

NAGW-4263

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Ambient effects on basalt and rhyolite lavas under Venusian, subaerial, and subaqueous conditions

Nathan T. Bridges¹

Department of Geosciences, University of Massachusetts, Amherst

Abstract. Both subaerial and subaqueous environments have been used as analog settings for Venus volcanism. To assess the merits of this, the effects of ambient conditions on the physical properties of lava on Venus, the seafloor, and land on Earth are evaluated. Rhyolites on Venus and on the surface of Earth solidify before basalts do because of their lower eruption temperatures. Rhyolite crust is thinner than basalt crust at times less than about an hour, especially on Venus. At later times, rhyolite crust is thicker because of its lower latent heat relative to basalt. The high pressure on the seafloor and Venus inhibits the exsolution of volatiles in lavas. Vesicularity and bulk density are proportional, so that lavas of the same composition should be more dense on the seafloor and less dense on land. Because viscosity depends partly upon the fraction of unvesiculated water in a melt, basalts with the same initial volatile abundance will be least viscous on the seafloor and most viscous on land. Assuming the same preeruptive H₂O contents, molten rhyolites on Venus will have viscosities ~10% that of rhyolites on land. Despite lower expected viscosities, underwater flows are more buoyant and should have heights like subaerial and Venusian lavas of the same composition and extrusive history. In cases where the influence of crust is insignificant, a volume of rhyolite will have a higher aspect ratio than the same volume of basalt, no matter what the environment. If flow rheology is dominated by the presence of strong crust, aspect ratios differ little among environments or between compositions. These analyses support a rhyolitic interpretation for the composition of Venusian festooned flows and a basaltic interpretation for the composition of Venusian steep-sided domes. Although ambient effects are significant, extrusion rate and eruption history must also be considered to explain analogous volcanic landforms on Earth and Venus.

Introduction

Background

Images of Venus returned by the Magellan spacecraft reveal a plethora of volcanic landforms [Head *et al.*, 1992]. In most cases, the morphology of these features seems consistent with formation by low-viscosity lavas, nominally assumed as basalt [Head *et al.*, 1992]. However, festooned flows [Moore *et al.*, 1992] and over a hundred steep-sided domes [McKenzie *et al.*, 1992; Pavri *et al.*, 1992; Fink *et al.*, 1993] bear a superficial resemblance to structures that on Earth are formed from more viscous, silicic lava. These interpretations are not without controversy. A major problem with a silicic interpretation for the domes is that they are much larger than silicic domes on Earth. Their widths are up to 70 km, and volumes range from ~1-1000 km³ versus 2 km and 10⁻⁵-0.1 km³, respectively, for terrestrial domes [Pike, 1978]. It has been suggested that this is a consequence of the 90 bar pressure of the Venusian atmosphere, which should inhibit the exsolution of volatiles in volcanic rocks [Head and Wilson, 1986]. Large volumes of volatile-rich silicic rock may therefore be preserved at the surface instead of being dispersed in volcanic eruptions, as they are on Earth [Pavri *et al.*, 1992]. However, there is also evidence for a basaltic interpretation. The radar properties of the domes are unlike terrestrial silicic domes [Plaut *et al.*, 1994] and are indistinguishable from the radar properties of the surrounding plains [Ford, 1994]. It has

been suggested that crystal-rich [Sakimoto, 1994; Sakimoto and Zuber, 1995] or vesicular [Pavri *et al.*, 1992] basalts could emulate the behavior of viscous silicic lavas during eruption and therefore may have formed the Venusian domes. It also has been noted that large Venusian domes have dimensions and appearances more similar to basaltic seamounts than silicic domes [Bridges, 1994, 1995; Sakimoto, 1994].

The composition of Venusian volcanoes and their products has broad ramifications. If silicic lavas exist, this implies that they formed by differentiation from a more mafic parent or by melting of silicic crust. Both possibilities have important implications for the composition of the Venusian crust and upper mantle. The formation of large volumes of silicic rock implies a complex crustal history and possibly a water concentration equivalent to that in Earth's mantle [Campbell and Taylor, 1983]. A basaltic composition for the domes and festooned flows would imply that they formed under eruptive or tectonic conditions distinct from those that produced basaltic shield volcanoes and lava plains.

Clearly, interpretations of the composition of domes and other enigmatic volcanic features on Venus are varied and somewhat ambiguous. Fortunately, considerable advances in the last decade have increased our understanding of the factors that control lava flow morphology. This has in part been driven by the desire to know the origin and composition of flows and volcanoes on other planetary bodies based on remote data. On Earth, models can be tested against analogs in the field. This option is, of course, unavailable on other planets at the present time. Therefore, the analysis of extraterrestrial flows has utilized terrestrial and laboratory analogs in order to estimate composition and other parameters. Both land [Pavri *et al.*, 1992] and seafloor [Bridges, 1994, 1995; Sakimoto, 1994] analogs have been used to explain volcanic landforms on Venus. However, critical to the effective use of these analogs is a sufficient

¹Now at Jet Propulsion Laboratory, California Institute of Technology, Pasadena.

Table 1. Planetary Conditions

Environment	Pressure, bars	Ambient Density, kg m ⁻³	Temperature, K	Gravity, m s ⁻²
Venus	90	65	750	8.8
Earth, land	1	1.2	290	9.8
Earth, seafloor	250*	1000	280	9.8

* Common value at the East Pacific Rise

understanding of the effect ambient factors have on flow morphology (Table 1).

Important Ambient Factors

Lava flow cooling is a function of several ambient factors and has received considerable attention in the literature [Head and Wilson, 1986; Fink and Griffiths, 1990; Griffiths and Fink, 1992a, b; Gregg and Greeley, 1993; Manley, 1993; Keszthelyi, 1994]. Cooling rate determines how long the surface of a flow remains molten and the rate of crust growth. Crust thickness will, in turn, exert an important influence on flow morphology [Fink and Griffiths, 1990; Griffiths and Fink, 1993]. Heat loss occurs through a combination of radiative and convective fluxes from the flow surface. Radiative cooling is a function of the temperature contrast between the lava and ambient environment whereas convective cooling also depends upon the ambient density. Griffiths and Fink [1992a,b] showed that for a given lava flow surface temperature, radiative losses are approximately the same on land on Earth (hereafter simply referred to as land) and the seafloor because of similar ambient temperatures. Higher temperatures on Venus cause radiative losses to be slightly less. The high density of water and, to a much lesser extent, the Venusian atmosphere, causes convective losses for underwater and Venusian flows to exceed losses on land [Head and Wilson, 1986; Griffiths and Fink, 1992b]. Considering both radiation and convection, total fluxes are greatest on the seafloor, followed by Venus, and then land. Therefore submarine lavas should cool the fastest and those on land the slowest.

It has also been recognized that volatile exsolution in lavas is inhibited in high-pressure environments like Venus [Head and Wilson, 1986] and the seafloor [Moore, 1965, 1970; MacPherson, 1984]. For eruptions on land, the rise of magma toward the surface causes volatile nucleation into bubbles, followed by expansion [Wilson and Head, 1983; Head and Wilson, 1986]. If magma is sufficiently enriched in volatiles, expansion can be catastrophic, resulting in explosive eruptions. For nonexplosive eruptions, a significant amount of volatiles is lost after leaving the conduit [Cashman et al., 1994]. This occurs as preexisting fluids passively rising through the lava, as new fluids coming out of solution, and as gases released as incompatible phases upon crystallization (often referred to as "second boiling"). Volatile exsolution forms vesicles in the lava, lowering bulk density [Moore, 1965; Moore et al., 1985]. Nonvesiculated volatiles remain dissolved in the magma. In this state, H₂O acts to depolymerize the melt structure, lowering viscosity [McBirney and Murase, 1984; Lange, 1994]. Therefore lavas in high pressure environments are more apt to retain their volatiles and have lower viscosities and higher densities than lavas in low-pressure environments.

Gravity can also be considered an ambient factor because its effect varies among land, the seafloor, and Venus. Combined with lava density and integrated over the volume of a flow, gravity controls the magnitude of the driving force causing lava to spread [Griffiths and Fink, 1993]. On land and

Venus the effective gravitational accelerations are about the same (Table 1). However, on the seafloor, the high density of water (1000 kg m⁻³) is a significant fraction of that of basalt (>2500 kg m⁻³), so that gravity is effectively less. With lava density and other factors being equal, this reduced gravity makes seafloor lavas more buoyant than on land or Venus.

These effects are quantified in this paper. This treatment differs from previous studies in several ways. First, rhyolite, in addition to basalt, is considered. Second, the effect of vesiculation on bulk density is evaluated and its influence on lava flow cooling considered. Third, this and the effects of viscosity and buoyancy are utilized in models of flow aspect ratio. Finally, the effects of ambient conditions on flow and volcano morphology are compared to the effects of extrusion rate and eruption history. The following two sections will explore how flows in the three environments should differ, based on theoretical and, in the case of land and the seafloor, empirical evidence. In the discussion, implications of these differences relative to volcanic features on Venus are evaluated. It will be shown that, as a result of ambient conditions, Venusian rhyolites flow more readily than rhyolites on Earth do. This means that very large volumes of lava, such as the "unusual" volcano with festooned flows near Imdr Regio [Moore et al., 1992], are consistent with a silicic composition. It will also be shown that Venusian basalts closely emulate the behavior of basalts both on land and the seafloor. This strongly suggests that steep-sided domes with morphometries and morphologies very similar to seamounts can be composed of basalt.

Properties of basalt and rhyolite lavas on Venus, land, and the seafloor

In this section, the effects of the ambient environment are evaluated for basalt on Venus, the land, and the seafloor and for rhyolite on Venus and land. Several assumptions are made for these analyses. First, it is assumed that Venusian lavas contain the same amount of preeruptive water and other volatiles as terrestrial lavas do. As the discussion below will show, lava water content can significantly affect lava properties. The preeruptive H₂O content of Venusian lavas depends in large part on the amount of water in the Venusian mantle. The present abundance of water in Venus' atmosphere is very low relative to Earth, but this probably was not the case in the past. The greenhouse effect on Venus strongly suggests that the atmosphere was once wetter than it is today [Shimazu and Urabe, 1968; Rasool and de Bergh, 1970; Chassefiere, 1996]. A terrestrial ocean's worth of water is thought to have been removed by photodissociation and hydrodynamic escape of hydrogen over the last few billion years [Watson et al., 1981; Kasting and Pollack, 1983; Chassefiere, 1996]. Therefore, it is reasonable to assume that abundances of water and other volatiles in the mantles of Venus and Earth are similar. Second, a depth of 2500 m will be used for the seafloor throughout this paper. This is a common depth of mid-ocean spreading centers, the region where most seamounts form [Fornari et al., 1987; Abers et al., 1988;

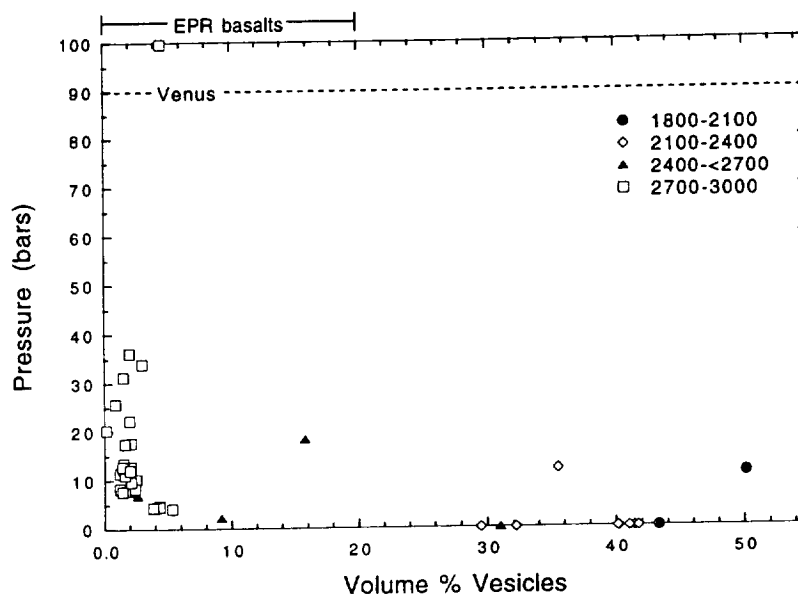


Figure 1. Vesicle contents of Hawaiian submarine basalts as a function of depth [Moore *et al.*, 1985]. Symbols are shown as a function of density (in units of kg m^{-3}). Points plotted at 0 bar are subaerial flows. Bar at top of graph represents the range of porosity values for East Pacific Rise (EPR) basalts, which are generally from a depth of ~ 2500 m (pressure 250 bars) [Batiza and Vanko, 1984]. Horizontal dashed line represents the pressure on Venus.

Smith and Jordan, 1988; Batiza, 1989; Smith, 1991]. Seamounts have been compared to Venusian domes [Bridges, 1994, 1995; Sakimoto, 1994], so a depth of 2500 m is appropriate. Finally, only basalt and rhyolite will be considered. These roughly approximate the two endmember lava compositions found in nature and both have been used as Venusian analogs [e.g., Head *et al.*, 1992; Moore *et al.*, 1992; Pavri *et al.*, 1992; Fink *et al.*, 1993; Bridges, 1995]. Variations in chemical and physical properties of "basalt" and "rhyolite" occur in nature. Nevertheless, differences in ambient conditions should affect flows of slightly different properties in similar ways. In this paper, in which relative differences in flow properties among environments are examined, these variations are not considered in detail.

Vesicle and Volatile Contents

The pressure dependence of vesicularity and volatile contents in basalt is empirically known from samples dredged from varying depths on the seafloor [Moore, 1965; Batiza and Vanko, 1984; Moore *et al.*, 1985] and has been modeled quantitatively [MacPherson, 1984]. Submarine tholeiitic basalts generally contain ~ 0.1 - 0.5 wt % H_2O [Moore, 1965, 1970; MacPherson, 1984; Dixon *et al.*, 1988] and < 0.2 wt % CO_2 [MacPherson, 1984]. The relationship between depth (pressure) and solubility is different for each volatile phase and depends in part on the amount of initial volatiles in the melt [MacPherson, 1984]. However, in general, it has been found that CO_2 exsolves at all ocean depths, H_2O mostly at depths < 500 - 2000 m (~ 50 - 200 bars), and SO_2 mostly at depths < 200 m (~ 20 bars) [Moore, 1970; MacPherson, 1984; Dixon *et al.*, 1988]. To illustrate how pressure-dependent exsolution affects vesicularity, a plot of vesicle content versus depth for Hawaiian basalts is shown in Figure 1. Two families of points are seen, each scattered along one of the axes. Those distributed about the x axis are confined to pressures less than 20 bars and have a wide range of vesicle contents from 1-50 vol %. The second family of points is along the y axis and shows that the Hawaiian lavas at pressures greater than 20 bars have vesicle contents restricted to less than 5 vol %. The

distribution of the two families of points can be explained most simply as due to H_2O and SO_2 exsolution at shallow depths. When ultradeep samples from the East Pacific Rise (EPR) are shown (illustrated by the bar in the upper left of the plot), an apparent conundrum is seen. The EPR is at a depth of 2500 m (~ 250 bars), where very low vesicle contents might be expected. However, most of the scatter in these deeper samples is easily attributable to different preeruptive volatile contents [Batiza and Vanko, 1984; MacPherson, 1984], exsolution in the magma chamber [MacPherson, 1984], or incomplete CO_2 degassing before solidification [Dixon *et al.*, 1988]. The latter two factors depend upon the dynamics of the magma chamber and eruption process and cannot be easily modeled. If these are neglected, vesiculation is then strictly a function of initial volatile content and confining pressure [MacPherson, 1984]. If the same preeruptive volatile content is assumed, then a plot such as Figure 1 would show a curve that is steep at high pressures and shallow at low pressures. For the comparative purposes here, the same initial volatile contents of magmas on Venus, the seafloor, and land are assumed. The different pressures in the three environments will result in differential vesiculation upon eruption. Seafloor lavas will be vesicle-poor and volatile-rich, subaerial lavas the reverse, and Venusian lavas in between.

For the purposes of the comparative analyses in this paper, a standard vesicle content of 30 vol % on land is used to compare to Venus and the seafloor (Table 2). This vesicularity

Table 2. Dissolved Water and Vesicularity

Environment and Lava Type	H_2O , wt %	Vesicles, vol %
Venus basalt	0.3	10
Venus rhyolite	4	10
Land basalt	0.05	30
Land rhyolite	3	30
Seafloor basalt	0.5	5

is similar to that measured for basalt collected in a molten state at intermediate vent distances [Cashman *et al.*, 1994] and for naturally emplaced samples [Moore *et al.*, 1985]. Although 30% is common, values can be greater than 40% [Moore *et al.*, 1985]. Basalts at the 90 bar pressure of Venus will have slightly more vesicles than seafloor basalts due to exsolution of CO₂ and perhaps some H₂O. In order to extrapolate from land values to pressures on Venus and the seafloor, the same limited range of preeruptive volatile content of the lavas must be assumed. MacPherson [1984] derived theoretical curves of pressure versus vol % vesicles for various initial amounts of H₂O and CO₂. Using his curves for preeruptive CO₂ and H₂O contents of 0.2 and 0.5-1.0 wt %, respectively, basalts at Venusian (90 bars) and EPR (250 bars) pressures are predicted to have ~10 and 5 vol % vesicles, respectively [MacPherson, 1984, Figure 2]. The negligible effect of H₂O on vesiculation at these pressures relative to that of CO₂ lets a range of preeruptive water contents be assumed as opposed to specific values. Only in the unlikely case in which the preeruptive H₂O content of basalt exceeds ~1 wt % will water act to increase vesiculation over the amounts considered here.

Volatiles not exsolved will remain in the melt. Of these, water is of primary interest due to its effect on viscosity (see below). Tholeiitic basalts on land commonly have H₂O contents of the order of 0.05 wt % [Peck *et al.*, 1966], whereas those on the seafloor range from about 0.1 to 0.5 wt % [Moore, 1965, 1970; MacPherson, 1984; Dixon *et al.*, 1988]. Unfortunately, the solubility of H₂O as a function of pressure at low pressures is poorly known, making it difficult to predict the amount of water retained in a melt as a function of pressure and initial volatile content. To get around this, MacPherson [1984] extrapolated a function from 0.05 wt % H₂O at 1 bar to a known solubility at 1000 bars that followed a linear relationship between mole fraction H₂O and the square root of pressure, a functional dependence found at pressures greater than 1000 bars [Hamilton *et al.*, 1964; Khitarov and Kadik, 1973]:

$$\text{H}_2\text{O wt \%} = 0.0889p^{1/2} - 0.039 \quad (1a)$$

The shape of such a curve reproduces, to a certain extent, the greater degree of exsolution observed at pressures less than ~50-200 bars, but it also predicts too much water for seafloor lavas. A better approximation is a function that intercepts 0.05 wt % at 1 bar and 0.1-0.5 wt % at 250 bars. Here, 0.5 wt

% H₂O is used for the seafloor to provide consistency with the discussion of vesiculation above. The resulting function is

$$\text{H}_2\text{O wt \%} = 0.0304p^{1/2} + 0.019 \quad (1b)$$

Using equation (1b), a lava with 0.05 and 0.5 wt % H₂O on land and the seafloor, respectively, would have ~0.3 wt % H₂O at the 90 bar pressure of Venus (Table 2).

Rhyolite is not known to erupt on the seafloor, so empirical measurements of vesicularity needed to extrapolate to Venusian conditions are lacking. For the purposes of doing the comparative studies that are the focus of this paper, it is assumed that subaerial and Venusian rhyolite contain 30% and 10% vesicles, respectively, just as basalt does. The range of water contents in rhyolites is much larger than that in basalt. Common values are 2-4 wt%, although they can be much greater or less depending on the mode of petrogenesis [Whitney, 1988]. Laboratory experiments on volatile solubility in rhyolites have generally been done at the kilobar level. Therefore it is necessary to extrapolate to lower pressures to estimate the difference in water contents between rhyolites on land and Venus. When this is done, it is found that granitic melt loses ~1 wt % H₂O for a 100 bar (0.01 GPa) decrease in pressure at low pressures [Burnham, 1979; Philpotts, 1990, p. 194]. Therefore, a typical subaerial rhyolite with 3 wt % H₂O would have about 4 wt % H₂O on Venus (Table 2).

Density

The average bulk density of basalt is greater at depth on the seafloor than it is near the surface or on land. This is illustrated in Figure 1, where Hawaiian basalts erupted at pressures greater than 20 bars are seen to have densities greater than 2700 kg m⁻³. At pressures less than 20 bars, density is variable and can be as low as 1850 kg m⁻³. The density dependence on depth is primarily due to the decrease in vesiculation with increasing pressure [Moore *et al.*, 1985]. Bulk density (ρ) is related to the solid density (ρ_s), and vesicle fraction x by

$$\rho = \rho_s(1-x) \quad (2)$$

This is easily shown in a plot of vesicularity versus bulk density for 61 samples of Hawaiian basalt measured by Robertson and Peck [1974] (Figure 2). Taking the vesicle

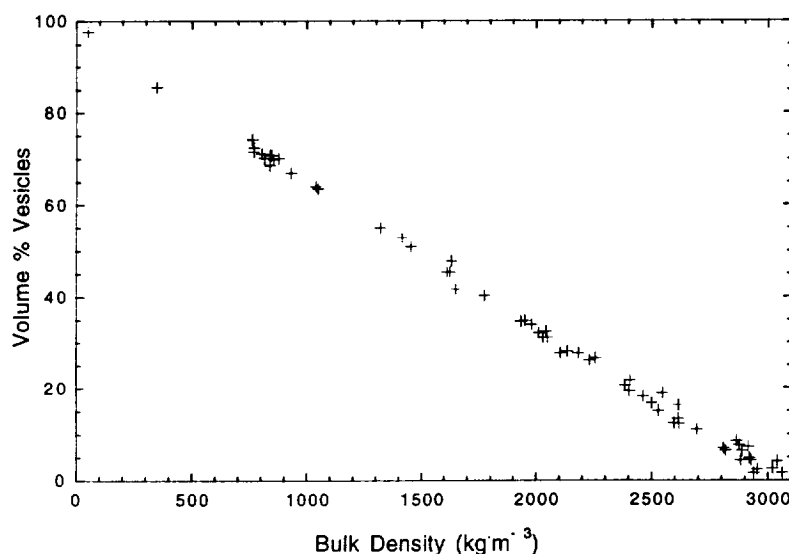


Figure 2. Volume percent vesicles versus bulk density of Hawaiian basalt (data from Robertson and Peck [1974]).

Table 3. Driving Forces in Planetary Environments

Environment and Lava Type	Lava Bulk Density*, kg m ⁻³	Reduced Gravity, m s ⁻²	Driving Force/Unit Volume, N m ⁻³
Venus basalt	2700	8.6 (8.5)	2.3x10 ⁴ (1.8x10 ⁴)
Venus rhyolite	1980	8.5 (8.4)	1.7x10 ⁴ (1.3x10 ⁴)
Land basalt	2100	9.8	2.1x10 ⁴
Land rhyolite	1540	9.8	1.5x10 ⁴
Seafloor basalt	2850	6.4 (5.1)	1.8x10 ⁴ (1.1x10 ⁴)

Values in parentheses assume land densities. Seafloor values are for a depth of 2500 m, typical of the East Pacific Rise.

* Bulk densities assume a solid density for basalt and rhyolite of 3000 and 2200 kg m⁻³, respectively.

contents discussed above and assuming a solid density of 3000 kg m⁻³, basalts on the seafloor, Venus, and land should have bulk densities of 2850, 2700, and 2100 kg m⁻³, respectively (Table 3). Similarly, with a solid density of 2200 kg m⁻³ [Murase and McBirney, 1973], the bulk densities of rhyolite on Venus and land are 1980 and 1540 kg m⁻³, respectively. Therefore, lavas on Venus and at 2500 m on the seafloor will be denser than those on land by a factor of ~1.3 and 1.4, respectively. For the remainder of this paper, bulk density is assumed wherever "density" is discussed.

Thermal Parameters

The effective thermal conductivity (k) in vesicular rock depends both upon the conductivity of the rock itself and radiative and convective heat transfer through the gas in the vesicles [Robertson and Peck, 1974; Keszthelyi, 1994]. The heat transfer through the gas is relatively inefficient compared to that in the rock, so that the effective conductivity of the bulk rock decreases as vesiculation increases. The lower vesicularities predicted on Venus and the seafloor relative to land will cause conductivities to be higher. Using values for basalt at 1423 K [Keszthelyi, 1994, Figure 4], conductivities are predicted to be ~1.3 and 1.5 times greater on Venus (10% vesicles) and the seafloor (5%), respectively, compared to land (30%). Rhyolite conductivities are approximately the same as those for basalt [Robertson, 1988] and should show a similar dependence on vesicularity.

A more useful thermal parameter is the thermal diffusivity, κ , which controls the rate at which temperature changes in a conductively cooling body. It depends on the conductivity, density, and specific heat (c) of the rock:

$$\kappa = \frac{k}{\rho c} \quad (3)$$

Specific heat is per unit mass and therefore is not a function of vesicularity. Detailed physical modeling by Keszthelyi [1994] shows that the differences in diffusivity for basalts at the vesicularities and temperatures of interest here are insignificant. The increase in conductivity and density with decreasing vesicularity are roughly the same, resulting in negligible effects on diffusivity. Therefore, diffusivities of basalt and rhyolite used in the subsequent thermal analyses are assumed to be the same in the three environments and are taken as 6 and 7.5 x 10⁻⁷ m² s⁻¹, respectively [Peck et al., 1977, Williams and McBirney, 1979]. Specific heat is taken as 1200 J kg⁻¹ K⁻¹, a common value for both felsic and mafic rocks [Robertson, 1988].

Lava Flow Cooling

Basalt and rhyolite typically erupt at temperatures (T_i) of ~1423 K and 1073 K, respectively [Williams and McBirney, 1979]. Initial cooling is very rapid and can be modeled by assuming that the lavas behave as semi-infinite half-spaces cooled by a combined convective and radiative flux (F) [Griffiths and Fink, 1992a,b]. The temperature (T) is a function of depth beneath the surface of the flow (z) and time (t) such that

$$T(z,t) = T_i - \int_0^t \frac{F(\lambda) e^{-z^2/4\kappa(t-\lambda)}}{\rho c \sqrt{\kappa\pi(t-\lambda)}} d\lambda \quad (4)$$

where λ is a dummy variable used in integration that has units of time and ranges from 0 (initial time) to t (final time). Radiative and convective values of F are taken from Griffiths and Fink [1992a,b]. F depends on ambient properties and iteratively on the surface temperature of the flow. It is not directly dependent on the physical properties of the lava. Thermal conductivity is encapsulated within the thermal diffusivity term (κ), where as discussed above, its proportionality with density nullifies its effect. Density, independent of diffusivity, is the only lava physical property in (4) that varies in the three environments. As the structure of the equation shows, density is inversely proportional to the change in temperature with time and depth. In other words, the ability for a flow to retain heat is proportional to density. For a given cooling flux F , the increased density of Venusian and seafloor lava inhibits cooling by a factor of ~1.3 and 1.4, respectively, relative to flows of the same composition having subaerial densities.

Lava surface temperatures as a function of time are plotted in Figure 3. Rhyolite and basalt are shown in all three environments, except for the improbable case of seafloor rhyolite. The solidification temperatures are taken as those at which the outer surfaces of the lavas quench [Griffiths and Fink, 1992a,b]. This is the glass transition, and because chilling involves supercooling, it is below the solidus. Lavas at the glass transition temperature have very high viscosities (~10¹² Pa s) and, as a result, exhibit some solid-like properties, including brittle failure (see viscosity section below) [Bottinga, 1994; Bottinga et al., 1995; McMillan and Wolf, 1995; Richet and Bottinga, 1995]. The glass transition temperatures of basalt and rhyolite are taken as 1003 K [Ryan and Sammis, 1981] and 943 K [Westrich et al., 1988], although it is recognized that other estimates exist (e.g., 1125

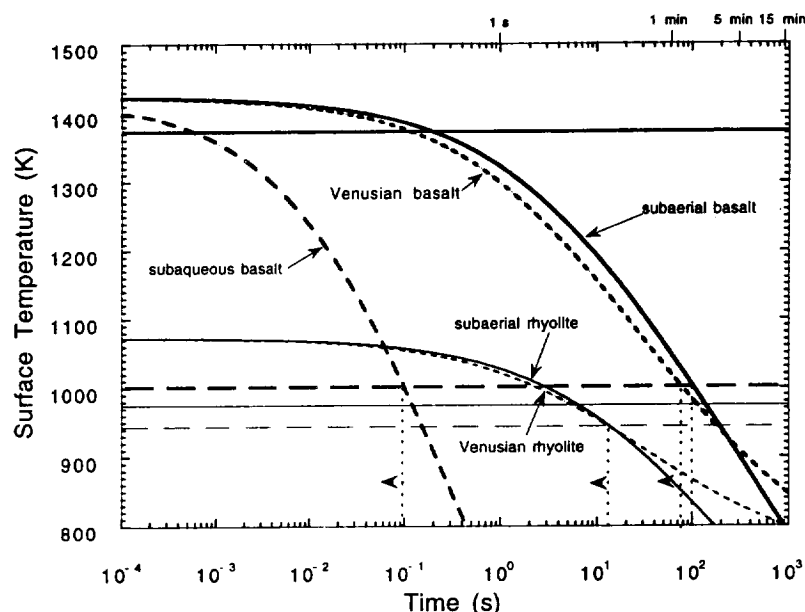


Figure 3. Lava surface temperature as a function of time for basalts and rhyolites on Venus, the land, and the seafloor. Heavy horizontal lines are the solidus (solid) and glass transition (dashed) temperatures for basalt. Light horizontal lines are for rhyolite. Solidification occurs when the temperature reaches the glass transition, marked by vertical dashed lines. Arrows show when solidification would occur if subaerial lava properties were used in place of those expected on Venus or the seafloor.

K for rhyolite by *Neuville et al.*, [1993]). The addition of H_2O lowers the glass transition temperature of synthetic melts substantially [*Dingwell*, 1995]. However, there are no available measurements of its effect on natural melts for the range of water contents of interest here. Therefore glass transition temperatures are assumed to be the same for a given composition, regardless of H_2O content.

Subaerial basalt and rhyolite take 100 and 14 s, respectively, to reach their glass transition temperatures following eruption. Venusian basalt and rhyolite solidify in 75 and 13 s, respectively. Submarine basalt solidifies in 0.1 s. If we instead assume that Venusian and submarine lavas have densities like their subaerial counterparts, cooling times decrease slightly. Cooling times are reduced by 20 and 5 s, respectively, for Venusian basalt and rhyolite and 0.04 s for submarine basalt. These minor differences are illustrated by the horizontal arrows in Figure 3. Thus, cooling times increase slightly when adjusted to reflect the different lava densities on Venus and the seafloor. However, the adjustments are relatively minor compared to the differences in cooling times among the flows in the three environments and do not change the solidification sequence. Submarine basalts solidify the fastest because of the rapid convective heat loss. Rhyolites on Venus and land quench before basalts do because they erupt at a lower temperature. The slightly greater cooling flux due to convection on Venus causes basalts and rhyolites there to solidify a bit faster than their compositional counterparts on land. The relative magnitude of the basalt cooling times is similar to that found by *Griffiths and Fink* [1992a,b], although the precise values are different because of the different densities and diffusivities used.

Once the lava reaches the glass transition temperature, it starts to develop a glassy crust that thickens a distance z as a function of time t (equation (4)) [*Griffiths and Fink*, 1992a,b]. However, this only applies for times less than about 1000 s, when crystallization is minimal [*Uhlmann et al.*, 1979; *Griffiths and Fink*, 1992a,b]. After about 10^4 s, or a few hours, crystallization can be significant, and latent heat must be considered [*Lipman et al.*, 1985; *Crisp et al.*, 1994]. For

example, *Crisp et al.* [1994] found that the volume percent of crystals in Mauna Loa basalts increased from < 10% in the first few hours of eruption to up to 30% at later times. The thickness, z_c , of a crystallized layer in a cooling lava flow is [*Turcotte and Schubert*, 1982, p. 169]

$$z_c = 2\lambda_1 \sqrt{\kappa t} \quad (5a)$$

The symbol λ_1 is a constant and is found iteratively using (5a) and the following equation:

$$\frac{e^{-\lambda_1^2}}{\lambda_1 \operatorname{erf} \lambda_1} = \frac{L\sqrt{\pi}}{c(T_m - T_0)} \quad (5b)$$

where L is the latent heat of crystallization, T_m the solidification temperature, and T_0 the temperature at the top of the flow. Here, the latent heats of basalt and rhyolite are taken as 4.2×10^5 and 2.9×10^5 J kg $^{-1}$ [*Huppert and Sparks*, 1988]. The solidus is used for the solidification temperature and for basalt is taken as 1373 K (1100°C) [*Griffiths and Fink*, 1992a,b]. For rhyolite, 973 K (700°C) is assumed, a common value for silicic magmas [*Piwocki*, 1968; *Best*, 1982]. The ambient temperature (T_a , Table 1) is used for T_0 because at times of 10^4 s and more, the lava surface will have cooled close to this value.

Crust thickness as a function of time since eruption is shown in Figure 4. For times before 1000 s, quenching is assumed and the thickness is that of glass (equation (4)). At 1000 s, the depth of the solidus isotherm is of the order of 1 cm deeper than that of the glassy crust. Crystallization should begin in earnest by about 10^4 s. After this time, thickness is computed from equation (5). At times between 1000 and 10^4 s, the amount of glass will be decreasing and crystals increasing. Therefore the curves for glass from equation (4) at times before 1000 s have been manually connected in Figure 4 to those for crystallization at times after 10^4 s. Although schematic, this illustrates the transition from the production of glassy crust to that of crystal-rich crust with latent heat at times of a few thousand seconds and depths of a few centimeters.

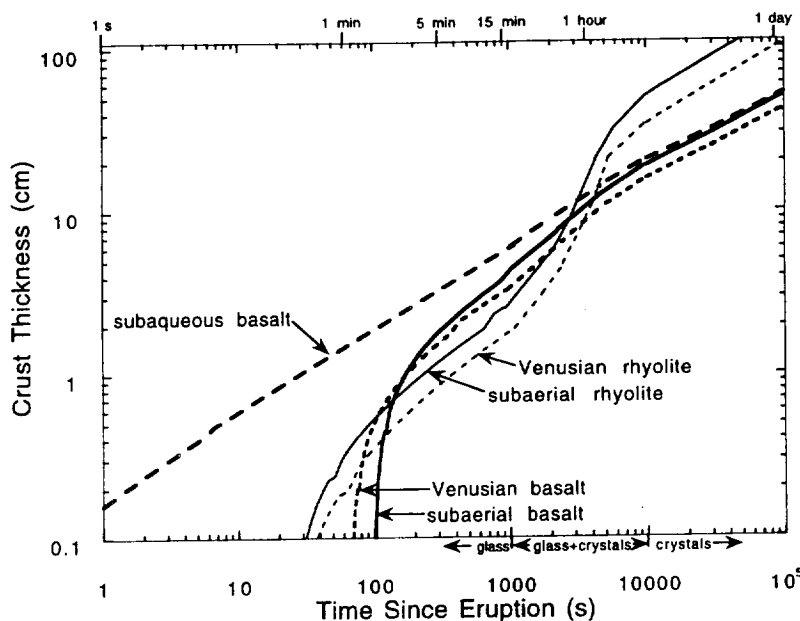


Figure 4. Thickness of crust as a function of time since eruption for basalt and rhyolite on Venus, the land, and the seafloor. Solidification begins once the glass transition temperature is reached. Before 1000 s, thickness is equivalent to the depth of the glass transition isotherm assuming cooling of a semi-infinite half space (equation (4)). After 10^4 s, thickness is that of the solidus isotherm assuming latent heat of crystallization (equation (5)). Half-space and latent heat curves are manually connected between 1000 and 10^4 s. Early thickening is strongly controlled by radiative and convective fluxes from the flow surface. At times after several hundred seconds, represented by the shallowing of the curves, crust thickness is more dependent on the temperature contrast between the interior and exterior of the flow and is roughly proportional to the square root of time. The change to cooling by latent heat after 10^4 s causes rhyolite crusts to ultimately be thicker than basaltic crusts.

Crust thicknesses among flows in the three environments change relative to each other with time (Figure 4). Due to rapid convective heat loss, submarine basalt has the thickest crust at times less than about 50 min. The higher convective heat losses on Venus compared to land cause Venusian basalt to solidify faster than subaerial basalt, producing thicker crust at early times. This quickly changes as the effect of the high Venusian surface temperature acts to slow growth, so that subaerial basalt crust is ultimately thicker (for an earlier treatment of this, taking into account only latent heat for basalt, see *Head and Wilson* [1986]). A similar relationship is seen for subaerial and Venusian rhyolite flows. Before about 50 minutes, rhyolite's low eruption temperature relative to basalt causes it to have a thinner crust at all but the earliest times. However, at later times crystallization is significant, causing rhyolites to have thicker crusts because of their lower latent heat relative to basalt. Thus, for times less than about an hour, submarine basalts should have the thickest crusts, followed by subaerial basalt, Venusian basalt, subaerial rhyolite, and Venusian rhyolite. However, at later times, rhyolites should have the thickest crusts.

Viscosity

Lava viscosity is mainly dependent upon composition, temperature, pressure, abundance and size of crystals, vesicularity, and dissolved volatile content [*Murase and McBirney*, 1973; *McBirney and Murase*, 1984]. In silicate melts, viscosity is proportional to the number of bridging oxygens per tetrahedrally-coordinated cation [*Mysen*, 1990]. The viscosity of basaltic melt, with few bridging oxygens, typically ranges from hundreds to thousands of Pa s whereas rhyolitic melt, which is more polymerized, is several orders of magnitude greater [*Shaw*, 1963; *McBirney and Murase*, 1984; *Lange*, 1994]. Adding water to silicate melts lowers the

polymerization of SiO_4 tetrahedra, reducing viscosity. The extent of this reduction depends on the degree of polymerization before H_2O is added [*McBirney and Murase*, 1984]. Rhyolitic melts, being more polymerized than basaltic melts, have viscosities that are more strongly affected by water content.

Tholeiitic lavas on Venus and the seafloor are both at pressures sufficient to inhibit some H_2O exsolution (see vesiculation discussion above) and should therefore have lower viscosities than those on land. Taking the relationship between log viscosity and wt % H_2O for 1423 K (1200°C) basalt in Figure 4 of *McBirney and Murase* [1984; after *Khitarov and Lebedev*, 1978] and extrapolating the curve to low water contents, it is found that adding a few tenths of a weight percent of H_2O to a nearly dry melt at ~1000 Pa s lowers its viscosity by a few hundred Pa s. Unfortunately, ambiguity in how to extrapolate the curve and in the precise shape of the curve itself prevents an accurate determination of viscosity for the solubilities used here for Venus (0.3 wt % H_2O) and the seafloor (0.5 wt % H_2O). Rather, it is assumed that molten basalt viscosities on land, Venus, and the seafloor are 1000, 800, and 700 Pa s, respectively. These are correct in a relative sense and are probably accurate to within a few tens of percent.

As discussed above, the pressure dependence of volatile exsolution should cause rhyolites on Venus to have ~1 wt% more H_2O than those on land. For 1073 K rhyolite, going from 3 wt % H_2O (land) to 4 wt % H_2O (Venus) reduces viscosity from $\sim 10^8$ to 3×10^7 Pa s (800°C curve in Figure 5 of *Shaw* [1963]). However, the log viscosity versus wt % H_2O curve used to derive these numbers is very generalized, so such exact values are relatively meaningless. Therefore, for the sake of simplicity, it will be assumed that molten rhyolites on land have a viscosity of 10^8 Pa s and those on Venus 10^7 Pa s (Table 4). This is consistent with other experimental data showing that the addition of 1 wt % H_2O to silicic melt reduces

Table 4. Rheologic Parameters in Planetary Environments

Environment and Lava Type	η , Pa s	η_c , Pa s	σ , Pa	σ_c , Pa	E , $m^{-1} s^{-1}$	E_c , $m^{-1} s^{-1}$	Γ , m^{-1}	Γ_c , m^{-1}
Venus basalt	800	10^{12}	1000	10^8	29 (18)	$2.3 \times 10^{-8} (1.8 \times 10^{-8})$	2.3 (1.8)	$2.3 \times 10^{-4} (1.8 \times 10^{-4})$
Venus rhyolite	10^7	10^{12}	10^7	10^8	$1.7 \times 10^{-3} (1.3 \times 10^{-4})$	$1.7 \times 10^{-8} (1.3 \times 10^{-8})$	$1.7 \times 10^{-3} (1.3 \times 10^{-3})$	$1.7 \times 10^{-4} (1.3 \times 10^{-4})$
Land basalt	1000	10^{12}	1000	10^8	21	2.1×10^{-8}	2.1	2.1×10^{-4}
Land rhyolite	10^8	10^{12}	10^7	10^8	1.5×10^{-4}	1.5×10^{-8}	1.5×10^{-3}	1.5×10^{-4}
Seafloor basalt	700	10^{12}	1000	10^8	26 (11)	$1.8 \times 10^{-8} (1.1 \times 10^{-8})$	1.8 (1.1)	$1.8 \times 10^{-4} (1.1 \times 10^{-4})$

Here, η , viscosity; η_c , crust viscosity; σ , yield strength; σ_c , crust yield strength; E , driving force per unit volume + viscosity; E_c , driving force per unit volume + crust viscosity; Γ , driving force per unit volume + yield strength; and Γ_c , driving force per unit volume + crust yield strength. Values in parentheses assume land densities, viscosities, and yield strengths.

viscosity by about an order of magnitude [McBirney and Murase, 1984; Lange, 1994]. Thus basalt viscosities are very similar on Venus and Earth whereas rhyolite viscosities are very different, resulting in different viscosity contrasts between basalt and rhyolite on the two planets. The viscosity of rhyolite exceeds that of basalt by a factor of $\sim 10^4$ on Venus versus a factor of $\sim 10^5$ on land.

In some cases, vesicularity can change viscosity. The effect is relatively minor in mafic magmas but can be significant in silicic ones [Cas and Wright, 1987]. At low strain rates, increasing the vesicularity of rhyolite from 10% (Venus) to 30% (land) lowers its viscosity by 40-50% [Bagdassarov and Dingwell, 1992]. In this case, the viscosity contrast between rhyolites on Venus and land is reduced from an order of magnitude to about half an order of magnitude. In contrast, at high strain rates, increasing the vesicularity of GeO_2 melts by 20% is predicted to increase the viscosity by a factor of 2.6 (extrapolation of data in Figure 8 of Stein and Spera [1992]). In both examples, viscosities of rhyolites on Venus are still expected to be lower than on Earth, but by how much depends on the model chosen. Because of this, it is assumed in the forthcoming analyses and discussion that vesicularity does not affect viscosity.

The viscosities discussed above apply to the early stages of eruption, when cooling and crystallization are relatively insignificant. At later times, as lava cools, its viscosity increases and the influence of crust must be considered. For example, in subaerial Hawaiian basalt eruptions, bulk viscosity increases 1-3 orders of magnitude over tens of hours [Fink and Zimbelman, 1990] and can be as high as 10^7 Pa s 25 km from the vent [Moore, 1987]. The viscosity of the glassy crust itself is much higher than that of the molten or bulk flow. As measured in the laboratory, the glass transition temperature is approximately coincident with a viscosity of 10^{12} Pa s for a range of silicate compositions [Bottinga, 1994; Bottinga et al., 1995; McMillan and Wolf, 1995; Richet and Bottinga, 1995]. Therefore this value will be used for the viscosity of glassy crusts of basalts and rhyolites on Venus, land, and the seafloor (Table 4). This viscosity value is only relevant for a limited amount of time because as the outer parts of the glassy crust cool below the glass transition, the viscosity will increase in an Arrhenian (exponential) fashion [Bottinga et al., 1995]. Furthermore, after about 10^4 s, the influence of glassy crust will wane and that of crystal-rich crust will increase (Figure 4). However, the drop in temperature as glass cools and crystallization commences will increase the yield strength of the lava [Murase and McBirney, 1973; McBirney and Murase, 1984] (see below), so that unless applied stresses are very high, viscous deformation will cease. Therefore at times greater than about 10^4 s flow morphology will probably be more affected by the magnitude of the yield strength, and the consideration of even higher crust viscosities is probably not necessary.

Yield Strength

Many lavas possess a yield strength that must be exceeded for deformation and flow to occur. Like viscosity, yield strength is inversely proportional to temperature and crystal content [McBirney and Murase, 1984] and may be proportional to silica content [Hulme, 1974; Moore et al., 1978]. However, the relative effects of composition versus those of temperature and crystallinity are not completely understood [McBirney and Murase, 1984] and there is evidence that the apparent yield strengths of some lavas are due in part to the influence of a solidified breccia blanket [Fink and Zimbelman, 1990] or carapace [Griffiths and Fink, 1993]. Based on laboratory measurements, the yield strength of both Mauna Loa and Columbia River basalts at 1373 K is about 5000 Pa [McBirney and Murase, 1984], whereas that for Mount Etna basalts at subliquidus temperatures is no more than 100 Pa [Pinkerton and Norton, 1995]. From flow dimensions, Moore et al. [1978] estimated yield strengths of 100 - 10^4 Pa for lavas with less than 50% SiO_2 and Fink and Zimbelman [1990] judged the yield strength of freshly erupted Hawaiian basalt to be ~ 1000 Pa, increasing to about 10^4 Pa as the eruption progressed. It is clear that a range of values may represent a "common" yield strength for basalt. Here, 1000 Pa is assumed. Based on the trend of yield strength versus SiO_2 , the yield strengths of rhyolites are of the order of 10^7 Pa [Hulme, 1974; Moore et al., 1978; McBirney and Murase, 1984], a value that will be used here. The yield strengths for lava flows with solid crusts are assumed to be 10^8 Pa, a value determined by Griffiths and Fink [1993] that seems compositionally invariant.

Lava Flow Shapes

This section examines how the different densities and rheologies in the three environments affect lava flow shapes. A radially extruding lava flow on a flat surface in a gravity field will spread due to the buoyancy driving force equal to [Griffiths and Fink, 1993]

$$\rho g H^2 R \quad (6a)$$

The driving force per unit length L of a flow extruded from a line source, such as a rift, is

$$\rho g H^2 \quad (6b)$$

where g' is the reduced gravity, H is the height at the center of the flow (elevation difference between the flow top and underlying flat surface), R is the flow radius, and ρ is the lava bulk density [Griffiths and Fink, 1993]. The reduced gravity, or buoyancy, is expressed as

$$g' = g \left(\frac{\rho - \rho_a}{\rho} \right), \quad (7)$$

where g is gravity and ρ_a is the density of the ambient fluid. The value of g' is significant for submarine flows. The density of water (1000 kg m^{-3}) causes a submarine flow to have about 1/3 less effective gravity than it would on land.

The analyses in this paper assume lava flows are extruded on to a flat surface. However, many flows are erupted on the flanks of volcanoes or on sloping plains. In these cases, the driving force in the downslope direction is increased by an amount proportional to $\sin\theta$, where θ is the topographic gradient. However, for a given slope the differences in the relative magnitudes of driving forces, and therefore, flow morphology, between compositions and among environments will be the same as they would be on a flat surface. Therefore slope is not explicitly considered in the analyses and discussions below.

The driving forces per unit volume ($\rho g'$) for the model basalts and rhyolites used in this paper are given in Table 3. Driving forces on Venus and the seafloor are derived using the lava densities appropriate for those environments. The values in parentheses show what the driving forces would be using land densities. On Venus the higher lava densities are somewhat offset by the lower gravity, resulting in driving forces that are similar to those on land. If a subaerial lava density is used, then driving forces are slightly lower than on land. Buoyancy on the seafloor offsets the high lava density expected in that environment, resulting in driving forces that are about 15-20% less than on Venus or land. If a subaerial lava density is used, then driving forces are only 1/2 the value on land and Venus. Thus the effects of high lava density and reduced gravity (buoyancy) on Venus and the seafloor work against each other, resulting in driving forces that are close to those on land.

Flow shape partly depends upon the balance between the driving force and a retarding force. *Griffiths and Fink* [1993] derived equations relating radius or length and height of flows for five dynamical regimes that incorporate these forces. Four describe the buoyancy driving force balanced by (1) Newtonian shear stresses in the flow interior (regime buoyancy-Newtonian-interior (BNI)), (2) yield stresses in the flow interior (buoyancy-plastic-interior (BPI)), (3) Newtonian shear stresses in the crust (buoyancy-Newtonian-crust (BNC)), and (4) yield stresses in the crust (buoyancy-crust-strength (BCS)). The fifth is a regime in which overpressure from the lava extrusion is dominant (pressure-crust-strength (PCS)) and is resisted by a brittle shell [*Iverson*, 1990]. The other regimes do not consider extrusion pressure. However, this factor will not systematically vary in the three environments. Therefore, for the purposes of evaluating relative differences of lavas on Venus, the seafloor, and land, it can be effectively ignored.

Aspect ratio (A) is a more useful parameter than height or radius alone for characterizing gross shape and is commonly used in comparing flows and constructs on different planets, where detailed morphometric data are generally lacking [e.g., *Fink et al.*, 1993; *Bridges*, 1995]. An understanding of the factors that control aspect ratio is therefore important in the study of planetary volcanology. Aspect ratio is defined as:

$$A = \frac{H}{2R} \quad (8a)$$

[from a point source) and

$$A = \frac{H}{L} \quad (8b)$$

[from line source). Equation (8a) applies to radially emanating

flows and (8b) to nonsymmetrical lobate and digitate flows. By assuming that the flows erupt at a constant rate and converting time to volume + volumetric effusion rate ($t = V/Q$), the equations of *Griffiths and Fink* [1993] can be rearranged to give aspect ratio as a function of volume (V). The rearranged equations are as follows.

Regime 1 (BNI)

$$A = \frac{1}{2} \left(\frac{Q^3}{V^4 E^3} \right)^{1/8} \quad (9a)$$

Regime 2 (BPI)

$$A = \frac{1}{2} \left(\frac{1}{V \Gamma^3} \right)^{1/5} \quad (9b)$$

Regime 3 (BNC)

$$A = \frac{1}{2} \left(\frac{\kappa^3 Q^3}{V^7 E_c^6} \right)^{1/14} \quad (9c)$$

Regime 4 (BCS)

$$A = \frac{1}{2} \left(\frac{\kappa^3}{Q^3 V \Gamma_c^6} \right)^{1/8} \quad (9d)$$

Regime 5 (PCS)

$$A = \frac{1}{2} \quad (9e)$$

E and Γ are the driving forces per unit volume divided by the viscosity and yield strength, respectively:

$$E = \frac{\rho g'}{\eta} \quad E_c = \frac{\rho g'}{\eta_c} \quad (10a)$$

$$\Gamma = \frac{\rho g'}{\sigma} \quad \Gamma_c = \frac{\rho g'}{\sigma_c} \quad (10b)$$

The subscript c refers to parameters of the outer lava crust. Complete derivations of the original equations are given by *Griffiths and Fink* [1993]. Appropriate values of E , E_c , Γ , and Γ_c , for basalt and rhyolite in the three environments are given in Table 4.

Density (ρ), viscosity (η , η_c), yield strength (σ , σ_c), buoyancy (g'), and thermal diffusivity (κ) are the parameters that determine the aspect ratio of flows for a given volume and effusion rate. To get a relative sense of how aspect ratios compare within each regime, values are normalized to basalt on land in Table 5 and plotted against volume (assuming an effusion rate of $10 \text{ m}^3 \text{ s}^{-1}$) in Figure 5. Note that changing the effusion rate changes the relative differences in aspect ratios among regimes, but does not change the relative differences within a regime.

The table and plot show that composition is dominant over environment in determining aspect ratio within a regime. This is especially true for regimes BNI and BPI, where the high viscosity of rhyolite results in aspect ratios ~30-300 times greater than those of basalt. The effect is strongest for regime BPI, reflecting a greater dependence of aspect ratio on yield strength than viscosity (Figure 5b). Aspect ratios are virtually the same in regimes BNC and BCS because crust viscosities and yield strengths are assumed to be invariant in regard to composition and environment. Within a given compositional field, significant differences among environments are only found for rhyolite in the BNI regime. The lower viscosity expected for Venusian rhyolite results in aspect ratios that are ~1/3 of those on land. In contrast, basalts have virtually the same aspect ratio in the three environments. The effect of buoyancy underwater is largely nullified by the increased density and lower viscosity assumed for submarine lavas. Where overpressure is dominant (PCS), flows have aspect ratios of 1/2, regardless of composition, environment, or volume.

Table 5. Normalized Aspect Ratios

Environment and Lava Type	BNI	BNC	BPI	BCS	PCS
Venus basalt	0.88 (1.1)	0.95 (1.1)	0.93 (1.1)	0.91 (1.1)	1.0
Venus rhyolite	34 (89)	1.1 (1.3)	283 (331)	1.3 (1.5)	1.0
Land basalt	1.0	1.0	1.0	1.0	1.0
Land rhyolite	84	1.2	303	1.4	1.0
Seafloor basalt	0.92 (1.3)	1.1 (1.3)	1.1 (1.5)	1.1 (1.6)	1.0

Values in parentheses assume land densities, viscosities, and yield strengths. Land basalt = 1.

It is important to emphasize that the aspect ratios derived here are individually non-unique. They result from the values of density, viscosity, yield strength, buoyancy, and thermal diffusivity used. Nevertheless, the relative differences among environments and between compositions should be approximately accurate.

Discussion

Evaluation of Volcanic Landforms

Ambient influences on lava flows are evaluated best when cooling and shape are considered together. As equations (9a)-(9e) show, extrusion rate (Q) partially determines the relative differences in aspect ratio among regimes. This factor depends on the eruption process and is largely independent of ambient conditions. Therefore, to effectively compare aspect ratios of flows of different compositions or in different environments, one of two assumptions must be made: (1) lava flows of different compositions in different environments are in the same regime, or (2) lava flows pass through more than one regime as they cool, but aspect ratios are "imprinted" in one regime and change little in other regimes.

We can consider flows to be either crust-rich or crust-poor. This is dependent upon both the cooling rate and extrusion rate [Fink and Griffiths, 1990]. The driving forces of flows with very thin or no crust are inhibited mostly by interior viscosity (BNI) or interior yield strength (BPI). On Earth, eruptions with rapid extrusion rates, in which large volumes of lava are extruded before crust forms, are strongly affected by these regimes. Examples of such high extrusion rate lavas include submarine sheet flows [Gregg and Fink, 1995, 1996], subaerial

flood basalts and shield flows, and voluminous rhyolite flows and domes on land [Manley, 1993]. Conversely, where cooling is rapid or extrusion rates are slow, the crust will be more dominant. Terrestrial examples include submarine pillow flows, which have thick crust due to rapid cooling (Figure 4), and low-volume subaerial lava domes, whose thick crust is caused, in large part, by low extrusion rates [Griffiths and Fink, 1993]. In these cases regimes BNC, BCS, and PCS will apply.

We now evaluate the effects the above analyses have on interpretations of two enigmatic volcanic landforms on Venus. First we consider the "unusual" Venusian volcano located between Artemis Chasma and Imdr Regio. It is unique on Venus in having large flows with festoons like those seen on terrestrial silicic flows [Moore et al., 1992]. However, the flows extend 360-400 km from the volcano's center, much longer than rhyolite flows on Earth [Manley, 1993]. The reasons for this can be explained by the effects of the Venusian environment on silicic lava flows. Due to the high surface temperature on Venus (~750 K), lava crusts will be thinner than they are for the same composition on Earth (Figure 4 and discussion above). Venusian lavas will therefore be able to spread farther before being arrested by the high viscosity or strength of solid crust. The slightly higher driving forces on Venus will enhance spreading even more. Furthermore, the thinner crust of Venusian rhyolites implies that they are more influenced by regimes BNI and BPI than terrestrial rhyolites are. Given the factor of 10 lower viscosity expected for Venusian rhyolites, silicic flows should be longer and have lower aspect ratios than those on Earth. It is therefore suggested that the lava flows from this volcano are silicic, as the festoons suggest [Moore et al., 1992], but due to the unique conditions on Venus are able to flow much farther than

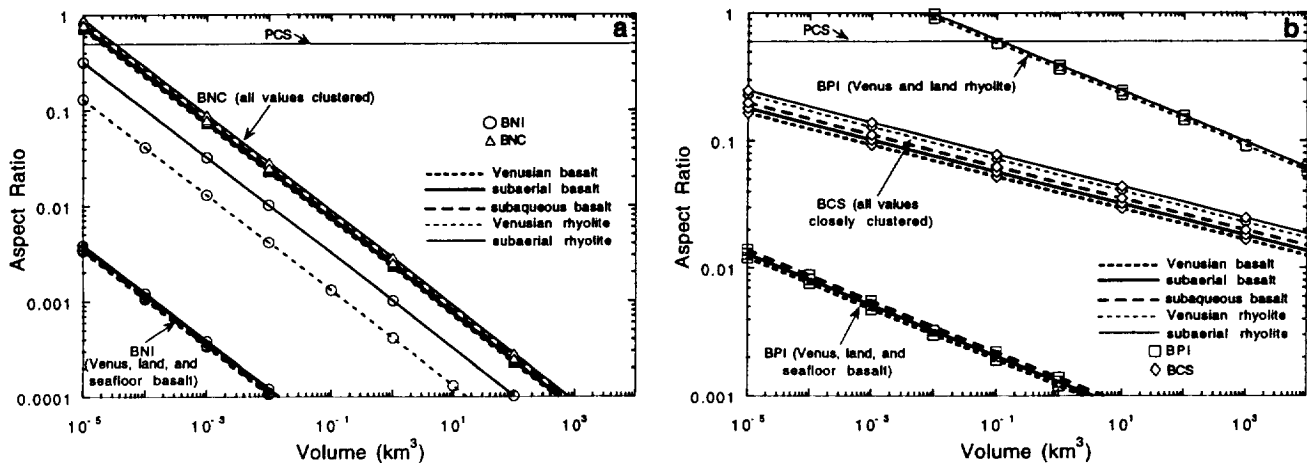


Figure 5. Aspect ratio versus volume for lava flows extruded at a rate of $10^3 \text{ m}^3 \text{ s}^{-1}$. Abbreviations refer to the regimes discussed in the text and are defined in equations (9a)-(9e). Viscous regimes (BNI and BNC) are shown in Figure 5a. Plastic regimes (BPI and BCS) are shown in Figure 5b. Regime PCS (lava overpressure resisted by a brittle shell) has a volume-independent aspect ratio of 1/2 and is shown in both figures.

rhyolites on Earth can. Conversely, digitate and lobate flows lacking festoons, such as those composing fluctii [Roberts *et al.*, 1992] and emanating from shields like Sif Mons [Senske *et al.*, 1992], are more consistent with basaltic compositions. The prediction here that basalt properties on Venus and Earth are nearly the same is supported by the fact that these flows have dimensions similar to flows in terrestrial flood basalt provinces [Roberts *et al.*, 1992].

Steep-sided domes are another interesting, apparently volcanic, landform on Venus. Many of these bear a remarkable resemblance to seamounts [Bridges, 1994, 1995; Sakimoto, 1994]. Both structures commonly have steep sides, flat tops, and similar ancillary features [Bridges, 1995]. The widths of cataloged Venusian steep-sided domes, at ~10 to over 70 km, are similar to seamounts with widths of tens of kilometers [Smith, 1988]. This is in contrast to subaerial silicic domes, which are generally less than 1 km wide [Pike, 1978]. The aspect ratios of silicic domes are above 0.1 and those of seamounts are near 0.1. Some Venusian domes have apparent aspect ratios that are lower than this, in a few cases below 0.01. However, many of these appear to be nonpristine domes that have had their heights lowered by tectonization and embayment from surrounding plains lavas [Bridges and McGill, 1997]. Therefore seamounts make better morphologic and morphometric analogs to Venusian domes than terrestrial silicic domes do. Does this mean that steep-sided domes are basaltic like seamounts are? Venusian basalts are expected to have thinner crusts and higher driving forces than those on the seafloor (Table 3). However, in the initial stages of eruption, prior to the formation of thick crust, aspect ratios will be determined by regimes BNI and BPI. Because the differences in internal viscosities and yield strengths for basalts on Venus and the seafloor are negligible, these regimes will contribute little to any systematic differences in aspect ratio. This is in contrast to rhyolites, which are expected to have very different aspect ratios in regimes BNI and BPI on Venus and Earth and much higher aspect ratios than basalt (Table 5). Once crust forms on basalt, regimes BNC, BCS, and PCS will dominate, preserving the similar aspect ratios. Therefore, for a given extrusion rate, basalts and the volcanoes they produce are expected to be similar on Venus and the seafloor. This supports a basaltic interpretation for Venusian steep-sided domes. Nevertheless, a silicic model cannot be ruled out. As mentioned in the introduction, large volumes of silicic lava may be stable under Venusian pressures [Pavri *et al.*, 1992]. Nevertheless, a basaltic composition seems just as likely for the Venusian domes and, if true, indicates that Venus is very deficient in silicic volcanics.

Caveats

Eruption history has an important influence on morphology. Many volcanic structures are made of compound lava flows, each extruded in a distinct eruptive episode. For example, the aspect ratios of terrestrial lava domes depend not only on extrusion rate and cooling rate, but also on the number and duration of repose periods between eruptions [Bridges, 1992; Fink and Bridges, 1995]. Pillow lavas in seamounts are layered, indicating episodic deposition [Batiza and Vanko, 1983; Staudigel and Schmincke, 1984]. The episodic layering, combined with enhanced cooling underwater [Searle, 1983], may be necessary ingredients in the formation of steep-sided and flat-topped seamounts. Whether such an episodic process formed the Venusian domes is unknown.

These examples serve to illustrate that all factors must be considered when trying to explain lava flow morphology. Nevertheless, ambient factors cannot be overlooked. Regardless of extrusion rate or eruption history, lava flows are influenced by the surrounding environment to some degree. Future studies should consider these ambient factors when

comparing volcanic morphologies on Venus, the land, and the seafloor.

Conclusions

1. The high-pressure environments of Venus and the seafloor repress the exsolution of carbon dioxide and, to a lesser extent, water, in lavas. The inhibition of volatile exsolution causes lavas to have fewer vesicles. Submarine and Venusian lavas are expected to be 1.4 and 1.3 times denser, respectively, than lavas of the same composition on land. This assumes that preeruptive volatile contents are the same on Venus as they are on Earth.

2. Submarine basalts cool the fastest of any lava composition in any environment. Rhyolites solidify more quickly than basalts on Venus and land. For a given composition, solidification times are similar on Venus and land, especially for rhyolite. The effect of increased lava density on the solidification of submarine and Venusian lavas is negligible.

3. Crust growth is fastest for basalts and slowest for rhyolites at times less than about 1 hour. In this case, basalts, especially those on the seafloor, have the thickest crusts. However, at later times, the low latent heat of rhyolite relative to basalt causes it to have the thickest crust. For a given composition, crusts on Venus will be thinner than those on land or the seafloor at all but the earliest times.

4. Confining pressures on Venus and the seafloor are sufficient to inhibit the exsolution of H₂O. By itself, this reduces basalt viscosities by 20-30% compared to values on land if the same preeruptive water contents are assumed in the three environments. Viscosities of Venusian rhyolites are reduced by an order of magnitude.

5. Gravitational driving forces of lava flows scale as the product of lava density and reduced gravity. These factors work against each other on the seafloor, resulting in driving forces that are only slightly less than on land. Similarly, the higher lava density on Venus is counteracted by the slightly lower gravity, resulting in driving forces that are also close to those on land for a given composition.

6. All of these factors affect the interpretations of Venusian volcanoes and lava flows. Because basalts on Venus are expected to behave much like basalts on the seafloor, the similarity between Venusian steep-sided domes and submarine basaltic seamounts implies similar compositions. In contrast, large festooned flows on Venus are consistent with a silicic composition. Their huge lateral dimensions are a likely consequence of rhyolite having a lower viscosity and being able to flow farther than rhyolite on Earth. A major caveat to these hypotheses is that other factors, such as extrusion rate and eruption history, also influence flow morphology. They must also be considered, in addition to ambient factors, in explaining volcanic landforms on Venus.

Acknowledgments. Detailed, constructive reviews by S.M. Baloga and an anonymous referee were extremely helpful and are gratefully acknowledged. Unofficial reviews by C.R. Manley and G.E. McGill significantly improved the manuscript. R.W. Griffiths kindly supplied the computer program used to compute lava cooling. Discussions with D.A. Clague and J.G. Moore improved the section on vesicularity. This work was performed under NASA Planetary Geology and Geophysics grant NAGW-4263 to the University of Massachusetts.

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N.T. Bridges, Department of Geosciences, University of Massachusetts, Amherst, MA 01003. (e-mail: bridges@wawa.geo.umass.edu)

(Received October 2, 1996; revised January 28, 1997; accepted February 5, 1997.)