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

Nathan T. Bridges

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. It 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, eruption 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 10⁻³-0.1 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 formation 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
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 crystalization (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;
have vesicle contents restricted to less than 5 vol %. The Hawaiian lavas at pressures greater than 20 bars have a wide range of vesicle contents from 1-50 vol %. The second family of points is along the 20 bars and have a wide range of vesicle contents from 1-50 vol %. Differences in flow properties among environments are not considered in detail. These roughly approximate the two endmember 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 % H2O [Moore, 1965, 1970; MacPherson, 1984; Dixon et al., 1988] and < 0.2 wt % CO2 [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 CO2 exsolves at all ocean depths, H2O mostly at depths < 500-2000 m (~50-200 bars), and SO2 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 H2O and SO2 exsolution at shallow depths. When ultradepth 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 CO2 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

<table>
<thead>
<tr>
<th>Environment and Lava Type</th>
<th>H2O, wt %</th>
<th>Vesicles, vol %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Venus basalt</td>
<td>0.3</td>
<td>10</td>
</tr>
<tr>
<td>Venus rhyolite</td>
<td>4</td>
<td>10</td>
</tr>
<tr>
<td>Land basalt</td>
<td>0.05</td>
<td>30</td>
</tr>
<tr>
<td>Land rhyolite</td>
<td>3</td>
<td>30</td>
</tr>
<tr>
<td>Seafloor basalt</td>
<td>0.5</td>
<td>5</td>
</tr>
</tbody>
</table>

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.
The shape of such a curve reproduces, to a certain extent, the pressure, a functional dependence found 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 (\(\rho\)) is related to the solid density (\(\rho_s\)), and vesicle fraction \(x\) by

\[
\rho = \rho_s (1-x).
\]

This is easily shown in a plot of vesiculality versus bulk density for 61 samples of Hawaiian basalt measured by Robertson and Peck [1974] (Figure 2). Taking the vesicle...
Table 3. Driving Forces in Planetary Environments

<table>
<thead>
<tr>
<th>Environment and Lava Type</th>
<th>Lava Bulk Density, kg m$^{-3}$</th>
<th>Reduced Gravity, m s$^{-2}$</th>
<th>Driving Force/Unit Volume, N m$^{-3}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Venus basalt</td>
<td>2700</td>
<td>8.6 (8.5)</td>
<td>2.3 x 10$^4$ (1.8 x 10$^5$)</td>
</tr>
<tr>
<td>Venus rhyolite</td>
<td>1980</td>
<td>8.5 (8.4)</td>
<td>1.7 x 10$^4$ (1.3 x 10$^5$)</td>
</tr>
<tr>
<td>Land basalt</td>
<td>2100</td>
<td>9.8</td>
<td>2.1 x 10$^4$</td>
</tr>
<tr>
<td>Land rhyolite</td>
<td>1540</td>
<td>9.8</td>
<td>1.5 x 10$^4$</td>
</tr>
<tr>
<td>Seafloor basalt</td>
<td>2850</td>
<td>6.4 (5.1)</td>
<td>1.8 x 10$^4$ (1.1 x 10$^5$)</td>
</tr>
</tbody>
</table>

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.

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, $\kappa$, 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:

$$\kappa = \frac{k}{pc},$$

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 basalt 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$^{-7}$ m$^2$s$^{-1}$, respectively [Peck et al., 1977, Williams and Mc Birney, 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 Mc Birney, 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

$$T(z,t) = T_i - \frac{1}{pc} \int_0^t \frac{F(X) \gamma^2 \lambda}{1 \gamma^2 \lambda - \lambda} d\lambda,$$  \hspace{1cm} (4)

where $\lambda$ 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 ($\kappa$), 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 (~10$^{12}$ Pa s) and, as a result, exhibit some solid-like properties, including brittle failure (see viscosity section below) [Bottigia, 1994; Bottigia et al., 1995; McMillan and Wolf, 1995; Richet and Bottigia, 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...
composition, regardless of H2O content. Transition temperatures are assumed to be the same for a given range of water contents of interest here. Therefore glass substantially [Dingwell, 1995]. However, there are no cooling times is similar to that found by counterparts on land. The relative magnitude of the basalt there to solidify a bit faster than their compositional they erupt at a lower temperature. The slightly greater cooling change the times among the are relatively minor compared to the differences in cooling increase slightly when adjusted to reflect the different lava 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, \( z_* \), of a crystallized layer in a cooling lava flow is [Turcotte and Schubert, 1982, p. 169]

\[
z_* = 2\lambda_1/\kappa T,
\]

(5a)

The symbol \( \lambda_1 \) is a constant and is found iteratively using (5a) and the following equation:

\[
e^{-2z_1^2/L} = \left( \frac{L\sqrt{\pi}}{\lambda_1 t_{\text{erf}}(\lambda_1)} \right) e^{C(T_m - T_0)},
\]

(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 \( \times \) 10^6 and 2.9 \( \times \) 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 [Piwinskii, 1968; Best, 1982]. The ambient temperature \( (T_a, \text{Table I}) \) is used for \( T_0 \) because at times of \( 10^5 \) 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 \( 10^5 \) 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.
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 crusts 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 SiO₄ tetrahedra, reducing viscosity. The extent of this reduction depends on the degree of polymerization before H₂O 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 H₂O exsolution (see vesiculation discussion above) and should therefore have lower viscosities than those on land. Taking the relationship between log viscosity and wt % H₂O 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 H₂O 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 % H₂O) and the seafloor (0.5 wt % H₂O). 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 H₂O than those on land. For 1073 K rhyolite, going from 3 wt % H₂O (land) to 4 wt % H₂O (Venus) reduces viscosity from ~10⁸ to 3 x 10⁷ Pa s (800°C curve in Figure 5 of Shaw [1963]). However, the log viscosity versus wt % H₂O 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 10⁵ Pa s and those on Venus 10⁴ Pa s (Table 4). This is consistent with other experimental data showing that the addition of 1 wt % H₂O to silicic melt reduces...
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 \(10^4\) on Venus versus a factor of \(10^2\) 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 \(\text{GeO}_2\) 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 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 \(10^6\) 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 stresses are very high, viscous deformation will cease. Therefore at times greater than about \(10^5\) 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% \(\text{SiO}_2\) and Fink and Zimbelman [1990] judged the yield strength of freshly erupted Hawaiian basalt to be \(\sim 1000\) 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, 100 Pa is assumed. Based on the trend of yield strength versus \(\text{SiO}_2\), the yield strengths of rhyolites are of the order of \(10^3\) 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 \(10^5\) 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]

\[
p g H^2 R. \tag{6a}
\]

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

\[
p g H^2. \tag{6b}
\]

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 \(p\) is the lava bulk density [Griffiths and Fink, 1993]. The reduced gravity, or buoyancy, is expressed as

---

**Table 4. Rheologic Parameters in Planetary Environments**

<table>
<thead>
<tr>
<th>Environment and Lava Type</th>
<th>(\eta), Pa s</th>
<th>(\eta_c), Pa s</th>
<th>(\sigma), Pa</th>
<th>(\sigma_c), Pa</th>
<th>(E), m s(^{-1})</th>
<th>(E_c), m s(^{-1})</th>
<th>(\Gamma), m(^{-1})</th>
<th>(\Gamma_c), m(^{-1})</th>
</tr>
</thead>
<tbody>
<tr>
<td>Venus basalt</td>
<td>800 (10^{12})</td>
<td>1000 (10^8)</td>
<td>29 (18)</td>
<td></td>
<td></td>
<td>2.3 (10^4) (1.8 (10^5))</td>
<td>2.3 (1.8)</td>
<td>2.3 (10^4) (1.8 (10^5))</td>
</tr>
<tr>
<td>Venus rhyolite</td>
<td>(10^7) (10^{12})</td>
<td>(10^7) (10^8)</td>
<td>1.7 (10^3) (1.3 (10^4))</td>
<td></td>
<td></td>
<td>1.7 (10^4) (1.3 (10^5))</td>
<td>1.7 (10^4) (1.3 (10^5))</td>
<td></td>
</tr>
<tr>
<td>Land basalt</td>
<td>1000 (10^{12})</td>
<td>1000 (10^8)</td>
<td>21</td>
<td></td>
<td></td>
<td>2.1 (10^4)</td>
<td>2.1</td>
<td>2.1 (10^4)</td>
</tr>
<tr>
<td>Land rhyolite</td>
<td>(10^7) (10^{12})</td>
<td>(10^7) (10^8)</td>
<td>1.5 (10^4)</td>
<td></td>
<td></td>
<td>1.5 (10^4)</td>
<td>1.5 (10^4)</td>
<td></td>
</tr>
<tr>
<td>Seafloor basalt</td>
<td>700 (10^{12})</td>
<td>1000 (10^8)</td>
<td>(0.26) (11)</td>
<td></td>
<td></td>
<td>1.8 (10^4) (1.1 (10^5))</td>
<td>1.8 (1.1)</td>
<td>1.8 (10^4) (1.1 (10^5))</td>
</tr>
</tbody>
</table>

Here, \(\eta\), viscosity; \(\eta_c\), crust viscosity; \(\sigma\), yield strength; \(\sigma_c\), crust yield strength; \(E\), driving force per unit volume + viscosity; \(E_c\), driving force per unit volume + crust viscosity; \(\Gamma\), driving force per unit volume + yield strength; and \(\Gamma_c\), driving force per unit volume + crust yield strength. Values in parentheses assume land densities, viscosities, and yield strengths.
where $g'$ is gravity and $p_a$ is the density of the ambient fluid. The value of $g'$ is significant for submarine flows. The density of water ($1000 \text{ kg m}^{-3}$) 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 $\theta$ 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 ($p g'$) 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 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\% 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:

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

[from a point source] and

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

[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 ($t = \frac{V}{Q}$), 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)

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

Regime 2 (BPI)

$$A = \frac{1}{2} \left( \frac{1}{V^4} \right)^{1/5} \quad \text{(9b)}$$

Regime 3 (BNC)

$$A = \left( \frac{1}{V^4} \right)^{1/14} \quad \text{(9c)}$$

Regime 4 (BCS)

$$A = \frac{1}{2} \left( \frac{1}{V^4} \right)^{1/8} \quad \text{(9d)}$$

Regime 5 (PCS)

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

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

$$E = \frac{p g'}{\eta} \quad \text{and} \quad \Gamma = \frac{p g'}{\sigma_c} \quad \text{(10a)}$$

$$E_c = \frac{p g'}{\eta_c} \quad \text{and} \quad \Gamma_c = \frac{p g'}{\sigma_{c_c}} \quad \text{(10b)}$$

The subscript $c$ refers to parameters of the outer lava crust. Complete derivations of the original equations are given by Griffiths and Fink [1993].

The driving forces per unit volume ($p g'$) 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 about 15-20\% less than on Venus or land. If a subaerial lava density is used, then driving forces are only 1\% 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.
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 very thin or no crust are strongly affected by these regimes. Examples of such high extrusion rate lavas include submarine pillow 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

**Table 5. Normalized Aspect Ratios**

<table>
<thead>
<tr>
<th>Environment and Lava Type</th>
<th>BNI</th>
<th>BNC</th>
<th>BPI</th>
<th>BCS</th>
<th>PCS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Venus basalt</td>
<td>0.88 (1.1)</td>
<td>0.95 (1.1)</td>
<td>0.93 (1.1)</td>
<td>0.91 (1.1)</td>
<td>1.0</td>
</tr>
<tr>
<td>Venus rhyolite</td>
<td>3.4 (89)</td>
<td>1.1 (1.3)</td>
<td>283 (331)</td>
<td>1.3 (1.5)</td>
<td>1.0</td>
</tr>
<tr>
<td>Land basalt</td>
<td>1.0</td>
<td>1.0</td>
<td>1.0</td>
<td>1.0</td>
<td>1.0</td>
</tr>
<tr>
<td>Land rhyolite</td>
<td>84</td>
<td>1.2</td>
<td>303</td>
<td>1.4</td>
<td>1.0</td>
</tr>
<tr>
<td>Seafloor basalt</td>
<td>0.92 (1.3)</td>
<td>1.1 (1.3)</td>
<td>1.1 (1.5)</td>
<td>1.1 (1.6)</td>
<td>1.0</td>
</tr>
</tbody>
</table>

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

![Figure 5](image-url)  
**Figure 5.** Aspect ratio versus volume for lava flows extruded at a rate of 10 m$^3$ 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.
Despite extrusion rate or eruption history, lava flows are considered when explaining lava flow morphology. The prediction here that basaltic properties on Venus and Earth are nearly the same is supported by the fact that these flows have dimensions similar to those in terrestrial flood basalt provinces. 

Steep-sided domes are another interesting, apparently volcanic, landform on Venus. Many of these bear a remarkable resemblance to seamounts. Both structures commonly have steep sides, flat tops, and similar ancillary features. 

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 suggests 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 thinnest 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 large lateral dimensions are a likely consequence of rhyolite having a lower viscosity and being able to flow farther than rhyolite on Earth. A major cave 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.

**References**


