoyant ascent and low crystal content are valid, this model also applies to magmatic ascent from the relatively near-surface proposed magma reservoirs in neutral buoyancy zones and/or shallow magma chambers on Venus (Head and Wilson, 1986, 1992). The additional effects of an extensional stress field on magmatic ascent are not considered in this analysis; however, an extensional tectonic regime such as those seen in Beta or Atla Regio (Schaber, 1982; Campbell et al., 1984; Head et al., 1992) would further facilitate magma reaching the surface.

For all venusian thermal structures considered, the ascent of plutons through the venusian lithosphere is significantly facilitated primarily due to the high surface temperature compared to Earth. Thus, our model results would suggest a wider range of sizes of intrusive and/or extrusive volcanic events over a wider spatial distribution of locations. This increased ease of volcanic activity in our model is independent of time. Therefore, our model neither lends support to catastrophic (Schaber et al., 1992) or episodic (Turcotte, 1992) resurfacing mechanisms that have been invoked to explain the nearly spatially random distribution of impact craters (Phillips et al., 1992) on the Venus surface, nor helps distinguish between these models and equilibrium resurfacing models (Phillips et al., 1992). However, our model is consistent with the broad spatial distribution of surface volcanism on Venus.
7. Conclusions

For a wide variety of possible venusian thermal structures, smaller and slower magma bodies are capable of reaching the surface on Venus compared to Earth. This result is a consequence primarily of Venus' high surface temperature. Elevated mantle temperatures, elevation-dependent surface temperatures, and details of the curvature of the near-surface thermal gradient contribute relatively little - if any - significant additional thermal effects on magma cooling and transport. The enhanced ability of magmas to reach the surface may be at least a partial explanation for the broad spatial distribution of surface volcanism on Venus.

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References


