understand which aspects influence on the cooling of ascending magma, we have constructed magma cooling curves for both plutonic and crust buoyant ascent mechanisms, and evaluated the curves for variations in the planetary mantle temperature, thermal gradient curve with depth, surface temperature gradient, and surface temperature. The planetary thermal structure is modeled as

\[
\frac{T}{T_0} = 1 - \tau \left(1 - \frac{Z}{Z_0}\right)^n
\]

(1)

where \(T\) is the temperature, \(T_0\) is the source temperature, \(\tau = 1 - \left(T/T_0\right)\) where \(T_s\) is the planetary surface temperature, \(Z\) is the depth, \(Z_0\) is the source depth, and \(n\) is a constant that controls thermal gradient curvature with depth. Equation (1) is used both for mathematical convenience and flexibility, as well as its fit to the thermal gradients predicted by the cooling half-space models [6]. We assume a constant velocity buoyant ascent, body-averaged magma temperatures and properties, an initially crystal-free magma, and the same liquidus and solidus for both Venus and Earth.

The cooling model for the plutonic ascent has been described in detail in earlier publications [2-5], and is a low Reynolds number, high Peclet number problem of heat transfer through a thin thermal boundary layer around a sphere. The resulting plutonic cooling curves, which are dominated by the convective cooling terms and strongly influenced by the planetary thermal structure, are then expressed mathematically by

\[
\frac{T}{T_0} = \frac{\int_0^\infty e^{-(J+\gamma)n} + e^{-\left[J+\gamma\right]} \left[\frac{r}{(J+\gamma)^n}\right]^{\frac{1}{n}}}{\sum_p (-1)^p \frac{\left[J+\gamma\right]}{(n-p)!}^{n-p}}
\]

(2)

**Fig. 1.** Example of a typical cooling curve plot for platon ascent. The curves are plotted for a given thermal structure and/or planet, and are contoured with their associated dimensionless ascent (\(h_0\)) values. In order for the magma to reach the surface unsoftened, the cooling curve must not cross the solidus. This plot is for a dry olivine tholeiite. The liquidus and solidus are obtained from [11]. Thermal structure parameters: \(n = 2; dT/dZ = 1.0\) (dT/dZ of Earth); \(Z_0 = 1.0\) (\(Z_0\) of Earth).
where \( J = 3 \nu u K / a^2 \), \( \gamma =\gamma g V / C_p \), \( t \) is time, \( t_0 \) is total ascent time, 
\( \nu = 0.8 \nu_0 \left( r / a \right)^2 \), \( \nu_0 = \nu g / a \), \( K \) is the thermal diffusivity, \( n \) is the velocity of magmatic ascent, \( a \) is the body radius, \( K(=1 \times 10^{-4} \text{ m}^2/\text{s}) \) is the thermal diffusivity, \( \alpha = 6 \times 10^{-2} \text{ deg}^{-1} \) is the coefficient of thermal expansion, \( g \) is the gravitational acceleration, and \( C_p(=1.25 \times 10^{4} \text{ ergs Kg}^{-1}\text{K}^{-1}) \) is specific heat capacity. \( T \) 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 (equation (1)). Equation (2) reduces to the expression for the thermal gradient (equation (1)) for an infinitely slow ascent (dimensionless ascent time \( J_0 = \infty \)), and to the adiabatic curve \( T / T_0 = e^{-\gamma t} \) for an infinitely fast ascent (\( J_0 = 0 \)). A typical plot of the resulting cooling curves for terrestrial conditions, contoured in \( J_0 \) values, is illustrated in Fig. 1. In order for the magma to reach the surface unsoftened, the cooling curve must not cross the solidus before it reaches the surface. The allowable \( J_0 \) values obtained from the cooling curve plots for Venus and Earth can be directly compared to obtain relative minimum magma ascent velocities, source depths, and body sizes. The results are shown in Fig. 2.

The cooling model for the buoyant crack ascent has previously been described briefly in [2]. It is the problem of a magma at an initial temperature \( T_0 \) placed in contact with the wall rock of temperature \( T_w \). This problem was initially solved by [7], and their solution for the average temperature \( T \) as a function of time is

\[
\frac{T - T_m}{T_0 - T_m} = 8 \sum_{m=1}^{\infty} \frac{\exp\left(-\frac{(2m-1)^2 \pi^2 K t / 4a^2}{(2m-1)^2}ight)}{\pi^2}
\]

(3)

where the notation is the same as in equation (2), and the right side is constant for any single dimensionless ascent time \( (K t / a^2) \). This result is for a constant wall rock temperature, but can be adapted to a variable wall rock (thermal gradient) temperature by approximating an incremental magma ascent in a simple numerical scheme where the initial magma temperature \( T_0 \) at any location \( m \) is, instead of the source depth temperature, the final magma temperature at the previous location \( m-1 \) [2]. If heat is conducted ahead of the magma body, the boundary temperature of the magma and wall rock will not be constant \( (T_{contact} = T_m) \) at any given location, but will be the average of the two initial temperatures of the magma and wall rock at any location for the majority of the cooling time [8]. For this case, the contact temperature at the mth position is the average of the final temperature of the magma at the \( m-1 \) position \( (T_{m-1}) \) and the initial wall rock temperature at the mth position \( (T_{m0}) \). The preliminary results from this model indicate that the effect of the planetary thermal structure is of the same order of the effect seen in the pluton model.

In general, for both ascent mechanism models presented here, the influence of the planetary thermal structure parameters for Venus in the probable order of decreasing importance is surface temperature, surface temperature gradient, thermal gradient curve with depth, and planetary mantle temperature. The higher surface temperature of Venus, for otherwise similar planetary thermal structures, allows considerably smaller minimum possible crack sizes and/or magma body sizes, and slower ascent velocities than would be possible on Earth for a reasonable range of Venus source depths and surface thermal gradients. This surface temperature effect is greater for more primitive magma compositions, and may be greater for magmas of higher crystallinity. A higher venusian surface thermal gradient has the same effect of the higher surface temperature on magma transport, but to a much lesser degree. Similarly, for higher values of thermal gradient curvature with depth (higher \( n \) in equation (1)), the minimum possible ascent velocity and body/crack size also decreases slightly. If the mantle temperature for Venus is elevated by a hundred degrees or so over that of Earth [9], it should result in a modest increase of melt production and magma transport to the surface compared to Earth. The effect of the range of Venus surface temperatures with elevation \((390^\circ-470^\circ C \text{ or } 660^\circ-740^\circ K)\) is under investigation, and is also anticipated to have a significant effect on magma transport, possibly greater than that of the higher mantle temperature.


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**EVIDENCE FOR RETROGRADE LITHOSPHERIC SUBDUCTION ON VENUS.**

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Though there is no plate tectonics per se on Venus [1], recent Magellan radar images [2] and topographic profiles [3] of the planet suggest the occurrence of the plate tectonic processes of lithospheric subduction [4] and back-arc spreading [5]. The perimeters of several large coronae (e.g., Latona, Artemis, and Ethelinhah) resemble Earth subduction zones in both their planform and topographic profile. McKenzie et al. [4] have compared the planform of accretionary structures in Eastern Tethys with subduction zones of the East Indies. The venusian structures have radii of curvature that are similar to those of terrestrial subduction zones. Moreover, the topography of the venusian ridge/trench structures is highly asymmetric with a ridge on the concave side and a trough on the convex side; Earth subduction zones generally display this same asymmetry.