VOLCANIC PROCESSES AND LANDFORMS ON VENUS: THEORY, PREDICTIONS, AND OBSERVATIONS

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Abstract. Volcanic activity is a fundamental mechanism of heat transfer from planetary interiors, and the characteristics, distribution, and morphology of volcanic deposits provide significant insight into (1) the relation of volcanism and tectonism, (2) eruption style, and (3) the chemistry and volatile content of the eruption products. Eruption styles and processes on the planets are known to be strongly influenced by such factors as gravity, temperature, and atmospheric characteristics. We model the ascent and eruption of magma on Venus in the current Venus environment, taking into account the influence of the extreme surface temperatures (650-750 K) and pressures (4-10 MPa) on these processes. These conditions produce a thermal gradient difference such that the temperature is higher at a given depth on Venus than on Earth, and a pressure distribution difference leading to much smaller ratios of subsurface to surface pressure on Venus than on Earth. Among the more significant consequences of this for volcanic style on Venus is that there will be less cooling of magma in the initial stages of ascent and that once the magma reaches the surface, convective heat losses will be much more important than in the subaerial terrestrial environment because of the higher atmospheric gas density. In general, however, our treatment suggests that there is no reason to expect large systematic differences between lava flow morphologies on Venus and Earth. On the other hand, conditions on Venus will tend to inhibit the subsurface exsolution of volatiles, and pyroclastic eruptions involving continuous magma disruption by gas bubble growth may not occur at all unless the exsolved magma volatile content exceeds several weight percent. However, Strombolian activity, in which bubble coalescence can cause sufficient concentration of gas to produce intermittent explosions in low-viscosity magmas ascending slowly toward the surface, can occur at much lower volatile contents. If pyroclastic eruptions do occur, pyroclastic fragment velocities and clast cooling will be less than on Earth, and the higher atmospheric pressure and temperature will cause convective cloud rise heights to be considerably lower, and pyroclasts to be much less widely dispersed, than on Earth. For example, eruption cloud heights of 50 km (suggested as a means of raising sulfur dioxide into the upper atmosphere) (Esposito, 1984) could only be reached if exsolved magma volatile contents exceeded 4 wt %, regardless of gas species. On the basis of our analysis, a series of predictions can be made concerning the expected characteristics of volcanic deposits and landforms on Venus. Comparison of these predictions with recent observations from Venus, Arecibo, and Venera data support the view that regional pyroclastic deposits are very rare, that magma volatile contents do not commonly exceed about 4 wt %, and that the atmospheric pressure has been about the same as the present value over a time period equivalent to the average age of the northern areas of the northern hemisphere (500-1000 Ma (Barukov et al., 1986)). Volcanism and morphology of volcanic deposits and landforms on Venus. Comparison of these predictions with recent observations from Venus, Arecibo, and Venera data support the view that regional pyroclastic deposits are very rare, that magma volatile contents do not commonly exceed about 4 wt %, and that the atmospheric pressure has been about the same as the present value over a time period equivalent to the average age of the northern areas of the northern hemisphere (500-1000 Ma (Barukov et al., 1986)). Volcanic processes and landforms on Venus: theory, predictions, and observations.

1. Introduction

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Fig. 1. Radar images of possible volcanic features on Venus. (a) Arecibo radar image of south central Beta Regio; north at top of image. Variations in brightness are related to differences in small-scale surface roughness values (wavelength size is 12.6 cm). Darkest areas are smoothest; brightest areas are roughest. Resolution is approximately 2 km. Large bright area is Theia Mons, interpreted to be a shield volcano [see Campbell et al., 1984]. Parallel lines northeast of Theia Mons are interpreted to be faults associated with the Devana Chasma rift system. (b) Sketch map of area of Figure la; Theia Mons, flanked by flowlike lobes, is situated on the western bounding fault of the Devana Chasma rift system, filling the rift valley and covering much of the fault system. B indicates bright rough areas, and the central dark smooth region at the summit of Theia Mons is portrayed in black. P-P' indicates location of profile shown in Figure 1c. (c) Comparison of the topography of Theia Mons on Venus, Hawaii on Earth (above ocean floor), and Olympus Mons on Mars (above surrounding plain). Theia Mons is superposed on the general positive topography of Beta Regio and rises less than about 3 km above this broad topographic swell. Location of radar bright zone seen in Figures la and lb are noted. (d) Arecibo radar image of area southeast of Ishtar Terra showing possible volcanic terrain [Head et al., 1985a]. Arrow at 1 points to base of bright shieldlike structure between 200 and 300 km in width and containing a dark summit region. Individual flows extend downslope and apparently change width and direction at the base. At 2, several narrow flows diverge and extend downslope for several hundred kilometers to the southeast. Several other more degraded sources and shieldlike structures are seen in this image. Global topographic maps of Pettengill et al. [1980] or Head et al. [1980b] show location of these regions.
planets [Basaltic Volcanism Study Project, 1981]. Study of the nature and distribution of volcanic deposits, combined with models of the basic principles of the ascent and eruption of magma [Wilson and Head, 1981], can provide information on mechanisms of lithospheric heat transfer; planetary chemical and thermal evolution; links between volcanism, tectonism, and the state of stress in the lithosphere; and the nature of planetary resurfacing processes. Comparison of volcanic deposits and features on planets with different bulk physical properties and atmospheric characteristics is also an important factor in assessing and refining the models for magma ascent and eruption [Wilson and Head, 1983].

The mechanisms for lithospheric heat transfer on Venus are not well understood [Solomon and Head, 1982], although volcanism is thought to play an important role [Morgan and Phillips, 1983]. On the basis of radar images and Pioneer Venus mission results, several regions on Venus appear to be influenced by volcanic activity [McGill et al., 1981; Masursky et al., 1980; Campbell et al., 1984; Head et al., 1985a]. High-resolution radar images reveal volcanic centers and geologic units interpreted to be flows [Campbell et al., 1984; Barsukov et al., 1984a, b, 1986; Head et al., 1985a]. In this paper we model the ascent and eruption of magma on Venus in the current Venusian environment, taking into account the influence of the extreme surface temperatures and pressures on these processes. We predict the nature of effusive and explosive eruptions on Venus and compare these to similar eruptions under Earth conditions. Finally, we use observations and measurements of probable volcanic deposits on Venus (Figure 1) to
make a preliminary assessment of the nature and spectrum of Venusian volcanic activity. We conclude with a discussion of surface observations required during future exploration of Venus in order to assess the full range of volcanic activity and to understand the role of volcanism in the evolution of the planet.

2. Production and Ascent of Magma on Venus

2.1. Conditions in the Lithosphere

To produce quantitative assessments of the nature of possible eruptions on Venus, we first need to assume values for those physical properties of the magmas which control the eruption process, the most important of which are the rheological parameters defining stresses in which the magmas respond to shearing. We need values for the bulk densities of the magmas and for the pressure gradients or density contrasts driving them to the surface from high-level magma reservoirs.

Although the rheological properties of magmas, especially at subliquidus temperatures, are in general complex [e.g., Shaw, 1969], there can be modeled adequately, for the purposes of describing many aspects of their eruptive behavior, as Bingham plastics [Hulé, 1974]. The response of the magma to shearing stresses is then defined by a yield strength γ, which the applied stress must exceed before deformation occurs, and a plastic viscosity μp, which is the (constant) ratio of applied stress in excess of the yield strength to strain rate. Both of these rheological parameters are functions of magma composition and temperature.

Modeling studies of the interior of Venus suggest that although there must be broad similarities to Earth, there may also be important differences in the chemistry [Wood et al., 1981]. It is by no means a trivial observation, therefore, that the major element compositions measured at two sites on the Venus surface by the Venera landers [Surkov et al., 1983] were very similar to those of two types of terrestrial basalt: an alkali basalt at the Venera 13 site and a tholeiite at Vesuvius. We will use the properties of these magmas as starting points for estimating the properties of candidate Venusian magmas.

No direct evidence was obtained from the Soviet probes about the minor element and volatile compositions of the rocks at the landing sites; however, the low water content and high halogen content of the carbon-dioxide-dominated atmosphere [Choffman et al., 1980] suggest that magmas erupted on Venus (at least in geologically recent periods) may be water depleted (and possibly carbon dioxide or halogen enriched) relative to otherwise similar terrestrial counterparts [Kingwood and Anderson, 1977]. Both of these factors tend to increase the solidus and liquidus temperature of basalts and would imply that the pressure gradients on the two planets were equal, magmas source regions and crustal reservoirs would not be found as close to the surface on Venus as on Earth. That the temperature gradients on the two planets are not the same, however, is one of the consequences of the very different atmospheres.

On Venus the large atmospheric mass is responsible for both a high surface pressure (4-10 MPa) and a high surface temperature (650-750 K) and so leads to smaller values for both the pressure and temperature gradients in the upper part of the crust relative to those of Earth. The pressure gradients are significantly different only at depths of about 1 km of the surface, whereas the temperatures in the two planets at a given depth differ over a much greater range of depths. The fact that the temperature is higher at a given depth on Venus than on Earth will, therefore, tend to counteract the effects of possible water depletion of Venusian magmas on pressures at great depths on the planet. A further consequence of the lower Venusian temperature gradient is that there will be less cooling of magma in the final stages of its approach to the surface. This means that the minimum fissure width needed to allow magma to reach the surface before significant chilling occurs will be smaller on Venus than on Earth and that the magma emerging through a fissure of a given width will be hotter on Venus. However, the differences between the two cases will not be very great: the temperature difference between the magma and the surface rocks will be about (1500-750 =) 750 K on Venus and about (1500-300 =) 1200 K on Earth, so that the magma cooling on Venus is about 100 K less than on Earth. The effect of these differences on the rheology of Venusian magmas is small.

For basaltic magmas in terms of a dimensionless temperature parameter φ given by \((T_0 - T_e)/(T_o - T_s)\), where \(T_e\) is the temperature of the environment (in this case near-surface rock), \(T_o\) is the eruption temperature of the magma, and \(T_s\) is its solidus temperature. Substituting typical values for these parameters on Earth and Venus shows that while \(\phi\) is close to 0.96 for terrestrial eruptions, the value will be closer to 0.94 for basalts on Venus. However, the numerical results of Delaney and Pollard [1982] demonstrate that except for eruptions in which the magma rise is very close to being suppressed as a result of cooling effects, differences in \(\phi\) of the above order will have very little effect on the amount of heat loss from an ascending magma. We will assume, therefore, that the temperatures of magmas moving through the Venusian crust lie between the liquidus and a value about 100 K lower than the liquidus.

Figure 2 shows some of the few data currently available on the variation with temperature of plastic viscosity and yield strength for tholeiitic and alkali-basaltic magmas [Murase, 1962; Shaw, 1969] and an anodesite sample [Mc Birney and Murase, 1984]. Comparison of the data for an anhydrous tholeiite (curve A in Figure 2a) with measurements on a similar hydrated melt (curve B) shows that anhydrous Venusian melts may have viscosities about one order of magnitude larger at a given temperature, and liquid at some 50 K higher, than terrestrial equivalents. However, as we have argued above, the Venusian magmas should erupt at similar temperatures relative to their liquidi. The two sets of dashed lines in Figure 2a link
pairs of corresponding viscosities implied for terrestrial and Venusian magmas on the basis of the above arguments and demonstrate that anhydrous magmas on Venus may commonly be a factor of 2–4 more viscous than terrestrial counterparts. For tholeiitic melts, plastic viscosities will probably lie in the range 300–3000 Pa s.

Figure 2b shows the variation of yield strength with temperature for a hydrated tholeiitic magma and an andesite. There are no other experimental data relevant to tholeiites or alkali-basalts. To estimate yield strengths of Venusian magmas, we therefore appeal to the fact that yield strength increases with decreasing melt temperature (due to polymerization or crystallite formation [Murase, 1962]) for any fixed composition and decreases with any trend toward more mafic composition at a fixed temperature. Given our above arguments concerning temperatures of Venusian magmas moving through the crust, we adopt values of $Y$ ranging up to $10^4$ Pa. Furthermore, the fact that both the viscosity and yield strength of a magma decrease with increasing temperature at fixed composition suggests that we should associate these two parameters systematically, using say $(Y = 0.01$ Pa, $\eta = 400$ Pa s) and $(Y = 10^7$ Pa, $\eta = 3300$ Pa s) to define the extremes of the range of variation of rheological properties of tholeiitic magmas approaching the Venusian surface: the relationship is shown in Figure 2c.

Fig. 2a. Variation of viscosity with temperature for three terrestrial basaltic melts: curve A, a dry tholeiite; curve B, a hydrated tholeiite; and curve C, a hydrated alkali basalt. Data are taken from Murase [1962] and Shaw [1969]. The dashed lines show the consequences of assuming that basalts on Venus are systemati- cally water-depleted and hotter than otherwise similar basalts on Earth (see text for details).

Fig. 2b. Variation of yield strength with temperature for hydrated terrestrial tholei-ites: curve D, andesite [McBirney and Murase, 1984]; other data sources as for Figure 2a.

Fig. 2c. The combinations of viscosity and yield strength implied for anhydrous tholeiites on Venus by combining the data from Figures 2a and 2b.
We can now estimate the minimum fissure width $w_{\text{min}}$ in the crust which will allow magma to reach the surface from a given depth of origin, $H$, Wilson and Head [1981] show that the need to avoid excessive cooling of the magma dominates over other considerations and that the critical relationship for magma rise to the surface in an open fissure is

$$g\Delta p w_{\text{min}}^4 - 2\omega_{\text{min}}^3 = 48\kappa H$$  \hspace{1cm} (1)

where $g$ is the planetary gravity, $\Delta p$ is the effective density difference driving the magma toward the surface (which may include a component due to excess volatile pressure in the source reservoir if such exists and is likely to be close to 100 kg m$^{-3}$ [Wilson and Head, 1981]), and $\kappa$ is the thermal diffusivity of the magma (close to $7 \times 10^{-7}$ m$^2$ s$^{-1}$ for essentially all compositions). For the above ranges of magma rheological properties, $w_{\text{min}}$ lies in the range 0.3-1.1 m if the source depth $H$ lies between 1 and 10 km; if $H$ is as large as 100 km, the upper limit on $w_{\text{min}}$ rises to 1.9 m.

These minimum fissure widths are very similar to values calculated for basalts on Earth and the moon [Wilson and Head, 1981] (a natural consequence of the dominantly fourth root dependence of $w_{\text{min}}$ on the other variables). Such calculated minimum fissure widths are only about a factor of 3 less than the actual dike widths commonly observed in eroded volcanic terrains on Earth (1-5 m [see Delaney and Pollard, 1982]) and, since the latter widths are presumably controlled by the amount of tectonic strain which accumulates in an area before an eruption starts, this observation serves to stress the intimate relationships which must exist between the style of a volcanic episode and the tectonic environment in which it occurs. There is currently considerable debate about the kinds of tectonic regimes to be expected on Venus [Phillips et al., 1981; Solomon and Head, 1982; Phillips and Malin, 1983; Barukov et al., 1984a, b, 1986], but we note that we can obtain a conservative but plausible estimate of the mass fluxes to be expected in Venusian eruptions by assuming that actual fissure widths exceed minimum values by a factor of 2. The general formulae for rise rates and mass eruption rates of magma to an open fissure are given by Wilson and Head [1981]; on substituting the above ranges of magma properties into these formulae and using actual fissure widths which are a factor of 2 greater than the minimum values corresponding to a given set of magma properties, we find that mass eruption rates should range from 100 to about 6000 kg s$^{-1}$ min$^{-1}$ for fissure widths of horizontal length of fissure. However, it is well established that many basaltic magmas on Earth do not erupt through open fissures but rather migrate upward as bodies of discrete vertical extent through a dike which opens in front of the magma and partly closes behind it (see, for example, Shaw [1980] and Spera [1980]). Stevenson [1983] has shown that under these circumstances, magma velocities and hence mass fluxes may be up to about 10 times less than the values appropriate to open fissures of the same mean width. Thus the range of expected mass eruption rates on Venus is probably 10 kg s$^{-1}$ m$^{-2}$ to a few times $10^7$ kg s$^{-1}$ m$^{-2}$. This range of values is essentially identical to that observed for terrestrial basaltic eruptions [Wilson and Head, 1981].

2.2. Processes in the Vent

Although the mass discharge rate in an eruption is controlled mainly by the fissure width and magma rheology in the lithosphere, the details of eruption style are determined by processes occurring within the region extending from a few hundred meters below the surface to a few tens of meters above it. The vital subsurface process is exsolution of volatiles; the supersurface process is interaction of the magma with the atmosphere (see Figure 3). The latter is particularly important if there is an explosive component to the eruption style.

The main effect of the high surface pressure and lower subsurface pressure gradient on Venus on erupting magmas is the suppression or reduction of exsolution of volatiles [Wood, 1979; Head and Wilson, 1982; Garvin et al., 1982]. Vesiculation to the point of explosive disruption of the magma will not occur at all on Venus unless the total dissolved magma volatile content exceeds about 1.5 wt % at the 4-MPa level in the Venusian highlands or about 3.5 wt % at the 10-MPa level in the Venusian lowlands [Wilson and Head, 1983]; these values are very uncommon in terrestrial basalts. The only exception to this rule concerns low-viscosity magmas rising slowly to the surface in narrow fissures, for which coalescence of gas bubbles can occur leading to Strombolian explosions [Garvin et al., 1982]. Even when explosions do take place in Venusian magmas, less gas expansion occurs between the depth at which gas bubble nucleation begins and the level at or above the vent at which sufficient pressure is reached than on any other terrestrial planet. Substitution of the appropriate pressures into the generalized models of volcanic explosion processes given by L. Wilson [1980] then shows that the velocities of the pyroclastic fragments produced will be smaller by a factor of about 2.5 than those of pyroclasts from terrestrial volcanic explosions with similar volatile contents [Garvin et al., 1982; Wilson and Head, 1981].
either natural or forced convection will eventually dominate the radiation and conduction processes. Solid crust thickness will increase with time and hence distance from the vent.

The motion of a flow lobe will cease when cooling fronts have penetrated to a sufficient fraction of its thickness.

1983). As a consequence, fire fountains on Venus (Figure 3) will be denser (in terms of the number of pyroclasts per unit volume) and will have a greater optical thickness than fountains from terrestrial eruptions with the same mass flux. This will lead to less heat loss for most of the clasts (see, for example, Wilson et al. [1982]) and thus both welding of near-vent deposits and, more importantly, coalescence of pyroclasts on landing to form lava flows will be more likely. Thus a number of factors collaborate to ensure that explosive basaltic eruptions are much less likely to occur on Venus than on Earth and that if magmas do erupt explosively on Venus, the clasts falling near the vent are likely to be subject to less cooling on Venus than on Earth by the process of emerging through the vent.

A further aspect of the reduction of volatile release on Venus concerns the rheological properties of the lava flows which move away from the vent. Magma rheology is strongly controlled by its volatile content, especially the H₂O content, as we have seen above. Sparks and Pinkerton [1978] have pointed out that volatile release by terrestrial magmas can commonly lead to a significant increase in viscosity and yield strength. The solidus and liquidus temperatures rise as the volatiles are lost, and even if the temperature of the lava itself does not change at all during its passage through the vent, its temperature relative to the liquidus decreases, with a consequent increase in the rate of phenocryst formation.

In summary, both the temperature and the rheological properties of the near-vent lava flows are likely to be much closer to the properties of the magma at depth on Venus than on Earth. Next, we consider the subsequent motion of such flows.

3. Effusive Eruptions

3.1. Cooling of Lava Flows

The morphological evolution, surface texture, and maximum length of a lava flow are all controlled ultimately by its thermal state, which changes with distance from the vent as a result of cooling processes. Heat loss takes place by conductive heat transfer into the ground over which the flow moves, by radiation from all exposed lava surfaces, and if a planetary atmosphere is present, by either natural or forced convection (Figure 4).

All of these processes occur for lava flows on both Venus and Earth; there are, however, significant differences in their relative and absolute importance [Frenkel and Zabaluyeva, 1983]. The following analysis uses standard treatments of the various heat loss mechanisms taken from McAdams [1954].

Radiative loss from the surface of a flow is controlled by the surface temperature T₀ and emissivity ε; the heat energy loss per unit time per unit surface area is Fᵣ, where

$$ Fᵣ = ε (T₀^4 - Tₐ^4) $$

ε is the Stefan constant (5.6703 x 10⁻⁸ W m⁻² K⁻⁴) and Tₑ is the effective temperature of the environment into which the heat is lost. For both Earth and Venus, Tₑ is taken as the temperature of the near-surface atmosphere: the values used in the subsequent illustrative calculations are 300 K for Earth, 650 K for the Venusian highlands and 750 K for the Venusian lowland areas. Given the measured reflectivities of rock-forming minerals [Washburn et al., 1926], the emissivities of all basaltic rocks can conveniently be taken as 0.75, and so the main cause of differences in radiative cooling rates of lavas between Venus and Earth is the high Venusian surface temperature. Table 1 shows some typical values of Fᵣ as a function of T₀ for the two planets.

If a finite atmospheric wind is present over a lava surface, heat loss also occurs by forced convection. The forced heat loss rate per unit surface area of the flow Fₚ is obtained from

$$ Nu = Pr ^{3/4} (0.036 Re ^{0.8} - 836) $$

where the dimensionless Nusselt number (Nu), Frandtl number (Pr), and Reynolds number (Re) are defined by

$$ Nu = \frac{Fₚ L}{k (T₀ - Tₑ)} $$

$$ Pr = \frac{μ c}{k} $$

$$ Re = \frac{W L β}{μ} $$

In these equations, L is the typical horizontal length scale of the lava flow (generally taken as the arithmetic mean of the length and width of the flow, though, as we shall see, the exact value is not important), W is the mean speed of the wind, and K, c, μ, and β are the thermal conductivity, specific heat at constant pressure, viscosity, and density of the atmospheric gas, respectively (air in the case of Earth, essentially pure CO₂ in the case of Venus). The values of the latter four parameters are to be evaluated at the mean temperature of the lava-atmosphere interface, 1/2 (T₀ + Tₑ) and can be obtained from Washburn et al. [1926]. Table 1 shows examples of values of Fₚ.
TABLE 1. Thermal Properties of Lava Flows on Venus and Earth

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</tbody>
</table>

Lava surface temperature: $T_0$; forced convection, natural convection, and radiation heat loss fluxes: $F_P$, $F_N$, and $F_R$, respectively; total heat loss flux: $F_{T_0}$; time since eruption of the lava: $t_r$; thickness of solid crust on the surface of the flow: $C$.

calculated using $W = 1$ m/s on Venus [Marov et al., 1973; Avduevskii et al., 1976; Counselman et al., 1980; Seiff et al., 1980] and $W = 10$ m/s on Earth. The value of $L$ used is 1 km for both planets and probably lies within a factor of 10-30 of the appropriate value for most flows. Examination of equations (3), (4), and (6) shows that $F_P$ varies only as the fifth root of $L$, so that a thirty-fold error in $L$ will yield only a factor of 2 error in $F_P$.

If the ambient wind speed is sufficiently small, natural convection takes over from forced convection and the relevant relationship is

$$Nu = 0.14 \left( \frac{Gr \ Pr}{3} \right)^{1/3}$$

where now the naturally convected heat flux per unit area, $F_N$, replaces $F_P$ in equation (4); $Pr$ is still given by equation (5), and $Gr$ is the Grashof number:

$$Gr = g \alpha \left( \frac{T_0 - T_e}{L} \right) \frac{L^3 g^2}{\mu^2}$$

in which $g$ is the acceleration due to gravity and $\alpha$ is the volume expansion coefficient of the gas. Substitution of (4) and (8) into (7) shows that $F_N$ is independent of $L$, the size of the flow. Comparison of the values of $F_P$, $F_N$, and $F_R$ given in Table 1 shows that radiative heat losses dominate convective losses for terrestrial flows until the surface temperature falls below about 500 K. In contrast, convective losses are much more important on Venus, mainly due to the high atmospheric gas density, and even if forced convection is absent, natural convection losses exceed radiation losses at temperatures below about 1000 K. The total heat loss rate from a Venusian flow surface $F_{T_0}$, taken to be the sum of the radiative loss and whichever of the forced and natural convective losses is the greater, is seen to be about 1.5 times larger than the corresponding rate for a terrestrial flow at the same temperature at temperatures exceeding about 500 K.

The two aspects of lava flow cooling which are most relevant to the comparison of flows on Earth and Venus are the variations with time since learning the vent of (1) the temperature $T_0$ of a given element of the flow surface, and (2) the thickness $C$ of its rigid crust defined here as the depth at which the solidus temperature $T_s$ is reached. $T_s$ is taken as 1525 K on Venus and 1475 K on Earth on the basis of the thermal arguments presented earlier. Settle [1979a] has stressed the importance of latent heat release in retarding the interior cooling of lava flows, and we use an analysis based on this principle given by Turcotte and Schubert [1982]. $C$ is given as a function of time $t$ by

$$C = 2 \lambda \left( \frac{\kappa t}{2} \right)^{1/3}$$
where \( \lambda \) is a dimensionless scale factor given by

\[
\frac{1}{2} \frac{Q \pi^{3/2}}{c_L (T_s - T_0)} = \exp \left( -\lambda^2 \right) \frac{\lambda}{\text{erf} (\lambda)}
\]  

(10)

Q and \( c_L \) are the latent heat of fusion and the specific heat of the magma (taken as \( 4 \times 10^7 \text{ J kg}^{-1} \) and \( 1500 \text{ J kg}^{-1} \text{ K}^{-1} \), respectively), and erf (\( \lambda \)) is given by

\[
\text{erf} (\lambda) = \frac{2}{\pi^{1/2}} \int_0^\lambda \exp (-z^2) \, dz
\]  

(11)

convenient values of which are tabulated by Carslaw and Jaeger [1959]. The vertical temperature gradient in the flow between the surface and the base of the solid crust is [Turcotte and Schubert, 1982]

\[
\frac{\partial T}{\partial y} = \frac{(T_s - T_0)}{(\pi k t)^{1/2}} \text{erf} (\lambda)
\]  

(12)

The method of solution is to select a value of \( T_0 \) from Table 1 and evaluate the left-hand side of equation (10). Turcotte and Schubert [1982] give a graphical solution of (10) from which \( \lambda \) can be determined. This value of \( \lambda \) is inserted, together with \( T_0 \), into equation (13) to solve for \( t \), the time from the commencement of cooling on leaving the vent at which these values of \( T \) and \( \lambda \) are reached. Finally, \( \lambda \) and \( t \) are inserted into equation (9) to find the corresponding value of \( C \), the thickness of the solid crust. Figure 5 gives the values of \( T_0 \) and \( C \) as a function of time evaluated in this way for flows on Venus and Earth and shows that during most of the first hour after leaving the vent, surface temperatures on terrestrial flows will be greater than on Venusian flows: convective heat loss is very efficient in the dense Venusian atmosphere. At times significantly longer than 1 hour, however, temperatures will be higher on Venusian flows as the flow surface temperature asymptotes to the higher ambient temperature. The thickness of the rigid crust on a flow is greater on Venus than on Earth immediately after the flow leaves the vent; however, after about 30 min, by which time the crust thickness exceeds 60 mm, the crust is thicker on a terrestrial flow. These results are confirmed by an analysis given by Frenkel and Zabaluyeva [1983] and lead to the following comparisons between the expected features of lava flows on the two planets.

3.2 Lengths of Lava Flows

Flow units which come to rest after traveling for less than 1 hour will be systematically longer, if we consider only the consequences of cooling, on Earth than on Venus, whereas flows traveling for much more than 1 hour will be longer on Venus. This relationship follows from the nature of the cooling process in the interior of a flow. The thickness of cooled crust on a flow is an indicator of the penetration of a wave of cooling into the flow; and since the rheological properties of a lava are dominated by its temperature, the wave of cooling may be thought of as a wave of modification of rheology. A flow will continue to move on a given slope under a given stress regime when its interior rheology has been sufficiently modified [Kilburn, 1983]. It is found empirically for terrestrial basaltic flows that motion will cease when the value of the dimensionless Graetz number \( G_z \) has fallen from an initially high value to about 300 [Hulme and Fielder, 1977; Pinkerton and Sparks, 1978]. \( G_z \) is defined as

\[
G_z = \frac{u D^2}{k \kappa} = \frac{D^2}{\kappa t}
\]  

(14)

where \( D \) is the flow thickness, \( \kappa \) is the lava thermal diffusivity, \( x \) is the distance flowed by the lava at mean speed \( u \), and \( t \) is the time since leaving the vent. \( G_z \) is simply the reciprocal of the parameter \( \kappa t / D^2 \) which is commonly used to characterize cooling problems, and \( G_z = 300 \) corresponds approximately to a situation in which the central one half of the vertical thickness of a flow is unaffected by cooling (see section 3.4 of Carslax and Jaeger [1959]). In the treatment given above, we use the relative thicknesses of crusts as a function of time on Venusian and terrestrial lava flows given in Figure 5b as an index of the relative penetration of thermal waves, we find that for flows moving for periods much less than 1 hour, the time available for motion will be about 1.3 times longer on Earth than Venus; for flows moving for significantly longer than 1 hour (by far the commonest circumstance on Earth), the time available will be about 1.1 times longer on Venus than Earth. This ratio is significantly smaller than that found in an earlier analysis which neglected the buffering effect of latent heat release [Head and Wilson, 1982] and implies that the maximum length of lava flows can be written

\[
X = \frac{-D^2 u}{\Lambda \kappa}
\]  

(15)

where \( \Lambda \), the critical Graetz number, is taken as 300 for Earth and 270 for Venus. The lengths of lava flows given above is somewhat idealized relative to the behavior of real basaltic flows. It has been tacitly assumed here that the liquid lava motion is laminar beneath a rigid crust. In most lava flows the deformation is relatively laminar, but shearing is commonly observed to take place in the surface layers, especially near the vent, in such a way as to expose fresh, hot lava to the atmosphere. This process inevitably increases the rate of heat loss from a given part of the flow surface and ultimately leads to a greater rate of thickening of the crust with time than calculated. Also, the above treatment is intended to apply to lava moving in open channels rather than in the closed
tubes which form when the surface crust adheres to the channel sides and creates a stationary roof [Greeley, 1970; Peterson and Swanson, 1974]. It is well established that tube-fed flows lose heat less efficiently than channel-fed flows and hence travel farther [Hulme and Fielder, 1977]. The factors controlling the onset of termination of lava flows are not well understood, but it seems reasonable to assume that the relative enhancement of crust formation on Venusian lavas soon after eruption will encourage the onset of tube formation on Venus relative to Earth.

3.3. Pahoehoe and Aa Textures on Flows

It is common on Earth for basaltic lavas to be erupted with a pahoehoe surface texture and to undergo a transition to aa texture at some distance from the vent. Peterson and Tilling [1980] have shown that the transition can occur over a wide range of conditions of motion but that for a given lava, there is a specific set of combinations of strain rate and apparent viscosity (defined as the instantaneous ratio of the shear stress to the strain rate) at which the transition will occur. Kilburn [1981] has extended this analysis by noting that the set of combinations of strain rate and apparent viscosity defined by Peterson and Tilling in fact corresponds to the stress-strain rate curve for a Bingham plastic material at a specific temperature. This suggests that cooling of a flow is the dominant factor in determining the texture transition and that the faster a flow cools, the earlier the transition is
likely to occur; however, this is necessarily a statistical argument since other factors such as the local topographic slope and the effusion rate control the detailed shear history of a flow and hence the exact distance from the vent at which the transition occurs. The lava surface temperature curves of Figure 5a indicate that pahoehoe-aa texture changes on flows on Venus should occur, on average, after times which are about 2/3 those which would apply on Earth.

The surfaces of pahoehoe lavas show a characteristic "ropy" texture, consisting of folds generated where the flow surfaces undergo horizontal shortening when encountering changes of slope or constrictions [Fink and Fletcher, 1978]. On Earth the tightest folds commonly have characteristic horizontal wavelengths when seen from above of a few tens of millimeters and are superimposed on second-generation folds with wavelengths which are longer by a factor of about 2, leading to a spectrum of wavelengths between about 0.05 and 0.25 m. Fink and Fletcher [1978] have developed a detailed model of the formation of these folds in terms of the variation of viscosity with depth in the outermost layers of a laminar flow and show how the characteristic wavelength can be expressed as a function of three parameters, Y, R, and S. Examination of equations (1), (2), and (10) and the definition of R by Fink and Fletcher [1978] show that if we define \( T_{00} \) to be the temperature of a flow below its cooled crust (so that \( T_{00} \) is equal to the value of \( T_0 \) as the lava leaves the vent at \( t = 0 \)), we can express Y, R, and S in terms of the variables defined earlier:

\[
\begin{align*}
\gamma &= \frac{(T_{00} - T_0)}{c} \quad (16) \\
R &= \exp \left( \frac{T_{00} - T_0}{\gamma} \right) \quad (17) \\
S &= \frac{g c}{(T_{00} - T_0)} \exp \left( \frac{T_{00} - T_0}{\gamma} \right) \quad (18)
\end{align*}
\]

where for simplicity we have omitted all the mathematical and material property factors which would be the same on both Earth and Venus. Figure 5 gives the variation of \( T_0 \) and C with time for representative flows on the two planets and so allows the relative values of \( \gamma \), R, and S to be found for these flows, also as a function of time. Figure 10a of Fink and Fletcher [1978] can then be used to find the relative values of the dominant wavelengths of the folds which will grow under similar shearing conditions on the two planets.

We find that at times up to about 30 min after eruption, the values of \( \gamma \) and R are larger on Venus than on Earth by factors of about 1.2 and 6, respectively, whereas the value of S is smaller by a factor of about 7. As a result, the dominant fold wavelengths will typically be systematically longer on Venus than on Earth by a factor of about 1.5, leading to predicted values up to about 0.4 m; Figure 10b of Fink and Fletcher [1978] shows that the higher values of R and lower values of S on Venus will also lead to a threefold to fivefold enhancement of the growth rate of the folds in pahoehoe flows on that planet. At times significantly greater than a few hours after eruption, relative conditions change so that \( \gamma \) and R are smaller on Venus than on Earth by factors of about 1.3 and 2.7, respectively, and S is larger by a factor of about 31, [1978]. Despite the simplification with some aspects of its general applicability [Sparks et al., 1976; Moore et al., 1976; Kibbun, 1983], it is the only complete, internally consistent model of lava flow morphology currently available. Lava motion is assumed to occur at an average speed \( u \) in a channel of width \( W_c \) and central thickness \( D_c \). A stationary levee or bank of width \( W_t \) and maximum thickness \( D_b \) flanks each side of the central channel. If the flowing material has density \( \rho \) and moves down a slope of \( \alpha \) radians (\( \alpha << 1 \)), it can be shown [Hulme, 1974] that

\[
\begin{align*}
D_b &= \frac{\gamma g}{\rho} \quad (19) \\
W_t &= W + 2 W_b \quad (22)
\end{align*}
\]

The central channel width \( W_c \) is a function of the rheological parameters and the volume rate of eruption of magma from the vent, \( F \), which can be expressed [Wilson and Head, 1983] as

\[
\begin{align*}
W_c &= \left\{ \begin{array}{ll}
(24 \frac{F \gamma}{\rho})^{1/3} & , W_c/2W_b \leq 1 \\
(24 \frac{F \gamma}{\rho})^{1/3} & , W_c/2W_b \leq 1
\end{array} \right. \quad (21)
\end{align*}
\]

If the total width of the central channel and levees is \( W_c \), then

\[
W_c = W + 2 W_b \quad (22)
\]

Hulme [1974] shows that the actively moving part of the flow has a cross-sectional area \( A \) given by

\[
A = \frac{2 \rho g}{3 Y} \left( \frac{D_c^3}{D_b^3} \right) \quad (24)
\]
The issue of differences in flow morphology rests very heavily, therefore, on the question of magma rheology. In deriving the relationships noted in Figure 2c we commented that if magmas on Venus are relatively dry and erupt at similar temperatures relative to the liquidus compared with terrestrial magmas, then viscosities might be about a factor of 3 greater on Venus than on Earth, while yield strengths should have about the same values. Examination of equations (19)-(27) then shows that we would expect flows on Venus having a given major element composition and effusion rate, and forming on a given slope, to have widths and depths which would be about the same as those on equivalent terrestrial flows but to have central channel widths larger by a factor of 1.2, mean flow speeds smaller by a factor of 1.8, lengths smaller by a factor of 1.1, and flow durations longer by a factor of 1.65. It will be noted that all of these factors are less than 2, so that no gross systematic differences exist between Venusian and terrestrial flows of similar magmas.

Table 2 shows some examples of the properties of flows forming on slopes typical of the major terrain units on Venus [Sharpton and Head, 1985]; the slopes used are taken from near the upper end of the slope distribution in each case and are 0.003 (lowland units), 0.0045 (upland rolling plains units), and 0.006 (interior of the Beta region) rad, respectively. Properties are calculated for a typical pair of rheological parameters from within the range defined by Figure 2c: $Y = 100 \text{ Pa}$, $\eta = 1500 \text{ Pa s}$. A spread of magma effusion rates has been chosen which includes the current terrestrial range for basalts, commonly extending up to about 300 m$^3$/s [Whitford-Stark, 1982] but sometimes reaching $7 \times 10^4$ m$^3$/s in basaltic Plinian eruptions [Walker et al., 1984], and the inferred range for the lunar mare-filling basaltic eruptions, extending up to $3 \times 10^5$ m$^3$/s [Head and Wilson, 1981]. If effusion rates on Venus are limited to the same range as on Earth, it is predicted that flow widths and lengths of several tens of kilometers can be reached on moderate slopes by hot, dry basalts. If larger values are assumed for any of the parameters slope, yield strength, or effusion rate, an increase in flow length is produced if all three parameters are doubled, for example, predicted flow lengths would increase by a factor of 3.5, thus ranging up to a few hundred kilometers.

This sequence of equations allows us to make a number of points. First, since the gravity on Venus is only slightly less than that on Earth and since the densities of lavas on Venus will be similar to (or, in view of the comments made earlier about low gas exsolution from Venusian magmas, slightly greater than) those of terrestrial lavas, the equations demonstrate that there is no reason to expect significant systematic differences between Venusian and terrestrial lava flow morphologies unless there are systematic differences in topographic slope, effusion rate, or lava rheology. Studies of the radar altimetry data from the Pioneer Venus orbiter have shown that the range of topographic slopes on Venus, 0.001-0.03 rad, is similar to that on Earth [Sharpton and Head, 1985]. We have argued that there is no obvious reason to expect different ranges of effusion rates on Venus from those on Earth unless there are systematic differences between magma rheologies on the two planets. The issue of differences in flow morphology rests very heavily, therefore, on the question of magma rheology.

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TABLE 2. Illustration of Geometric and Mechanical Properties of Dry Tholeiite Lava Flows on Venus and Earth

<table>
<thead>
<tr>
<th>F, Wc, Wl, Dc, uc, X, T,</th>
<th>Earth, Average Slopes: α = 0.0007; Db = 6.24 m; Wb = 4.46 km</th>
</tr>
</thead>
<tbody>
<tr>
<td>m³/s</td>
<td>km</td>
</tr>
<tr>
<td>1 0.9</td>
<td>9.8</td>
</tr>
<tr>
<td>10 1.9</td>
<td>10.9</td>
</tr>
<tr>
<td>10² 4.2</td>
<td>13.1</td>
</tr>
<tr>
<td>10³ 9.0</td>
<td>17.9</td>
</tr>
<tr>
<td>10⁴ 20.9</td>
<td>29.8</td>
</tr>
<tr>
<td>10⁵ 48.2</td>
<td>57.1</td>
</tr>
<tr>
<td>10⁶ 111.0</td>
<td>120.0</td>
</tr>
<tr>
<td>Venus, Modal Slope: α = 0.0015; Db = 2.91 m; Wb = 971 m</td>
<td></td>
</tr>
<tr>
<td>1 0.5</td>
<td>2.5</td>
</tr>
<tr>
<td>10 1.2</td>
<td>3.1</td>
</tr>
<tr>
<td>10² 2.6</td>
<td>4.5</td>
</tr>
<tr>
<td>10³ 6.0</td>
<td>7.9</td>
</tr>
<tr>
<td>10⁴ 13.8</td>
<td>15.7</td>
</tr>
<tr>
<td>10⁵ 31.8</td>
<td>33.8</td>
</tr>
<tr>
<td>10⁶ 73.5</td>
<td>75.4</td>
</tr>
<tr>
<td>Venus, Lowlands: α = 0.003; Db = 1.46 m; Wb = 243 m</td>
<td></td>
</tr>
<tr>
<td>1 0.3</td>
<td>0.8</td>
</tr>
<tr>
<td>10 0.8</td>
<td>1.3</td>
</tr>
<tr>
<td>10² 1.8</td>
<td>2.3</td>
</tr>
<tr>
<td>10³ 4.1</td>
<td>4.6</td>
</tr>
<tr>
<td>10⁴ 9.4</td>
<td>9.9</td>
</tr>
<tr>
<td>10⁵ 21.8</td>
<td>22.3</td>
</tr>
<tr>
<td>10⁶ 50.4</td>
<td>51.8</td>
</tr>
<tr>
<td>Venus, Upland Rolling Plains: α = 0.0045; Db = 0.97 m; Wb = 107.9 m</td>
<td></td>
</tr>
<tr>
<td>1 0.3</td>
<td>0.5</td>
</tr>
<tr>
<td>10 0.6</td>
<td>0.8</td>
</tr>
<tr>
<td>10² 1.4</td>
<td>1.6</td>
</tr>
<tr>
<td>10³ 3.3</td>
<td>3.5</td>
</tr>
<tr>
<td>10⁴ 7.6</td>
<td>7.8</td>
</tr>
<tr>
<td>10⁵ 17.5</td>
<td>17.7</td>
</tr>
<tr>
<td>10⁶ 40.4</td>
<td>40.6</td>
</tr>
<tr>
<td>Venus, Beta Interior: α = 0.006; Db = 0.73 m; Wb = 60.7 m</td>
<td></td>
</tr>
<tr>
<td>1 0.2</td>
<td>0.3</td>
</tr>
<tr>
<td>10 0.5</td>
<td>0.6</td>
</tr>
<tr>
<td>10² 1.2</td>
<td>1.3</td>
</tr>
<tr>
<td>10³ 2.8</td>
<td>2.9</td>
</tr>
<tr>
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<td>6.6</td>
</tr>
<tr>
<td>10⁵ 14.9</td>
<td>15.0</td>
</tr>
<tr>
<td>10⁶ 34.5</td>
<td>34.6</td>
</tr>
</tbody>
</table>

Flow parameters (defined in the text and Figure 6) are given for a lava density of 2600 kg/m³, yield strength of 100 Pa, and viscosity of 1500 Pa s.

and overlapping shields rather than well-defined lava flows.

We have included in Table 2 the possibility of very high effusion rate basaltic eruptions. High effusion rates have been associated, on both Earth and the moon, with heating of the ground underlying a flow to the point where thermal erosion occurs, and if the eruption continues for a sufficient period, a sinuous rille channel is formed [Hulme, 1973; Carr, 1974; Hulme and Fielder, 1977; Wilson and Head, 1980; Huppert et al., 1984; Wilson and Mouginis-Mark, 1984]. However, the essential prerequisite for efficient thermal erosion is turbulence in the moving lava rather than a high effusion rate per se [Hulme, 1973]. We have calculated the Reynolds number = d c / μ / ν for each set of flow conditions given in Table 2 and find that the highest value is 75, much less than the critical value 42000 needed for the onset of turbulence. None of the flows whose properties are given in Table 2 would be expected to lead to significant thermal erosion of their substrates during their emplacement periods. Turbulence, and hence thermal erosion, could occur, however, if
magnas with sufficiently low viscosities were erupted on Venus.

4. Explosive Eruptions

4.1. Volatile Sources

Explosive volcanic activity always involves the release and expansion of a vaporized volatile phase. This is true whether the activity accompanies relatively steady release of magma from a vent or consists of a series of abrupt and intermittent episodes of discharge of blocking debris. The vapor may be generated by volatile exsolution from the magma or may be derived from the decomposition and/or evaporation of substances existing on the surface in the vicinity of the vent or present as a component of crustal rocks through which the magma rises. On Earth the magmatic volatiles are generally dominated by \( \text{H}_2\text{O}\) or \( \text{CO}_2\), though sulfur compounds and halogens may be important in some magmas [Anderson, 1975; Luhr et al., 1984; Varekamp et al., 1984]. The main volatile derived from surface layers is \( \text{H}_2\text{O} \) in the form of liquid water or ice; water is also the main subsurface volatile, though decomposition of carbonate sediments cut by feeder conduits may release small amounts of \( \text{CO}_2 \) into some erupting magnas [Lilie et al., 1973]. On Venus the predominance of \( \text{CO}_2 \) in the atmosphere may suggest that \( \text{CO}_2 \) is currently the major magma volatile; however, while some models of the formation of Venus call for extreme initial \( \text{H}_2\text{O} \) depletion, there is as yet no direct evidence that Ven- usian magma need contain appreciable amounts of \( \text{H}_2\text{O} \). Additionally, as on Earth, other volatile species such as halogens or sulfur compounds may be present in significant amounts. However, data on the solubility of volatiles in magnas which are adequate for modeling the detailed eruption dynamics exist only for \( \text{CO}_2 \) and \( \text{H}_2\text{O} \), and we will use these species as being representative of the effects of low- and high-solubility volatiles, respectively. In considering nonjuvenile volatiles, we note that the surface temperature and geothermal gradient on Venus are such as to ensure that liquid water, water ice, and solid \( \text{CO}_2 \) cannot exist in the lithosphere (in contrast, liquid sulfur, if present, would be stable everywhere on the Venus surface under present conditions). We will examine the possibility that thermal decomposition of carbonates may release \( \text{CO}_2 \) to interact with some magnas.

4.2 Explosion Mechanisms

A fundamental feature of all explosive eruptions is the generation of gas (by exsolution from the magma or evaporation of accidental volatiles) at some pressure which is greater than that of the planetary atmosphere. It is the expansion of the gas down to atmospheric pressure which provides the energy to drive explosions. The amount of energy released is roughly proportional to the natural logarithm of the ratio of the initial and final gas pressures [L. Wilson, 1980]. In the case of a magma progressively exsolving volatiles as it approaches the surface, there is no single pressure at which gas expansion begins; however, the final velocity of the ejected gas and pyroclasts in the vent can be quite closely predicted if it is assumed that gas expansion effectively begins at the pressure, and hence depth below the surface, at which the magma is disrupted from a continuous liquid into a mixture of gas and fluid fragments [L. Wilson, 1980]. In the case of transcritical explosions, the pressure built up under a chilled retaining cap by either evaporation of accidental volatiles or exsolution of magmatic volatiles, the starting pressure differential for the gas expansion is essentially equal to the strength of the cap [Self et al., 1979]. Both of these initial pressure conditions are very close to being independent of the planetary environment, and so it is mainly the final pressure reached by the expanding gas, i.e., the atmospheric pressure, which controls the nature of the explosion. It is inevitable, therefore, that the high atmospheric pressure will lead to less gas expansion and hence smaller ejection velocities in explosive eruptions on Venus [Wood, 1982]. We examine the consequences of this phenomenon in detail in later sections.

When magma does rise steadily to the surface, exsolving gas into growing bubbles, disruption of the magma probably takes place when the volume fraction of the bubbles reaches about 0.75% [Sparks, 1978]. This requirement sets a limit on the minimum amount of gas needed to cause magma disruption. Because the solubility of all magmatic volatiles increases with pressure, the fact that Venus has a higher atmospheric pressure than Earth means that greater initial magma volatile contents are needed there to ensure that an eruption is explosive. We have used the method outlined by Wilson and Head [1981] for \( \text{CO}_2 \) and \( \text{H}_2\text{O} \) in common magnas to find the minimum volatile contents needed to produce explosions (see Table 3) for two sets of Venusian environmental conditions: atmospheric pressure = 100 MPa, ambient temperature = 750 K, corresponding to the lowland areas, and pressure = 4 MPa, temperature = 650 K, values more relevant to the upper highlands or the summits of proposed volcanic constructs [Campbell et al., 1986]. It is found that if the volatile phase is \( \text{H}_2\text{O} \), the minimum amounts needed lie within, but near the upper end of, the range of water contents inferred to be common in terrestrial magnas. If the volatile phase is \( \text{CO}_2 \), however, the amounts needed are very large compared with what is regarded as common on Earth; indeed, the gas solubility data show that a magma would need to rise from a source region at a depth between 35 and 83 km (for highland and lowland vents, respectively) to permit it to contain the required amount of the volatile. This result implies that if volatiles more soluble than \( \text{CO}_2 \) are not available to Venusian magnas, explosive activity may be rare on the planet. However, since there is no specific evidence precluding the presence of either of these volatiles in magnas on Venus, we will explore the nature of explosive eruptions involving both of them and draw some general conclusions relevant to events which may involve these or other volatiles.

4.3 Eruption Cloud Formation

All volcanic explosions produce some kind of mixture of gas and fragmental material which is
projected out into the atmosphere. These mixtures are generally denser than the surrounding atmosphere. Specifically, mixtures of silicate fragments and magmatic water vapor at magmatic temperatures on Earth will only be less dense than the atmosphere if the water vapor mass fraction exceeds 15%, a situation only likely to occur in explosions which concentrate the volatile into the upper part of the available magma: this can happen in Strombolian [Blackburn et al., 1976] and Vulcanian explosions [Self et al., 1979]. On Venus the corresponding limit is 45-50% (depending on the local atmospheric pressure) if CO₂ is the volatile phase and 18-20% if the volatile is H₂O.

Immediately after leaving the vent, the explosion products interact with the planetary atmosphere in two ways. Sufficiently coarse clasts decouple rapidly from the gas flow and follow approximately ballistic paths back to the ground. Sufficiently fine clasts remain locked to the gas flow regime and become involved in the next phase of the explosion, which is the entrainment and heating of atmospheric gases. In this way, thermal energy is converted to buoyancy (and hence kinetic energy), the bulk density of the mixture decreases, and a convecting cloud may form over the vent. A completely stable cloud can only form if atmospheric gas is mixed as far as the center of the cloud before the upward velocity of the material there decreases to zero as a result of the action of gravity [Sparks and Wilson, 1976]. If a stable condition is not reached, part or all of the ejected material will rise to a large fraction of the ballistic height which would have been reached in the absence of an atmosphere and then collapse back to the ground. Depending on the initial temperature and on the grain size and flight time, and hence amount of cooling, of the clasts, this material may form an unwelded ash/scoria deposit, a welded deposit of similar type, or a pyroclastic flow or may coalesce to form a lava flow.

Establishment of a stable convecting plume will be encouraged if the ejection velocity from the vent is large and the radius of the vent is small (since these conditions minimize the lateral distance to which the atmosphere must mix to reach the cloud center and maximize the time available for this to happen) and if the jet of material emerging from the vent is well-collimated. Calculations given by Sparks et al. [1978] can be used to show that the minimum eruption speed \( v_m \) needed to ensure stable convection is

\[
v_m = \frac{80 \, R_v \, (\rho_m - \rho_a)}{\beta_m} \quad (28)
\]

where \( \beta_m \) and \( \beta_a \) are the bulk density of the erupting gas-clast mixture and the density of the surrounding atmosphere, respectively. This relationship was developed for eruptions on Earth, but there is experimental evidence to suggest that the entrainment process is geometrically similar over a wide range of density contrasts between the inside of the cloud and the surrounding atmosphere [see Turner, 1979; Sparks and Wilson, 1982]. We shall use equation (28) in later sections to assess the stability of convecting eruption clouds on Venus and the likelihood of pyroclastic flow formation.

### 4.4. Heights of Eruption Clouds

Once stable convection is assured, the height to which a cloud can rise is determined by the structure of the surrounding atmosphere and the thermal input to the cloud. Theoretical studies of idealized plumes [Morton et al., 1956], numerical calculations for volcanic clouds, and field observations [Wilson et al., 1978; Settle, 1978] show that for clouds forming above a vent which is releasing material at a constant rate, the maximum height reached is proportional to the fourth root of the heat release rate (which is in turn proportional to the mass eruption rate of clasts over a time interval large enough to be entrained into the cloud). For clouds forming above sudden, discrete explosions, the height is proportional to the fourth root of the amount of heat released rather than the release rate [Morton et al., 1956]. However, if discrete explosions occur sufficiently close together in time, the result may be a relatively steady convecting cloud with a height determined by the time-averaged heat release rate. This is likely to happen if the time needed for the decay of the velocity field induced in the atmosphere by one explosion is less than the time interval between successive explosions. The decay time will certainly be less than the time which would be required for fragments projected upward at the same speed as the gas stream to come to rest under the action of gravity alone (since entrainment of the surrounding atmosphere will exert an additional retarding force on the ejecta). The ballistic rise time is equal to the initial velocity divided by the acceleration due to gravity and since the gravities on Venus and Earth are similar, we find that for both planets, a steadily convecting cloud will form over a vent producing discrete explosions if the time interval between explosions is less than the following values: for an ejection velocity of 10 m/s, 1 s; for a velocity of 30 m/s, 3 s; for 100 m/s, 10 s; and for 300 m/s, 30 s.

Early estimates of the rise heights of steady

### TABLE 3. Minimum Volatile Contents and Magma Source Depths Required to Ensure Explosive Activity in Venusian Magmas

<table>
<thead>
<tr>
<th>Magma-Volatile Combination</th>
<th>4 MPa Surface Pressure</th>
<th>10 MPa Surface Pressure</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Volatile Content, wt %</td>
<td>Source Depth, km</td>
</tr>
<tr>
<td>Water in basalt</td>
<td>1.1</td>
<td>1.0</td>
</tr>
<tr>
<td>Water in rhyolite</td>
<td>1.6</td>
<td>0.5</td>
</tr>
<tr>
<td>CO₂ in any melt</td>
<td>2.0</td>
<td>35</td>
</tr>
</tbody>
</table>

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eruption clouds on Venus [Head and Wilson, 1982] implied that they should reach heights about 0.35 times as large as those from equivalent terrestrial eruptions. Recently, Esposito [1984] has used improved data on the atmospheric thermal structure to show that the factor is probably closer to 0.6. Using this value and the observed relationship between cloud height \( H_c \) and mass discharge rate \( M \) observed for terrestrial eruption clouds [Wilson et al., 1978], we find

\[
H_c = 140 \left( \frac{M}{10^{-3} \text{ kg}} \right)^{1/4}
\]

(29)

for Venus, where \( H_c \) is in meters and \( M \) is in kilograms per second. The equivalent relationship for clouds from discrete explosions is

\[
H_c = 19.7 \left( \frac{M}{10^{-3} \text{ kg}} \right)^{1/4}
\]

(30)

where \( M \) is the total magma mass in kilograms injected into the cloud.

The height of an eruption cloud is the main factor controlling the pattern of dispersion of the pyroclasts which it releases. Model calculations [Wilson, 1976] suggest that the widths of air fall deposits some distance downwind of the vent are similar to the heights of the corresponding clouds, this relationship being largely independent of the nature of the planetary atmosphere [Wilson et al., 1982]. The downwind extent of an air fall deposit from a cloud of a given height is determined by the fine end of the grain size distribution in the explosion products and the ambient wind speed profile. We will consider some possible air fall deposit characteristics on Venus in a later section.

4.5. Explosion Classification Schemes

Explosive eruption styles on Earth have been classified in terms of the grain size characteristics and dispersal of pyroclastic products [Walker, 1973] and the mechanisms involved [L. Wilson, 1980]. There is some disagreement in the literature about what kind of activity is implied by some of the terms in common usage, and this is due in part to uncertainties about the nature of the mechanisms involved. Even when the processes giving rise to a given type of explosive activity are well understood, it is not safe to assume that the equivalent mechanisms operating in a planetary environment other than that of Earth will produce analogous eruption products [McGetchin and Head, 1973; Head and Wilson, 1979; Wilson and Head, 1981]. For this reason we will generally avoid terms based on the grain size and dispersal of pyroclasts when defining styles of explosive activity on Venus and will use the following small set of familiar terms which we feel can be closely related to the physical processes which take place in eruptions (on Earth and Venus): 1) Strombolian activity is the intermittent discharge of gas and fluid magma [Blackburn et al., 1976; Wilson and Head, 1981]. Magma may overflow the vent as an effusive lava flow as well as being involved in the explosions. 2) Hawaiian activity is the relatively steady discharge of magma which is disrupted below or at the surface into released gas and pyroclasts with a range of grain sizes sufficiently coarse that little of the pyroclastic material is entrained into a convecting gas cloud over the vent [Wilson and Head, 1981]. 3) Plinian activity is the relatively steady discharge of magma which has fragmented into pyroclasts having grain sizes sufficiently small that most of them are able to be entrained into an eruption cloud [Walker and Croasdale, 1971]. When magma discharge rates are sufficiently high, the exsolved magmatic gas fraction in the eruption products is sufficiently small, a stable convecting eruption cloud cannot be maintained [Sparks and Wilson, 1976; Sparks et al., 1978], and eruptions of this kind will instead produce large-scale pyroclastic flows (ignimbrites). 4) Vulcanian activity is the episodic discharge of magma, volatiles, and accidental debris where the ejecta are coarse-grained and consist mainly of the accidental debris [Self et al., 1979]. 5) Pelean activity is the episodic discharge of magma, volatiles, and accidental debris where the pyroclasts consist mainly of fine-grained juvenile material. Such eruptions, which commonly form small pyroclastic flows and small eruption clouds at each, have characterized the post-May 1980 activity of Mount St. Helens [Rowley et al., 1981].

We emphasize that while some of the above eruption styles are probably linked to specific magma types on all planets, this is not likely to be a general rule. Thus Strombolian activity is apparently only able to occur in low-viscosity magmas rising in narrow conduits or conduits controlled by impacts [Wilson et al., 1982]; however, whereas Vulcanian activity is not commonly associated with fluid basaltic magmas on Earth, this association appears to have occurred on the moon in the formation of the dark halo deposits inside the crater Alphonsus [Head and Wilson, 1979], probably as a result of the chilling of rising magma as it invaded the fragmental regolith layer forming in the uppermost few tens of meters of the lunar crust. Also, while Plinian activity is mainly, but not entirely [see Williams, 1983; Williams and Self, 1983; Walker et al., 1984] associated with rhyolitic and other silicic magmas on Earth, it is anticipated that the enhanced degree of magma fragmentation produced by the low atmospheric pressure on Mars will have led to the more common association of this eruption style with basaltic magmas there [Hougins–Mark et al., 1982; Wilson et al., 1982]. In devising the above list of eruption styles for Venus we have inevitably started from the terrestrial data base of what is known about eruption mechanisms, but we have attempted to minimize the number of assumptions which need to be made about the association between eruption styles and magma chemistry. We now explore the consequences of each of the above types of eruption taking place under the two sets of venusian environmental conditions defined earlier.

4.6. Strombolian Activity

The rise speed of any magma in a crustal fissure system depends on the magma rheology (yield strength and viscosity), the fissure width, and the driving stress (pressure gradient or magma-
STROMBOLIAN ACTIVITY

**EARTH**
- cooling of pyroclastics
- accumulation of pyroclasts in cone
- diffuse, optically thin cloud
- ambient atmosphere
- magma rise rate low
- bubble nucleation depth

**VENUS**
- less cooling of pyroclasts
- denser, optically thick cloud
- coalescence of hot pyroclasts
- lava flow
- less gas expansion, lower pyroclastic velocities

Fig. 7. A comparison of the characteristics of Strombolian eruptions on Earth and Venus. Smaller amounts of gas expansion produce lower pyroclast velocities, denser fire fountains, less opportunity for cooling of clasts, and smaller pyroclast ranges on Venus.

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If the magma exsolves volatiles, gas bubbles nucleate and grow in the magma as a result of both decompression during rise toward the surface and addition of the volatile phase by continuing diffusion. Each growing gas bubble is buoyant in the magmatic liquid and rises through it at a speed which depends on the magma rheology, the bubble size, and the ambient pressure (which controls the gas density). If the rise speed of the magma in the fissure is much greater than the rise speed of the bubbles in the magma, the bubbles remain effectively locked to the magma and do not move appreciably relative to one another: Sparks [1978] showed that these conditions hold for all highly silicic magmas (on all the terrestrial planets). However, if the bubbles can move through the magma at a sufficiently great rate relative to the magma rise speed through the crust, there is time for large bubbles to overtake and coalesce with smaller bubbles. A runaway condition can be reached where single, very large bubbles reach the surface having swept essentially all of the gas out of a long, vertical section of the magma-filled conduit. The eventual bursting of each of these bubbles at the surface is the cause of the explosion (Figure 7). This process can only take place to a significant extent in magmas with relatively low viscosities (≤ 1000 Pa s) rising in narrow (≤ 1 m wide) fissures at low (≤ 1 m/s) speeds [Wilson and Head, 1981]. Furthermore, it is a necessary condition for this kind of activity that the magma volatile content be small enough to prevent complete disruption of the magma into pyroclasts (i.e., smaller than the values given in Table 3); otherwise, the eruption style is by definition Hawaiian or Plinian.

Explosions of this kind on Venus have already been examined in some detail by Garvin et al. [1982], who showed that coalescence may produce bubbles with sizes up to at least 10 m. The excess pressure in these large bubbles, together with their appreciable rise speeds through the magma itself, cause them to blow off a layer of magma trapped between the upper boundary of the bubble and the free surface of the magma in the vent at speeds of up to 10-30 m/s. The sizes of the pyroclasts so produced can range from some meters (if the magma volatile content exceeds about 0.5 wt %) down to millimeters or less. The coarse fraction (clasts larger than 0.3 m) will decouple quickly from the upward gas flow since their terminal fall velocities in the rising and decelerating gas will be comparable to the upward gas speed; the ejection ranges of such clasts are likely to be no more than 100 m.

Clasts smaller than about 10 mm can be entrained into a convecting eruption cloud over the vent if such a cloud forms. If, as discussed earlier, the explosion repetition rate is great enough to produce such a cloud, the dispersal of small pyroclasts will depend on the height of the cloud (and the ambient wind conditions). Since intermittent explosive activity of this type can only occur when magma rise speeds are less than about 1 m/s in conduits or fissures having widths of order 1 m, it follows that typical magma discharge rates in this kind of activity should range up to 2 x 10^3 kg/s (for circular conduits) or 3 x 10^3 kg/s per meter of horizontal fissure length (for elongate fissures). In most terrestrial eruptions, individual active sections of fissures exhibiting this kind of activity are no more than a few tens of meters long, so that the total discharge rate of magma giving rise to an individual eruption cloud is not likely to be more than 10^5 kg/s. Furthermore, analyses of the gas to pyroclast mass ratio in terrestrial Strombolian eruptions shows that only a few percent of the total magma reaching the surface is expelled in the explosive events [Blackburn et al., 1976], so that...
the mass flux into the eruption cloud providing the heat for convection is likely to be of order 3 \( \times 10^7 \) kg/s. If we assume that similar geometrical conditions are applicable to Venus, then equation (29) indicates that eruption clouds over Strombolian explosion sites could have heights up to about 1100 m and thus be capable of forming pyroclastic cones with diameters of the same order consisting mainly of scoria with sizes up to a few millimeters. If we do not accept that terrestrial fissure system geometries may be applicable to Venus, we can get some idea of an upper limit on the sizes of pyroclastic deposits formed in explosions of this kind by arbitrarily raising the percentage of pyroclasts taking part in the convection process to 100\% and increasing the lengths of active fissures feeding individual clouds by a further factor of 10. The fourth root dependence of cloud height on heat release rate means that even though we have now increased the mass discharge rate into the cloud to 10 kg/s, the expected cloud height and hence cone diameter is only 4.2 km. We regard this as a generous upper limit on the possible size.

**4.7. Hawaiian Activity**

If most of the fragments or clots into which a steadily discharging magma disrupts are too large to be readily entrained into a convecting eruption cloud, the clots follow largely ballistic trajectories and form a fire fountain over the vent. The motion is close to ballistic because the steady nature of the discharge ensures that the atmosphere is displaced from the region of the vent; some entrainment by atmospheric gases takes place, but this only affects the outer parts of the ground. As a result, the maximum height of the fountain, \( H_f \), can be expressed in terms of the mean eruption velocity through the vent, \( V_v \), using simple energy conservation, so that

\[
H_f = \frac{V_v^2}{2g} \tag{31}
\]

and the maximum range of clasts, \( R_m \), is given by

\[
R_m = \frac{V_v^2 \sin 2\theta}{g} \tag{32}
\]

where \( \theta \) is the maximum angle from the vertical at which the stream of gas and entrained pyroclasts emerges at the edge of the vent.

If the fire fountain is optically thick, which will be the case if the discharge rate is large enough and the grain size of the magma clots is sufficiently small [Wilson and Head, 1981], little or no cooling can take place on the way. Clots, and so on landing, they will coalesce to form a lava flow or lava lake around the vent. If the fountain is optically thin, some cooling will occur for all clots, the amount now being a function of the grain size and the flight time (and hence \( V_v \)). The result will be an unwelded scoria cone, a partly or completely welded deposit, or formation of a lava flow. The parameters required to characterize this type of activity are, therefore, the magma mass discharge rate and the velocity and spread angle from the vertical of pyroclastic fragments emerging from the vent. Wilson et al. [1980] and Wilson and Head [1981] have described computer programs which can be used to find the exit velocity and maximum spread angle from the vertical of pyroclasts produced in the steady explosive eruption of any magma rheology:

- Volatile content combination given the mass discharge rate and the total volatile content of the magma.
- We have used these programs to calculate the exit velocity and maximum spread angle for a wide range of permutations of magma rheology and mass flux using both CO\(_2\) and H\(_2\)O as the volatile species and assuming a dry tholeiitic as the magmatic liquid.

The size of the smallest pyroclast that can avoid being entrained into a convecting eruption cloud over the vent can be estimated by making some simple assumptions about the ratio of the speed of the magmatic gas stream, \( V_g \), as it emerges from the vent and the terminal velocity, \( V_t \), of such a pyroclast in the gas stream. For example, a clast may be regarded as entrained if \( V_t \) is of order 0.2\( V_g \), since the upward speed of the clast relative to the ground is then 80\% of the mean gas stream velocity. \( V_t \) can be expressed in terms of the diameter \( d \) and density \( d_c \) of the clast (assumed for convenience to be spherical) and the density \( d_g \) of the magmatic gas [Wilson, 1976]:

\[
V_t = \left( \frac{4\pi d^2}{3d_c} \right)^{1/2}
\]

and so, putting \( V_t = 0.2V_g \), we have

\[
\phi = \frac{0.03 d_g V_v^2}{d_c g} \tag{34}
\]

All of the pyroclasts into which the magma disrupts which are significantly greater than the value of \( \phi \) given by the expression will participate in the formation of a fire fountain over the vent; all those which are significantly smaller will be dispersed by a convecting eruption cloud, assuming that the cloud satisfies the stability criterion defined earlier.

There is no simple way of estimating the size distribution of the pyroclasts into which an explosively erupting basaltic magma disrupts. Wilson and Head [1981] argued that the largest clasts would have sizes comparable to the sizes of the largest gas bubbles in the magma, on the grounds that these clasts would be derived from the liquid septs isolated by the interconnection of the gas bubbles as the magma disrupted. Carvin et al. [1982] have calculated gas bubble sizes in Venusian magmas as a function of magma rise speed and total gas content, and we have used their results to estimate the maximum pyroclast sizes likely to be produced in each of our model eruptions. We have then compared the maximum size with the value of \( \phi \) given by equation (34) for the corresponding value of \( V_v \) to see if a significant fraction of the erupted clasts should be involved in forming a fire fountain.

The results of these calculations are illustrated in Figures 8 and 9. Figure 8 shows the differences between conditions on Earth and Venus in schematic form, while Figure 9 shows the dis-
HAWAIIAN ACTIVITY

EARTH

VENUS

narrower, lower, hotter fountain

lower pyroclast speeds & ranges

less gas expansion

Fig. 8. The characteristics of fire fountains produced in Hawaiian-style eruptions on Earth and Venus. For magmas with similar rheological properties and volatile contents, gas bubble nucleation will commence at shallower depths on Venus than Earth; there will be less gas expansion prior to fragmentation of the magma into pyroclasts and also less gas expansion as the gas expands to atmospheric pressure. Pyroclast velocities will be less on Venus than Earth, and clasts will be ejected on paths that are more nearly vertical, leading to smaller ranges, more densely crowded fire fountains, and pyroclasts which are hotter on landing on Venus.

persal of coarse pyroclastic material on Venus. For each of the combinations of magma volatile composition and ambient surface pressure, contours are drawn (solid lines) showing a series of maximum ballistic pyroclast ranges as a function of magma volatile content (left-hand axes) and mass eruption rate (lower axes). The contours are labeled with the diameters of the resulting deposits in meters. Also shown are the minimum magma gas contents at which explosive eruptions can occur (indicated by the horizontal lines below which the contours are truncated) and the boundaries (dashed lines) of the sets of conditions under which fire fountains are likely to form. For mass flux-volatile content combinations to the left of the dashed lines, a significant proportion of the pyroclasts will be coarse enough to escape entrainment into an eruption cloud and will form fire fountains. For sets of combinations to the right of these boundaries, essentially all pyroclasts will be swept up into eruption clouds and will be deposited over a much wider area. The diameters (in kilometers) of the areas within which most of the resulting air fall deposits will lie are shown along the upper axes and have been calculated from the mass fluxes using equation (29) with the same assumptions about eruption cloud shapes as were made for Strombolian eruptions. These air fall deposits will consist entirely of cool pyroclasts for combinations of eruption conditions well to the right of the dashed line boundaries; however, for conditions near the boundaries, the near-vent parts of the deposits will include clasts which have not had time to cool appreciably and so will be partly or completely welded. When they form, the welded zones will probably have diameters in the range 2-4 km.

Figure 9 reveals some striking features of the deposits from the Venusian version of Hawaiian fire fountain activity which are the consequence of the high atmospheric pressure. First, the amount of gas decompression during the passage of the magma through the uppermost few tens of meters of the conduit system is so much less on Venus than on Earth that except at extremely high gas contents and mass fluxes, the mixture of magmatic gas and pyroclasts emerges from the vent as a jet with almost vertical sides; the range of angles from the vertical at which pyroclasts leave the vent is typically a factor of 10 less than on Earth with a similar reduction in the resulting
Fig. 9. The sizes of pyroclastic deposits produced by ejecta from Hawaiian fire fountains on Venus, shown as a function of magma volatile content (left-hand axes labeled in weight percent) and mass eruption rate of magma (lower axes labeled in kilograms per second). The horizontal line in each diagram shows the minimum volatile content for which explosive activity of this type can occur for the stated combination of volatile composition and planetary surface pressure. The solid contours show the ranges of ballistic ejecta and are labeled in meters; however, ballistic ejecta are only likely to dominate the deposits for combinations of eruption conditions to the left of the dashed line in each diagram. For all other conditions most of the ejecta will be entrained into convecting eruption clouds, and for these cases the upper axes show the diameters (in kilometers) of the resulting air fall deposits.

ranges of the clasts which follow ballistic trajectories. Second, the range of conditions (mass fluxes and volatile contents) under which fire fountains can form at all is much smaller than on Earth; the density of the magmatic gas is so high on Venus even after it has decompressed to the local atmospheric pressure that it is capable of sweeping much coarser clasts into a convecting eruption cloud than would be possible at similar eruption speeds on Earth. Eruption speeds are, in fact, smaller on Venus than on Earth for the same magma volatile content just because of the reduced amount of gas expansion, but the higher gas density much more than compensates for this factor. As a consequence of the combination of these two effects, the maximum distances from the vent which can be reached by clasts which form fire fountains are calculated to be extremely small, nominally only a few meters in all cases. No doubt there will be enough irregularities in the walls of real vent systems to produce a greater spread of pyroclast launch angles than is implied by the idealized calculations, but it is nonetheless regarded as likely that ballistic pyroclasts will land within at most one vent radius of the edge of the vent. When the implied eruption parameters are inserted into the equations given by Wilson and Head [1981] for the opacity (and hence heat retention capability) of fire fountains, it is found that in all cases where a fountain can form on Venus, the magma clots within it will suffer negligible cooling during flight and will coalesce on landing to form a lava flow or a pond around a vent.

4.8. Plinian Eruptions

We saw in the previous section that explosive eruptions of basaltic magmas on Venus at mass eruption rates greater than about $3 \times 10^2 \text{ kg/s}$ (and at lower mass fluxes if the magma gas content is very high) will inevitably lead to conditions in which most or all of the pyroclasts produced are entrained into a convecting eruption cloud. The same will be true of eruptions of silicic magmas on Venus (assuming these occur) since, as is the case on Earth, the high viscosity of silicic melts reduces the opportunities for gas bubble coalescence and increases the likelihood of thorough magma disruption [Sparks, 1978]. Progressively higher eruption clouds will be produced at progressively higher mass eruption rates, as indicated by equation (29), and the air fall deposits from these clouds will form the Venusian equiva-
Fig. 10. Parameters relating to Plinian-type eruptions on Venus for various combinations of magma type, magma-volatile type, and surface atmospheric pressure. The velocity of gas and small pyroclasts in the vent (left-hand axes, labeled in meters per second) is shown as a function of mass eruption rate of magma (lower axes, labeled in kilograms per second), and magmatic volatile content (solid curves, labeled in weight percent of the stated volatile). The lowest curve in each case shows the smallest magma-volatile content that will allow Plinian activity to take place. Eruption cloud heights along the upper axes are labeled in kilometers and are determined by the mass fluxes shown on the lower axes. Stable convecting eruption clouds can only form for combinations of eruption conditions lying above the dashed line in each diagram; for other conditions, pyroclastic flows will be formed instead. The sizes of the largest vesicular (density 1000 kg/m$^3$) pyroclasts that could be transported out of the vent, shown along the right-hand axes labeled in meters, are determined from the eruption velocities shown on the left-hand axes. The horizontal bars in three of the diagrams indicate ranges of mass eruption rates over which pyroclastic flows can never be formed on Venus in steady Plinian eruptions.

Elements of strong Strombolian, sub-Plinian, and Plinian deposits, differing from their terrestrial counterparts only in the details of their areal extents and grain size distributions. We have used the calculations of eruption velocity as a function of magmatic volatile content and mass discharge rate described in section 4.7 to define the conditions at the bases of the resulting eruption clouds and have then employed the methods given by Wilson [1976] and Sparks and Wilson [1987] to estimate the heights, shapes, and internal velocity fields of the resulting clouds. Figure 10 shows the eruption velocity of gas and small pyroclasts in the vent (left-hand axes) plotted as a function of mass discharge rate (lower axes) and total magmatic volatile content (solid curves labeled in weight percent) for the three magma-volatile combinations and two sets of environmental conditions defined in Table 3. The volatile content curve labeled with the lowest gas content is in each case the curve corresponding to the smallest volatile content which will allow explosive activity to occur at all. Eruption velocities are typically a factor of 2 less on Venus than those on Earth (given by Wilson et al. [1980], L. Wilson [1980], and Wilson and Head [1981]) as a result of the smaller amount of expansion which the magmatic gas must undergo to reach ambient atmospheric pressure. On the right-hand axes are indicated the diameters of the
largest vesicular (bulk density = 1000 kg/m$^3$) pyroclasts which can be transported out of the vent by the gas flow speeds give on the left-hand axes. The pyroclast sizes, which are typically a factor of 20 larger than those which would be involved on Earth for a given magma volatile content, are calculated using the appropriate atmospheric pressure and magma volatile species in each case. They represent the largest clasts which could be erupted if they were present. We stress that these large clasts (sizes up to many tens of meters are allowed) may well not be present; indeed, for some combinations of magma gas content and mass eruption rate, the predicted maximum clast size is larger than the width of the conduit through which the eruption would be taking place. In such cases the conduit width would of course set the upper clast size limit, and any attempt to eject clasts with sizes approaching the limit would in any case lead to strong, transient nonuniformities in the eruption conditions [Wilson et al., 1980].

On the upper axes of Figure 10 are plotted the heights of convecting eruption clouds (when these can remain stable) corresponding to the mass discharge rates on the lower axes. The dashed curve in each part of the figure is the stability criterion for convection to occur, taken from equation (28) of Wilson [1976]. For all sets of conditions above this line and a collapsed cloud, or fountainlike, structure, of the kind assumed to act as the source of large-scale pyroclastic flows, will form for all sets of conditions below it. It will be noted that both kinds of activity can occur at all mass eruption rates and at all eruption temperatures and all eruption cloud heights of approximately 50 km; we have used a mean wind speed of 3 m/s to calculate the downwind transport of clasts, a value which may well be something of an overestimate [Counselman et al., 1980]. In an attempt to give an impression of the maximum extent of a Plinian air fall deposit on Venus we have simulated a cloud arising from an eruption in which the magma is a rhyolite exsolving 7 wt % H$_2$O and the mass eruption rate is $10^{15}$ kg/s, yielding a cloud height of 65 km and a mean wind speed of 3 m/s. Examination of Figure 11 shows the variation of maximum grain size with downwind distance from the vent. It is clear that the fine-grained parts of air fall deposits may extend for distances well in excess of 100 km from the vent. However, clasts with sizes of the order of a few hundred millimeters, capable of causing noticeable roughness signatures at the wavelength used by currently employed radar systems, will be confined to regions extending no more than 15-20 km downwind of the vent. Figure 12 shows the variation of maximum grain size with position in the deposit resulting from the above model calculation. Few Plinian deposits on Venus should have a more extensive dispersal of clasts than that shown in Figure 12.

It has recently been suggested [Esposito, 1984] that sulfur dioxide levels in the middle atmosphere of Venus may be fluctuating on a time scale of decades as a result of episodic volcanic injections of this gas. The proposed mechanism would require convecting eruption clouds to reach a height of approximately 50 km. Examination of Figure 10 shows that there is always an upper limit to the height that can be reached by an eruption cloud produced in an eruption having any particular magma-volatile combination: if the mass flux increases beyond the limit corresponding to the critical cloud height, convective instability sets in, a collapsed eruption cloud forms, and it is inferred that pyroclastic flow production begins. Cloud heights of 50 km can in fact be reached by all of the magma-volatile combinations shown in Figure 10, though the magma must exsolve at least 4.5, 7.5, or 5 wt % gas for the water-basalt, water-rhyolite, and CO$_2$-any magma combina-
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Fig. 11. Some properties of a 45-km-high Plinian eruption cloud on Venus produced by an eruption with a mass flux of $10^{16}$ kg/s, a magmatic gas content of 7 wt% H$_2$O and an eruption velocity in the vent of 400 m/s. (a) The variation of upward gas speed on the cloud centerline as a function of height above the vent. (b) The size of the largest pyroclast (bulk density 1000 kg/m$^3$) which could be supported in the cloud as a function of height. (c) The maximum pyroclast size which would be found in the air fall deposit as a function of distance from the vent measured in the downwind direction with a mean wind speed of 3 m/s.

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tions, respectively (there is only a weak dependence on the surface atmospheric pressure). Clearly, we need to generate the detailed relationships shown in Figure 10 for SO$_2$ as the magmatic volatile; however, there are too few data available on the solubility of SO$_2$ in magma to make this possible [Carroll and Rutherford, 1984].

As a compromise, we have carried out the calculations for SO$_2$ assuming that its solubility in magmas is the same as for each of the above three magma-volatile combinations in turn. We find that the greatest cloud heights occur if a solubility function like that of H$_2$O in basalt is assumed; however, even with this assumption it is not a trivial matter to generate eruption clouds extending up to 50 km. Figure 13 shows the maximum height that can be reached before pyroclastic flow formation begins as a function of the exsolved magmatic volatile weight fraction for SO$_2$ (using the H$_2$O solubility function), CO$_2$ and H$_2$O (using basalt as the magma type) at both the 40- and 100-bar surface pressure levels on Venus. Volatile contents of at least 12 wt% of pure SO$_2$ are needed to enable an eruption cloud to reach 50 km. A simple way of transporting SO$_2$ to the required height without involving such large weight fractions is to carry it as a minor component in a cloud driven by H$_2$O, CO$_2$, or some other volatile, but the consequences of the high-level emplacement in the atmosphere of the other volatile species would then need to be explored, of course.

4.9. Vulcanian Explosions

Following Self et al. [1979], we have associated this eruption style with a buildup of gas pressure in the upper part of a magmatic system as a result of gas accumulation under a plug or cap blocking the vent. The presence of the plug may commonly be the result of the intrusion of magma to shallow depths where excessive cooling halts its progress. The strength of the plug may be quite high: modeling of terrestrial Vulcanian explosions suggests values up to 200-300 bars in tension [Self et al., 1979], these values relating to the part of the magma which has cooled sufficiently to acquire long-term strength but has not reached the stage of containing many large-scale brittle fractures due to cooling stresses.

There is no simple way of predicting an upper limit to the ratio of the masses of gas and clastic material which may be involved in Vulcanian explosions. If exsolving magmatic gases accumulate in situ in a trapped magma, the gas solubility arguments presented earlier restrict the gas weight fraction to considerably less than 10% for all magma-volatile combinations (on all the terrestrial planets). If the heat from a magma vaporizes volatiles trapped in the rocks surrounding it or produces gases by thermal decomposition of some of the country rocks (as might happen if, for example, carbonate or sulfate deposits were present), then higher gas fractions might in principle be produced [Luhr et al., 1984]. The weight fraction of trapped volatiles that can be invoked in this way is limited by the pore space of the host rocks: if we assume a generous value of 20 vol % for the pore space, filled with, say, liquid sulfur (which, as we noted earlier, is a thermody-

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Fig. 12. Dispersal pattern of pyroclasts and variation of maximum grain size with distance for a Plinian eruption on Venus. Maximum range in the air fall deposit is shown for a range of sizes of clasts having bulk density 1000 kg/m$^3$. Eruption conditions as for Figure 11.
We associate the Pelean eruption style with the explosive disintegration of a body of magma emplaced at a high level in a volcanic system: as a dome on the surface or as a near-surface intrusion. The explosive event may be initiated by the failure of an outer, cooled carapace as a result of a buildup of gas pressure or by the unloading of the magma body due to a landslide. The main difference between the Pelean and Vulcanian eruption styles appears to be that the former involves almost entirely juvenile material and magmatic volatiles, whereas the latter involves a substantial proportion of country rock or nonmagmatic volatiles or both. Both styles share the common feature that they involve unsteady motions as the eruption products are first accelerated up to a maximum speed by the expansion of the released volatiles and then decelerate again as a result of frictional interaction with the atmosphere (in the approximate and geochemically plausible candidate on Venus), having a density close to one third that of the rock, then complete vaporization of the sulfur will yield gas mass fractions of about 50% and 30%, respectively; again, these figures would be reduced when expressed in terms of the total mass of magmatic material involved. Also, if large masses of volatiles are produced from country rocks by either of these two mechanisms, consideration of the heats of vaporization and decomposition shows that a very large fraction of the magmatic thermal reservoir will be used up, reducing drastically the kinetic energy produced in the eventually resulting explosion. We conclude that all combinations of conditions likely to be encountered in Vulcanian explosions (again, on any terrestrial planet) can be more than adequately represented by assuming volatile mass fractions in the explosion products up to 30% and magmatic temperatures in these materials prior to a buildup of gas pressure or by the unloading of the material away from the vent in a transient eruption cloud during activity of this kind would be less efficient on Venus than on Earth due to the lower eruption cloud heights, already discussed at length in connection with Strombolian and Plinian activity.

4.10. Pelean Eruptions

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A. BEFORE EXPLOSION

B. DURING EXPLOSION (EARLY STAGE)

C. DURING EXPLOSION (LAST STAGE)

Fig. 14. (a) The geometry assumed for a cone being invaded by fresh magma prior to a Vulcanian explosion on Venus. (b) Conditions during the early phase of acceleration of the explosion products. (c) Conditions late in the explosion process.

When upward directed Pelean explosions produce convecting eruption clouds on Venus, these will rise to about 60% of the height of a cloud of the same mass on Earth, as discussed in earlier sections. Without estimates of the masses involved in such explosions it is hard to give a quantitative estimate of cloud rise heights and pyroclast dispersal areas. Unfortunately, little work has yet been done on the stability of structures such as extrusive domes and near-surface intrusions. It is clear that relationships must exist between such parameters as magmatic volatile content, magma extrusion or intrusion rate, cooling rate of the outer skin of the magma body, and accumulation rate of trapped gases, but we do not have even empirical information about these relationships for terrestrial cases. We note, however, that the widths and thicknesses of domes involved in Pelean activity on Earth generally range up to at most a few hundred meters, implying masses of order $10^{11}$ kg and that if all this material is injected into a transient eruption cloud, its rise height on Venus as given by equation (30) will be about 10 km, leading to an air fall deposit with a diameter of the same order.

Pelean explosions on Venus could produce pyroclastic flows, either by collapse of upward directed eruption clouds or by direct generation of laterally directed clouds. The run-out distances of flows from collapsing eruption clouds will be reduced relative to those on Earth by the same amount as flows produced in Plinian activity. The Venusian environment will also lead to a reduction in the run-out distances of flows produced in lateral blast explosions. This is a result of the effect noted for Vulcanian explosions, whereby the ratio of the absolute pressure in the compressed magmatic gases driving the explosion to the atmospheric pressure which they will ultimately reach after expansion is much less than in the corresponding events on Earth. We have modeled the maximum velocities likely to be reached by lateral blast clouds using the formulation given by Eichelberger and Hayes [1982] in which the magmatic material is accelerated by an expansion wave which propagates into it from the free surface and decelerated by a compression wave which forms between the expanding debris cloud and the undisturbed atmosphere. Table 4 shows values of the peak velocities of such clouds for three magma volatile contents on Earth and Venus (at the 40- and 100-bar pressure levels) using H$_2$O as the magmatic volatile. If CO$_2$ were used, the velocities would be about two thirds of those generated by H$_2$O: the velocity scales approximately as the reciprocal of the square root of the volatile molecular weight. Also shown in Table 4 are values for the speeds of acoustic waves within the blast cloud and in the surrounding atmosphere. It will be noted that terrestrial lateral blast clouds propagate at speeds which are potentially so high that the initial stages of their acceleration are in fact controlled by choking in the vent and involve the generation of complex patterns of shock waves, as modeled numerically by Kieffer [1981a, b]; however, this complication may well be absent for such events on Venus, since cloud velocities are so much less there. If we assume that the run-out distances of blast clouds are proportional to their initial kinetic energies (this was the assumption made earlier for pyroclastic flows from Plinian-type eruptions) and so to the squares of their velocities, comparison of the velocities in Table 4 suggests that such clouds should travel at least 10 times less far from the vent on Venus than on Earth, restricting their ranges to at most a few kilometers.

5. Volcanic Deposits and Landforms on Venus

5.1. Summary of Predictions

In previous sections we have outlined the nature of effusive and explosive eruptions in the
present environment on Venus and have developed the basis for understanding the nature and distribution of deposits from a variety of eruption styles. As a first step toward the interpretation of volcanic landforms on Venus, we outline a summary of the sizes of features and deposits predicted by our analyses (Figure 16). On the basis of these analyses, the size and geometry of landforms and deposits can be predicted (Figure 17). Lava shields with diameters up to 20 km would be relatively easy to produce from central vents in the Venus environment. Construction of larger (wider) shields would require effusion rates higher than those common on Earth (lunarlike, for example) or eruptions from multiple vents. Stratovolcanoes produced from Vulcanian or Pelean activity could commonly have diameters up to 10 km and some could possibly extend up to 20 km in diameter. Ignimbrites and air fall mantles could cover zones with long axes or diameters up to 100–200 km (Figure 16). Even if there is no primary volatile enhancement in the magmas, bubble coalescence in Strombolian activity provides a mechanism to produce episodic volatile enhancement and localized pyroclastic activity and deposits. Low effusion rates and protracted activity could produce basaltic cones up to 8–10 km in diameter (Figure 17). Central pit size should be dictated by ballistic transport range and should be less than a hundred meters in diameter. Cones could also result from primary volatile enhancement and high effusion rates in Hawaiian activity. Such cones would have less distinct edges, finer-grained material near the vent, and the possibility of larger central pits related to collapse of a near-surface reservoir. Strombolian cones could be distinguished from Hawaiian cones by their more distinct edges and their smaller summit pits. Single Vulcanian eruptions could produce deposits up to the 10-km diameter range, but many successive eruptions would be required to build a cone (Figure 17). More evolved silicic magmas could produce cones up to 10–16 km in diameter (Figure 17) from air fall and pyroclastic flow deposits. More silicic magmas would be characterized by higher viscosity, more chilling during emplacement, and lower effusion rate, with more intermittent eruptions producing Pelean activity and localized air fall and pyroclastic flow deposits. The range of these deposits would be rather small (5–8 km), and cones could build up from repeated activity. If the slopes of the cone exceeded the angle of repose, the base of the cone could widen beyond the general limits dictated by radial ranges of eruption deposits (Figures 16 and 17). When air fall and pyroclastic flow deposits occur together with lava flows of more evolved magmas in stratovolcanoes, the magmas will tend to be less viscous than terrestrial analogs because of the higher eruption temperatures and the reduced amounts of volatile loss, and the general proportion of flows to pyroclastics will be greater on Venus than on Earth. This will tend to produce landforms whose lateral dimensions are dictated by flow length (Figures 16 and 17) and whose vertical dimensions are governed by the number of events and the proportion of pyroclastic deposits.

The sizes of the summit craters formed in cones produced by the Vulcanian or Pelean activity on Earth tend to be controlled by the distribution of stresses in the region surrounding relatively chilled magma in the vent and are thus commonly much larger than the size of the conduit feeding the eruption. A similar pattern is expected on Venus, and by analogy with what is observed in terrestrial cases, we anticipate craters with diameters up to about 100 m in cones formed by Vulcanian activity and diameters up to at least several hundred meters in Pelean cones. The 1980 eruption of Mount St. Helens demonstrated that when lateral blast activity is involved, summit craters larger than 1 km can be produced by what is essentially a version of Pelean activity.

5.2. Observations and Analysis of Venus Volcanic Features

The predictions outlined above can be compared to surface features observed on Venus from the Pioneer Venus mission [Pettengill et al., 1980; Mauzerly et al., 1980], from the Venera lander and orbiter missions [Florensky et al., 1982, 1983a, b; Garvin et al., 1984; Barsukov et al., 1984a, b, 1986], and from images derived from Earth-based radar observations [Campbell et al., 1977, 1984; Head et al., 1985a]. Several areas are of particular interest because of the distinctive features that have been observed (Figure 1).

Flows and shields. Recent high-resolution radar images of Beta Regio [Campbell et al., 1984;
TABLE 4. Values of the Maximum Velocities of Lateral Blast Clouds on Earth and Venus

<table>
<thead>
<tr>
<th>Magma Water Content, wt %</th>
<th>Maximum Velocity of Blast Cloud, m/s</th>
<th>Acoustic Speed in Blast Cloud, m/s</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Earth (Sea Level)</td>
<td>4-MPa level</td>
</tr>
<tr>
<td>10</td>
<td>548</td>
<td>155</td>
</tr>
<tr>
<td>5</td>
<td>443</td>
<td>125</td>
</tr>
<tr>
<td>3</td>
<td>374</td>
<td>105</td>
</tr>
<tr>
<td>Atmospheric</td>
<td>350</td>
<td>412</td>
</tr>
<tr>
<td>Acoustic Speed</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Also given are the acoustic wave speeds within the blast material and in the surrounding atmosphere.

Stofan et al., 1985] have confirmed earlier proposals [McGill et al., 1981, etc.] that Beta Regio is the site of a major rift zone and that two centrally located topographic prominences (Rheia and Theia montes) are of volcanic origin. Theia Mons (Figures 1a, 1b, and 1c) appears as an approximately circular radar bright feature averaging 325 km in diameter and rising over 5 km above mean planetary radius and approximately 1.5-2.5 km above the surrounding crest of Beta Regio [McGill et al., 1981; Campbell et al., 1984] (Figure 1c). A smaller, irregular, radar dark feature 60-90 km in diameter is centrally located on the bright spot. This topographically distinct radar feature is located along the major western bounding fault of the central Beta Regio rift system and appears to be superposed on the structure, filling in the deep chasm in this area. The central dark area may be a volcanic caldera containing relatively smooth flows, a situation common in terrestrial volcanoes, although the density of topographic data is insufficient to establish the existence of a summit depression. The dark central region is surrounded by a halo which has a relatively rougher surface or an enhanced dielectric constant (or both). This region could consist of air fall materials or flows of either lava or pyroclastic origin. Farther down the flanks are observed features which extend approximately radially (downslope) away from the summit, particularly to the northwest, and from their shapes they show signs of topographic control (Figures 1a and 1b). The edges are very lobate and are much more suggestive of the margins of lava flows than of ignimbrites. The most distinctively lobate features are 25-75 km in width and extend on the average about 250 km from the edge of the radar diffuse zone to the northwest of Theia, about 375 km from the edge of the bright region and about 550 km from the summit. The geometry of the mapped units suggests that some of the lobate flows may originate in the vicinity of the diffuse zone on the west and northwest flank of Theia.

High-resolution radar images have also been obtained of a region south of Ishtar Terra in northern Sedna Planitia which contains extensive volcanic deposits (Figure 1d) [Head et al., 1985a]. The volcanic plains consist of patches of radar dark, bright, and intermediate terrain often with lobate and flowlike boundaries. In the central western portion of the plains occurs a 200-km-diameter intermediate-brightness patch with a 50-km dark central area that has been interpreted to be a volcanic source area and possibly a low shield area. Approximately 250 km to the north occurs a second low shieldlike structure of intermediate radar brightness and approximately 200-350 km in diameter. In this case, the central area is dark and about 100 km wide and contains an inner, irregularly shaped region about 50 km in diameter, which may be a caldera. This region is characterized by several flowlike features about 10-30 km in width and extending at least 200 km from the apparent source at the center of the structure. These elongate and lobate flowlike features then merge into broad units surrounding the structures (Figure 1d). A third occurrence of elongated flowlike features has been reported by Barsukov et al. [1984a, 1986] for the Lakshmi Planum area as observed on Venera 15/16 images. Here, a system of flows up to 100-300 km in length emanates from the vicinity of Colette, an apparent caldera in central western Lakshmi Planum. Flow lengths in the 100-200 km range are also reported by Barsukov et al. [1984a, 1986] for other parts of the northern hemisphere of Venus, with flow widths in the 10-20 km range.

To summarize, on the flanks of Theia Mons the lava flow features have widths averaging about 50 km and lengths of up to 300 km if they emerge from the middle of the diffuse zone or 550 km if they emerge from the summit. If the bright zone represents a carapace of rough flows, their widths are not known, but lengths should not exceed about 160 km. All of these flows have formed in areas where the regional slope measured from the Pioneer Venus orbiter radar altimetry [Pettengill et al., 1980; Sharpton and Head, 1985] averages 0.007 rad. In the region south of Ishtar Terra, similar features have average widths of 20 km and lengths of at least 200 km on regional slopes of 0.002 rad. The flows described by Barsukov et al. [1984a, 1986] surrounding Colette crater on Lakshmi Planum average 150 km in length and 15 km in width and occur on regional slopes of 0.01 rad.

For these sets of features, therefore, we can regard an average value of each of \( W_t \), \( X \), and \( \alpha \) as known. A range of possible rheological properties and effusion rates can be calculated using equations (19) to (27) and some simple assumptions. We choose a series of pairs of values of \( Y \) and \( \eta \) from the range specified by the curve in Figure 2c, and for each value of \( Y \) we calculate \( D_b \) from equation (19), \( W_b \) from equation (20), and \( D_t \) from equation (23). We obtain a value of \( \eta \) for each \( Y, \eta \) pair by subtracting \( 2W_b \) from \( W_t \) and use this...
to find $F$ from equation (21). Finally, we evaluate $A$, $u$, $X$, and $\tau$ in turn. The calculated values of $X$ found in this way can be compared with the average value observed in the radar images. Clearly, we cannot accept any $Y_\eta$ pair for which the calculated value of $X$ is less than the observed value (unless we wish to assume that lava tube formation is a sufficiently common process that equation (26) always understimates flow lengths). However, any $Y_\eta$ pair for which the calculated value of $X$ is greater than that observed represents a possible solution for the lava rheology, since flows commonly cease to advance (due to exhaustion of the magma supply) before they reach their maximum permitted length. The value of $F$ corresponding to the $Y_\eta$ pair for which the observed and calculated values of $X$ are equal can then be taken as the minimum plausible value of lava effusion rate for the observed flow, while the values of $Y$ and $\eta$ for this condition can be regarded as lower limits on the rheological parameters.

Table 5 shows the results of this kind of analysis. It is clear that the flows near Colette caldera imply eruption rates close to $6 \times 10^6$ m$^3$/s; the flow features to the south of Ishtar Terra require eruptions having values of $F$ greater than or equal to about $5 \times 10^6$ m$^3$/s, while those on Theia Mons would have to involve higher effusion rates, approaching $4 \times 10^5$ m$^3$/s. These latter values are more typical of the high eruption rates inferred for the ancient lunar basaltic flood eruptions than those currently common for basalts on Earth. There are four possible explanations for these results: (1) the features assumed to be flows have been misidentified and the analysis has no meaning, (2) the flows have been assigned widths or lengths which are too large as a result of instances of the combination of multiple flows into a single map unit, (3) the combinations of rheological parameters implied by Figure 2c are not appropriate to the magmas involved in the observed flows, and (4) the values are correct and high effusion rate basaltic eruptions are common on Venus. We regard explanation 1 as very unlikely in the light of our earlier discussion of the identification of flow features. Under explanation 2 it is possible that the flow widths have been overestimated as a result of the failure to recognize the boundaries of adjacent flows, but it is less likely that flow lengths have been affected in this way (unless, perhaps, compound flows have built up as a result of breakouts of lava from the flow fronts of earlier emplaced flow lobes); smaller values of either $W$ or $X$ would imply lower eruption rates than the values given above. Option 3, though a possibility, would probably not lead to very great changes in implied effusion rates: for example, if the viscosity associated with each yield strength value in Figure 2c were to be increased by a factor of 10, the implied effusion rates would only be approximately halved. We therefore conclude that option 4, that high effusion rate eruptions are relatively common on Venus, can be well supported. This must reflect some aspect of the tectonic state of the lithosphere in the areas where the long flows are located which encourages the opening of fissures in the crust which are long or wide (or both).

Some of the above ambiguities could be resolved, at least in part, if it were possible to resolve (and recognize) levees on the lava flows, so that values could be determined separately for
**Fig. 17.** Predictions of widths (based on Figure 16) and heights (see text for discussions) of volcanic landforms on Venus produced by various kinds of eruptions.

$W_c$ and $W_b$ in addition to $X$ and $\alpha$. Equations (20) and (21) then allow $Y$ and the product $(F \rho)$ to be determined uniquely; it is still not possible to separate $\eta$ and $F$, however, without assuming that a particular flow or flow lobe has a length $X$ equal to its maximum length permitted by the cooling constraint. If this is in fact the case, then for a flow for which $W_c >> W_b$ and $D_c >> D_b$, it is readily shown by suitable manipulation of equations (21), (24), (25), (26), and (27) that

$$X = \frac{0.84}{270c} \left( \frac{9 \alpha \gamma \beta}{6 \beta \beta \gamma \gamma} \right)^{1/11}$$

Thus, if we find values of $(F \rho) = A$ and $(9 \alpha / \beta)^{1/11} = B$, we can separate out values of $F = (A^2 B)^{1/11}$ and $\eta = (A^2 / B)^{1/11}$. If the approximations $W_c >> W_b$ and $D_c >> D_b$ are not valid, an equivalent but much more cumbersome expression can readily be derived in place of (35) using all of equations (19)-(26). This kind of procedure would always be subject to the uncertainty as to whether a particular flow unit had in fact reached its maximum possible length. It is interesting to note from Table 5 that for the Theia Mons flows, the calculations suggest that $W_b \sim 5 \text{ m}$, so that much higher-resolution radar systems than any currently planned for Venus exploration would be needed to enable us to proceed any further with the analysis of these features. For the flows south of Ishtar, however, the same calculations suggest that $W_b \sim 800 \text{ m}$ or larger, so that an improvement in resolution by less than a factor of 10 over that currently available in the Arecibo radar system would enable levees on these flows to be identified and measured, thus providing important additional evidence on lava effusion rates.

The Venera lander panoramas provide information about a series of sites on the eastern flanks of the Beta-Phoebe Regio area. From the many interesting types of structure visible [Florensky et al., 1982, 1983a, b; Garvin et al., 1984] we focus attention here on the relatively thin (of the order of a few tenths of a meter) fine-grained layers which appear to be continuous for at least several meters at various sites (Figure 18). It has been suggested that these layers may be air fall ash layers or pyroclastic flows [Florensky et al., 1983b] or lava flows [Garvin et al., 1984]. Based on earlier discussions in this paper, it is clearly possible for the deposits to be air fall ash layers: Figures 11 and 12 show that material with mainly submillimeter grain size may be deposited over areas of hundreds of square kilometers from high mass eruption rate Plinian-type events, assuming that volatile contents are high enough to permit explosive eruptions to occur at all. An additional possibility is that the thin layers represent eolianite air fall deposits made up of fine material elutriated from pyroclastic flows by the throughput of atmospheric gases. It is also possible that the thin layers are themselves pyroclastic flow deposits. However, since we do not have a good general model of thickness/travel distance/grain size relationships for terrestrial pyroclastic flows, we cannot predict what kinds of internal details we should look for in these layered deposits to test this possibility. If either of the above two possibilities is correct, the question remains as to how the deposits become lithified: options include welding (possible for a pyroclastic flow, even this thin, but much less likely for an air fall deposit) and chemical reactions with the atmosphere.

The third possibility is that the thin deposits are lava flows. Each layer could be a distinct flow unit or could be part of an internally stratified, thicker flow unit. In the latter case, initially suggested by Florensky et al. [1977] and developed by Garvin et al. [1984], the layering is interpreted to be analogous to that commonly observed in terrestrial lava flows which is a combination of upper thermal boundary layers (crusts) and horizontal sheets formed generally parallel to the top of the flow due to cooling and shearing during flow emplacement. Cross sections of terrestrial lava flows [Hammond, 1974] show that such
structures may be common in the upper 10-30% of individual pahoehoe lava flows. Thus, depending on whether we assume that each thin deposit represents all or part of a flow unit, we need to consider flows from 0.1-0.5 m thick which cover significant areas. Such features are not common on Earth but examples do exist [Greeley, 1974]. In modeling lava flows we have two basic options: (1) the flows are formed on topography approximating an infinite flat plain, so that the width and thickness of the flow units are determined solely by the rheological properties and effusion conditions of the lava itself, and (2) the flows occur on preexisting topography which provides channels for the lava to move in. In this case the rheological properties will determine the thickness and width of a central, unsheared plug, which will always exist if the yield strength is nonzero. In the first case we may assume that the typical layer thickness of 0.1-0.5 m corresponds either to the center of the flow or to some region near the edge, representing the levee thickness, for example. In the second case, the layer thickness may represent the thickness of the plug or the whole flow unit. Since plugs commonly have thicknesses which are between one-third and one-half the total flow unit thickness [Johnson, 1970], assuming that the layers are internal shear or cooling features: the flow unit thickness has to be 4 or 5 times larger than the layer thickness.

We treat both the above possibilities. First, assume that the flows occur on a wide flat plain. The topography and dynamics of the flow are controlled by the relationships embodied in equations (19)-(27). Initial, exploratory calculations were based on a wide range of assumed values for regional slopes and lava rheological parameters. It was found that to generate thin flows with widths on the order of a few tens of meters (required for consistency with the fact that the thin layers in the Venera panoramas extend for at least many meters), it was necessary to invoke regional slopes of 0.01 rad or more (the upper end of the range of values appropriate to the flanks of Theia and Rhea montes) and lava viscosities and yield strengths comparable to or less than those found for the assumed flows on the flanks of Theia Mons. Table 6a shows some flow parameters calculated using $\alpha = 0.01$ rad and levee thicknesses $D_b$ of 0.1 m and 0.01 m. The corresponding values of $Y$ and $\eta$ are given in Table 6a. Suitable flow geometries are produced for volume eruption rates $F$ of about 100 m$^3$/s, which are within the present-day range for basaltic eruptions on Earth.

If we now make the alternative assumption that

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Flanks of Theia Mons</th>
<th>South of Ishtar Terra</th>
<th>Area Described by Barsukov et al. (1984a, 1986)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$Y$, Pa</td>
<td>10</td>
<td>110</td>
<td>30</td>
</tr>
<tr>
<td>$\eta$, Pa s</td>
<td>1036</td>
<td>1497</td>
<td>1230</td>
</tr>
<tr>
<td>$D_c$, m</td>
<td>4.6</td>
<td>9.8</td>
<td>4.3</td>
</tr>
<tr>
<td>$W_b$, m</td>
<td>4.5</td>
<td>6.0</td>
<td>6.1</td>
</tr>
<tr>
<td>$W_c$, km</td>
<td>50.0</td>
<td>18.8</td>
<td>15.0</td>
</tr>
<tr>
<td>$F$, $10^8$ m$^3$/s</td>
<td>29</td>
<td>4.1</td>
<td>5.8</td>
</tr>
<tr>
<td>$D_b$, m</td>
<td>0.064</td>
<td>2.4</td>
<td>0.12</td>
</tr>
<tr>
<td>$u$, m/s</td>
<td>1.8</td>
<td>0.32</td>
<td>1.4</td>
</tr>
<tr>
<td>$X$, km</td>
<td>211</td>
<td>162</td>
<td>131</td>
</tr>
<tr>
<td>$t$, days</td>
<td>1.3</td>
<td>5.9</td>
<td>1.1</td>
</tr>
</tbody>
</table>

See text for definitions of parameters and discussion.

Fig. 18. Venera 14 panorama showing horizontal to subhorizontal layers with thicknesses ranging from a few to tens of centimeters. Edge of the 1.2-m-diameter lander basal ring is seen. Lander ring teeth are 7 cm in length. Photo courtesy of USSR Academy of Sciences.
thin flow units are produced by lava flows moving down valleys defined by the preexisting topography, there is no longer any simple coupling between the rheology, the effusion rate, and the morphology. There is still, however, the usual relationship between the mean velocity $u$ of the moving lava and its depth $D_c$ in the middle of the guiding depression, arising as a result of the balance between the gravity component down the slope and the viscous resistance of the fluid

$$u \approx \frac{D_c^2 g \rho \alpha}{8 \eta}$$

which can be used in place of equation (25) for $u$ to provide a value of the maximum flow length $X$ from equation (26). Table 6b shows some values of $X$ found for $\alpha = 0.01$ and two values of $\eta$ (1200 and 120 Pa s) for values of $D_c$ in the range 0.5-4 m. Again it is possible to produce flows hundreds of meters wide having internal layers with thicknesses of a fraction of a meter.

Although the above calculations show that the layered structures seen at the Venera landing sites can be produced in thin tholeiitic lava flow units erupted at rates similar to those commonly encountered on Earth, they also show that such flows can only extend for distances of at most a few hundred meters (a conclusion reached just on the basis of the rate of growth of the cooled skin by Frenkel and Zaboluyeva [1983]). If one wishes to interpret the layers seen in the lander images as being produced by such flows, therefore, it is necessary to accept that each lauder is located within a few hundred meters of a vent. The distances between the essentially randomly located Venera landing sites are typically a little greater than 2000 km, and so one would be driven to conclude that volcanic vents were spaced no more than about a kilometer apart over an area of more than $2 \times 10^7$ km$^2$. If this interpretation were to be supported by data from future missions, it would have profound implications for the structure of the Venus lithosphere. Such a situation has not been encountered on any of the other terrestrial planets and must currently be considered improbable. We therefore favor the conclusion of Garvin et al. [1984] that the surfaces viewed by the Venera landers in the Beta-Phoebe region represent the partly eroded upper surfaces of lava flows having thicknesses in excess of at least a meter.

**Domes and cones.** Barsukov et al. [1984a, b, 1986] reported on a series of domelike hills which often occur in clusters on the plains in numerous locations on Venus. These features have diameters ranging from the limits of resolution (1-2 km) up to 10-15 km. Occasionally, a summit crater is seen. Barsukov et al. [1986] point out the similarity in morphology of these features to volcanic domes or cinder cones on Earth and Mars. The widespread nature of these cone and domelike features is an indication of the local and regional significance of volcanic activity and the abundance of individual vents on Venus. However, the present level of description and knowledge of these features is insufficient to assign specific eruption conditions to them as a class or even to determine if they represent more than one class of eruption styles. On the basis of our predictions we offer some initial observations and some guidelines for further analysis in order to establish the eruption style or styles represented by these features. As described by Barsukov et al. [1984a, b, 1986], these features could represent edifices dominated by effusive activity, explosive activity, or a combination of the two. Predominantly effusive activity could produce small domelike shield volcanoes similar to those observed in the Snake River Plain on Earth [Greeley and King, 1977] and in the lunar maria [Head and Gifford, 1980]. On Earth and the moon these features range in size up to 15-20 km diameter and several hundred meters in height and tend to occur in clusters. Summit craters are often seen on lunar and terrestrial domes and are systematically larger (relative to dome base) on lunar domes. Summit crater diameters on the moon increase with dome size and are in the 1-3 km range [Head and Gifford, 1980]. Thus the Venus features share many of the characteristics of small shields built by effusive activity on Earth and the moon.

Explosive activity is also capable of building features matching this general description. Strombolian activity on Venus could produce pyroclastic cones with maximum diameters in excess of several kilometers without requiring the presence of more evolved magmas (Figures 16 and 17). Cinder cones produced from such activity are common on Earth and often occur in local clusters and fields in various tectonic environments [Settle, 1979b]. Although cone diameter sometimes exceeds 2 km in terrestrial situations, average cone diameter for a wide range of locations is less than 1 km [Settle, 1979b]. Although Strombolian activity is plausible in the Venus environment, we have no reason to believe that average cone diameters should greatly exceed terrestrial dimensions, which are approximately the limit of resolution of the Venera 15/16 radar system (1-2 km). Pyroclastic cones could also be built from air fall and

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**Table 6b. Calculated Properties of Thin Lava Flow Units on Venus: Flows for Which the Morphology is Controlled by Preexisting Topography**

<table>
<thead>
<tr>
<th>$D_c$, m</th>
<th>$\eta$, 1200 Pa s</th>
<th>$\eta$, 120 Pa s</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5</td>
<td>8</td>
<td>79</td>
</tr>
<tr>
<td>1.0</td>
<td>126</td>
<td>1260</td>
</tr>
<tr>
<td>1.5</td>
<td>638</td>
<td>6380</td>
</tr>
</tbody>
</table>

See text for discussion and definition of parameters.
pyroclastic flow deposits of Pelean activity on Venus (Figures 16 and 17) if more evolved magmas and higher volatile contents are involved. However, in this case we also anticipate that the cone diameter would on the average be less than about 5 km, although larger edifices could be built. Size-frequency data on the features observed on Venus would be helpful in determining whether the majority of the cones fall within, or if the upper end of, the size range of landsforms predicted for Pelean activity (Figures 16 and 17). Mixed effusive and explosive activity to produce mixed volcanoes also cannot be ruled out. An additional possible interpretation for the Venus features is that they represent extrusive domes comparable to domes formed from the extrusion, deformation, and subsequent intrusion of very viscous magmas. Such features are common on Earth within large calderas and in other volcanic provinces [Smith and Bailey, 1968] but are generally restricted in diameter to less than 5 km. Domes morphologically similar to the terrestrial examples are observed on the moon and reach diameters of 20 km [Head and McCord, 1978]. In general, Venus conditions would favor flow formation rather than internal dome growth for such magma types over terrestrial conditions. In addition, the total eruption volume would lead to lower viscosities, producing initially wider, lower flows, less susceptible to subsequent intrusion.

In summary, several types of additional data will be useful in determining the eruption style or styles of the features described by Barsukov et al. [1984a, b, 1986]. The detailed characteristics of size and shape of the edifice also will have extreme importance in determining average diameters and the general diameter range. Data on features of statistical data on landform shape in profile and planform, and information on the nature of the boundary between the base of the structure and surrounding terrain (sharp or subdued and transitional) will allow better assessment of the various eruption conditions. Finally, an assessment of associated features, such as nature, size, and abundance of central craters, evidence for cone breaches, presence of associated flows, cone/dome symmetry and asymmetry, and relation to surrounding structural features will provide data required to assess further the eruption conditions represented by these features.

Possible pyroclastic deposits. Barsukov et al. [1986] have described a section of the flanks of Bell Regio containing a distinctive, continuous, radar bright region which, based on its elongation downwind from the summit area, may be a pyroclastic fall deposit. In total extent the region measures 100 by 300 km, but its outline is irregular. Based on the plan view of an air fall deposit (see Figure 12), we estimate that the width of the deposit lies between 65 and 100 km.

Modeling calculations on the shapes of volcanic eruption clouds [e.g., Wilson, 1976; Sparks and Wilson, 1982] imply that the width of an air fall deposit near the vent is approximately equal to the height of the cloud which deposits it. Figure 10 then shows that eruption cloud heights in the range 65 to 100 km imply mass eruption rates in the range of 3 x 10^{15} to 3 x 10^{11} kg/s (i.e., 30-300 km/h). Though higher than eruption rates associated with basaltic activity on Earth or the moon, these values lie within the range deduced for Plinian and ignimbrite-forming eruptions of silicic magma on Earth. Examination of Figure 13 reveals that a high volatile content is required in a magma on Venus to allow eruptions forming air fall deposits to proceed at these rates: if the transition to pyroclastic flow formation is to be avoided, CO_2 contents in excess of 8 wt % or H,O contents in excess of 6 wt % are implied. Thus, if the air fall nature of the Bell Regio deposit is confirmed, its presence provides strong evidence for the eruption of a very volatile-rich magma on Venus.

6. Summary and Conclusions

6.1. Present Venus Volcanic Environment

Present conditions on the surface of Venus (high surface temperatures and atmospheric pressures) result in a thermal gradient difference such that the temperature is higher at a given depth on Venus than on Earth, and a pressure distribution difference leading to much smaller ratios of subsurface to surface pressure on Venus than on Earth. There are several significant consequences for styles of volcanism on Venus. Although the implied minimum reduction in eruption rates in the same range as terrestrial values, there will be less cooling of magma in the final stages of approach to the surface because the erupting magma will be hotter, and narrower fissure widths are required. For lava flows on Venus the high atmospheric gas density causes convective heat losses to be much more important than on Earth and to exceed radiation losses at temperatures less than about 1000 K. At temperatures above about 900 K the heat loss rate from a lava flow surface on Venus will be about 1.5 times greater than from a comparable terrestrial flow. Relationships of flow surface temperature and thickness of the rigid crust on the flow surface with time show that for half the time after leaving the vent is the crust thickness greater on Venus than on Earth, while for the first hour, surface temperatures will be greater on terrestrial flows. After this time, temperatures will be higher on Venus flow surfaces and the crust thinner. A consequence of this is that flows ceasing after less than about an hour will be longer on Earth by a factor of about 1.3 and those traveling for times greater than an hour will be systematically, but only slightly, longer on Venus than on Earth. These relationships also suggest that behavior in the first hour of flow may enhance the formation of a crust sufficiently to encourage tube-fed rather than open-channel flows, thus possibly enhancing flow lengths on Venus. The same cooling relationships suggest that the transition from pahoehoe to aa textures on flows should occur at times about one-third earlier on Venus (and so closer to the vent) than on Earth. In general, our treatment suggests that there is no reason to expect large systematic differences between lava flow morphology on Venus and Earth. For a given fissure width, we do not anticipate different effusion rates on Venus unless there are systematic differences between magma rheologies. Even if magmas are relatively dry on Venus compared to Earth and have viscosities a factor of 3 greater, yield strengths should be about the same, and levee widths and thicknesses about the same on the two planets. Other
factors, such as flow speed, channel width, and flow duration are all within a factor of 2 for Venus and Earth. Thus variations in lava flow morphology between Venus and Earth are much more likely to be related to variations in local topographic slope or effusion rate than to differences in environmental conditions between the two planets. As on Earth and the moon, sinuous rille channels could be formed by thermal erosion on Venus if mafic lavas are erupted having sufficiently low viscosities and high effusion rates to ensure turbulent flows. However, if volatile (especially CO$_2$) depletion (low magma volatile contents exceeding several weight per cent. The only major exception to this rule is provided by Strombolian activity, in which bubble bursting may not occur at all unless the exsolved magma volatile content exceeds several weight percent. The only major exception to this rule is produced. For example, clasts ejected in transient Strombolian or Pelean explosions in the submarine environment on Earth will also cause major differences in the development of volcanic landforms. Prehn et al. (1985) show that coalescence can concentrate gas sufficiently to cause intermittent explosions in low-viscosity magmas ascending slowly toward the surface. The lower temperatures and pressures characteristic of the Venus highlands are not sufficiently different relative to lowland conditions to dictate a significant change in eruption style; the presence of silicate liquid will tend to favor explosive eruptions. When pyroclastic eruptions do occur, for example, due to abnormally high magma volatile contents or the occurrence of Strombolian activity, pyroclastic fragment velocities will be less by a factor of about 2.5 than those of similar eruptions on Earth, less clast coalescence will take place on Venus than on Earth, and basaltic case pyroclastic eruptions will be more likely to produce lava flows than pyroclastic cones. The high atmospheric pressure and temperature will cause convecting cloud rise heights to be considerably lower on Venus, and pyroclasts will be much less widely dispersed. Eruption cloud heights of 50 km, suggested as a means of raising sulfur dioxide into the upper atmosphere (Esposito, 1984), could only be reached if exsolved magma volatile contents exceeded 4 wt % (regardless of gas species). If pyroclastic eruptions occur, eruption column collapse and pyroclastic flow formation should be much more common than in similar eruptions on Earth, although pyroclasts will be much more mobile on Venus primarily because of the high temperatures of ingested atmospheric gas results in less gas expansion and less consequent mobilization of the flows. Thus the identification of air fall pyroclastic deposits extending for distances in excess of a few tens of kilometers from their vents would directly imply the presence of volatile-rich magmas on Venus. If relatively silicic magmas exist on Venus, they may exhibit lower viscosities on eruption than terrestrial counterparts as a result of the reduced amounts of volatile exsolution that they experience.

3. It has been proposed (Wood, 1979) that conditions on the terrestrial ocean floors at depths between 0.5 and 1 km provide good analogs to conditions on the Venus surface, mainly because of the similarities in ambient pressure. Preliminary studies of the interactions between erupting magmas and seawater as a function of depth (L. Wilson et al., manuscript in preparation, 1985) show that there are both strong similarities to, and major differences from, the Venustian subaerial environment. The effects of water pressure on the suppression of gas exsolution are essentially identical to the effects of the high atmospheric pressure on Venus; similarly, the pattern of convective gas loss that permits Strombolian activity to occur is the same. However, the greater density and lower temperature of the water in terrestrial oceans compared to the Venus atmospheric gases ensure that heat transfer away from a lava flow is always somewhat greater on the seafloor on Earth than on Venus. Thicknesses of cooled crusts are about 5 mm greater on submarine lavas than on Venus lavas, this difference being achieved within 1 min after eruption from the vent. The high density of water as compared with the Venus atmosphere will also cause major differences in the dispersal of pyroclastic fragments, when these are produced. For example, clasts ejected in transient Strombolian or Pelean explosions in the submarine environment on Earth will be much less widely distributed than on Venus. It is less easy to be sure how the conditions differ in steady explosive eruptions (such as those which would produce Hawaiian fire fountains under subaerial conditions), however. The main cause of this uncertainty is the fact that there is currently no detailed model of the conditions of the boiling and subsequent condensation of water on contact with hot clasts in a densely packed jet of volcanic gas and pyroclasts, such as occurs in Surtseyan and phreato-Plinian eruptions on Earth.

4. The above considerations lead to the following conclusions concerning volcanic deposits and landforms. In optically wetter, more extensively non-Earth-like temperature and pressure conditions presently characterizing the surface of Venus, the full range of terrestrial eruption styles may occur. However, Venus surface conditions are such that some styles and resulting volcanic landforms are more likely than others. Several styles are influenced in such a way as to produce landforms with different characteristics than their terrestrial counterparts (Figures 16 and 17). In general, present Venus conditions favor the production of lava flows and discourage the formation of extensive pyroclastic deposits. We would thus predict a relatively higher proportion of areal coverage by lava flows on Venus than on Earth for a given range of conditions. However, since the state of stress in the lithosphere and consequent fissure widths are such important factors in determining effusion rates and vent spacing and density, it is difficult to make further comparisons with the terrestrial environment, in terms of configuration and morphology of regional deposits and provinces, without additional knowledge of Venus tectonic regimes.

Present Venus environmental conditions serve to prevent or strongly inhibit pyroclastic eruptions even at the highest Venus elevations. With the exception of localized secondary volatile enhancement as in the case of Strombolian or Volcanian
activity, pyroclastic activity requires exsolved magma volatile contents in excess of several weight percent. Thus unless these special conditions prevail, pyroclastic deposits and landforms should be very uncommon on Venus. Our treatment leads to a series of predictions concerning the detailed nature of deposits (Figure 16), and resulting landforms (Figures 17-18). Preliminary analyses of Venus radar images show evidence of extensive lava flows (some best explained by very high, lunarlike effusion rates), large low shield volcano structures, locally abundant cone/dome structures, and one possible example of a regional pyroclastic deposit. The treatment outlined here and summarized in Figures 16-18 has allowed possible modes of origin of these features to be specified and provides a framework for the interpretation of subsequent observations of the surface of Venus.

6.2. Past Volcanic Environments

It is widely accepted that the present-day pressure and temperature regime on Venus may not have been present throughout the history of the planet and that both parameters may have been smaller in the past. Most studies [e.g., Donahue et al., 1982] place the evolution of current conditions at an early stage, rather than late, stage of solar system history. Using our general eruption modeling calculations [Wilson and Head, 1983], we can readily explore the consequences of assigning any pressure and temperature values to the Venus environment. However, it is less easy to know what associated changes in the current slow rotation rate of Venus becomes ever more important as regards the disparity between dayside and nightside temperatures, and it becomes progressively more important to know if the planetary rotation rate should be changed to higher values in the past. Fortunately, calculations show that the decrease in pressure will be much more important than the decrease in temperature for most volcanic processes. However, the temperature decrease will have a significant effect on increasing the mobility of pyroclastic flows: if the greenhouse effect were absent from the Venus atmosphere, temperatures at the surface would still be greater than those on Earth because of the greater solar flux. As a result, the run-out distances of pyroclastic flows would then approach (but still not quite equal) those of terrestrial equivalents when the atmospheric pressures were equal.

The main effect of changing conditions on lava flows would be that a reduction of atmospheric density with decreasing pressure would lead to less efficient cooling of the flow surfaces. As we have shown in detail earlier, a pressure reduction to Earth-like conditions would lead to only a 10-30% change in any of the morphological parameters. Even if the pressure were to fall to zero, so that only radiative cooling operated, cooling rates would vary by less than the change from present-day Venus to present-day Earth conditions, leading to a negligible further modification in flow morphologies due to cooling alone. It is possible, however, that large pressure reductions might lead to systematic changes in flow morphology induced by corresponding changes in rheology of the erupted magmas as volatiles were lost by exsolution.

The other major consequence of assuming lower atmospheric pressures on Venus is the increased likelihood of explosive activity in any type of magma (as long as there are at least some volatiles present). Steady explosive activity (Hawaiian or Plinian) will occur at much lower gas contents as the pressure is reduced, will result in more thorough disruption of the magma (hence producing finer-grained pyroclasts), and will lead to increasing eruption velocities, increasing eruption cloud heights, increasing dispersal of air fall and pyroclastic flow deposits, and decreasing values of the size of the largest pyroclast which can be transported to a given range in the deposit.

An impression of the absolute changes in all these parameters can be obtained by examining values for the environments of Venus, Earth, Mars, and the moon by Wilson and Head [1983]. These calculations show that if the dispersal parameter changes are not linear with pressure changes, a two-order-of-magnitude pressure decrease to terrestrial values would lead to a 50% increase in dispersal of most kinds of pyroclastics, whereas a four-order-of-magnitude decrease of pressure to Martian values would yield about a sevenfold increase in dispersal distances.

Since lava flow morphologies are so insensitive to environmental conditions, we cannot use them as useful indicators of the history of the planet's atmosphere by examining lava flow deposits. Of much greater value is likely to be recognition of the presence or absence, at successive levels in the stratigraphic sequence, of numerous, large-scale air fall or pyroclastic flow deposits. These could, of course, be an ambiguity in interpreting the significance of the presence of such deposits, since although many pyroclastic deposits have the volatile contents of the magmas erupted they would have no way of deducing the atmospheric pressure reduction required to allow an eruption of the appropriate kind to take place. It is possible that magma volatile contents could be at least approximately estimated for some kinds of pyroclastic deposits: Wilson [1976] and Wilson et al. [1980] have shown that the sizes of the largest clasts found in a Plinian air fall or pyroclastic flow deposit in the vicinity of the vent can be used to determine the magma volatile content if the ambient pressure is known. In the Venus case a series of atmospheric pressures would have to be assumed, and for each, the clast size information would be used to obtain a volatile content; the combination of volatile content and pressure would then allow an assessment to be made of the likelihood of formation of the deposit observed. Unfortunately, this may not be a practicable way of interpreting pyroclastic deposits on Venus since pyroclast sizes in the meter size range cannot easily be deduced using radar systems: clast size is likely to be confused with general meter-scale roughness.

We can, however, make some deductions about past atmospheric conditions on Venus. The survey of about 30% of the Venus surface carried out by
the Venera 15/16 probes [Barsukov et al., 1984a, b, 1986] has revealed only one candidate for a large pyroclastic deposit. In view of the evidence for slow erosion rates on Venus [Head et al., 1985b], this implies that large explosive eruptions are not common at the present time and hence that magmatic volatile contents do not commonly exceed about 4 wt %. Impact crater counts [Barsukov et al., 1984a, b, 1986] show that the average age of the surface mapped lies between 500 and 1000 Ma. Since the likely trend of evolution of availability of magmatic volatiles on Venus is toward smaller amounts of volatiles being present in magmas at later times as the planet degasses, we take this to imply that large explosive eruptions have not been common for at least the last 500 m.y., even though magmatic volatiles may have been more plentiful in the past. This, in turn, implies that the atmospheric pressure has been at least as large as the present-day value for a time of at least this order.

6.3. Relation of Volcanism to Tectonism and Lithospheric Structure

On Earth, volcanism is closely linked to tectonism in terms of areal distribution, style, and tectonic characteristics of the crust and lithosphere. Furthermore, volcanism can be interpreted as a fundamental aspect of local and regional tectonic activity, and it is a major factor in determining effusion rates on the planets [Wilson and Head, 1983]. On the basis of theory and observations discussed in this paper, we can offer several tentative conclusions about the relation of volcanism and tectonism on Venus.

6.3.1. Volcanism and tectonic environment.

Although the nature of the surface of Venus is only partly known, preliminary observations show that many of the volcanic features are related to regional tectonic characteristics. Venus, the prominent shield volcano in Beta Regio, is situated on the western bounding fault of Devasa Chasma, a major extensional rift structure [Campbell et al., 1984]. Several other large volcanic edifices [Stofan, 1985] are located along regional structural trends, although it is unknown whether they are in extensional or compressional environments. The cone/dome structures reported by Barsukov et al. [1984a, b, 1986] occur in extensive fields in the plains and may reflect the structural fabric of the upper crust and lithosphere in these regions [e.g., Settle, 1979b]. Larger shield structures southeast of Ishtar Terra are related to preexisting linear trends, and they may also contain information about lithospheric structure [Head et al., 1985a].

When magma is able to pass completely through a planetary lithosphere to the surface, the heat transfer can be regarded as advective (though volcanic deposits in the form of lava flows or ash layers carry out the final stages of their cooling by conduction). When magma only partially penetrates the lithosphere and cools in intrusive dikes or sills, the activity represents a hybrid of advection and conduction and is effectively a way of increasing the average value of the temperature gradient in the outer layers of the planet. Intrusion of low-viscosity basaltic material in narrow, linear dikes which can cool quickly is a more effective means of heat loss than the formation of more equant high-level magma bodies. It is tempting, therefore, to assume that regions of Venus having high heat flow rates will be preferentially associated with linear surface patterns of volcanism and tectonism. Linear tectonic patterns are linked on Earth with other processes besides rifting and intrusion, and so the best chance of understanding these relationships on Venus will probably come from identifying and mapping the global distribution of volcanic centers.

6.3.2. Volcano heights and lithospheric properties.

The height to which volcanoes can grow depends in part on the depth of origin of the magma and the density contrast between the lava and the rocks between the source and the surface. Eaton and Murata [1960] and Carr [1973, 1981] have applied these concepts to Earth and Mars and have suggested that the source depths for the approximately 9-km-high Hawaiian shields are about 60 km, while those for the Martian shields in excess of 20 km altitude are at least 160 km. Even the largest and relatively youngest of the identifiable shield volcanoes on Venus (e.g., Theia Mons) do not appear to attain as great topographic heights as analogous structures on Earth and Mars, yet their diameters are similar to those of the largest shields on Mars. Even though viscous relaxation is an important process in the present environment [Weertman, 1979; Solomon et al., 1982], these youngest volcanoes do not appear to have structural or topographic features associated with these processes. This observation of volcano height may be related to the structure of the Venus lithosphere, and hence that magma volatile contents do not commonly imply that large explosive eruptions are not common for at least the last 500 m.y., even though magmatic volatiles may have been more plentiful in the past. This, in turn, implies that the atmospheric pressure has been at least as large as the present-day value for a time of at least this order.

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that the maximum height above the mean surface to
which magma can be raised to produce a volcanic edifice is an increasing function of the mean den-
sity difference. Lower density differences on
Venus would therefore be expected to lead to
smaller typical heights of volcanoes built up
largely from deep-seated eruptions. For example,
if density differences on Venus were limited to
the approximately 100 kg/m\(^3\) difference between
basaltic solid and liquid densities (implying that the
magma sources were located only within the crust),
then volcanic edifice heights would be
systematically about a factor of 4 less than on
Earth, where effective density differences for
magma originating in the mantle range up to about
400 kg/m\(^3\).

Analyzes of crack geometries given by Weertman
[1971] and Secor and Pollard [1975] show that the con-
sequence of reducing the driving density con-
trast acting on a migrating magma-filled crack is to
make the crack narrower for a given vertical crack
height or magma volume. This will increase the
amount of cooling experienced by a given batch of
magma and decrease its ability to reach the
surface. This implies that larger minimum magma
volumes may be required to ensure surface erup-
tions (as distinct from crustal intrusions) on
Venus than Earth, thus making intrusive rather than
eruptive events statistically more frequent
and adding to the problem of constructing high
volcanic edifices. Also, since the minimum fissure
width and length needed to permit an eruption
to take place are larger than on Earth and since the
mass flux per unit horizontal length of a fissure is an increasing function of fissure width
[Wilson and Head, 1981], it is likely that when an
eruption does take place, it will do so at a
higher mass eruption rate than on Earth, thus
helping to explain the apparently common occur-
rence of long lava flows on Venus.

The possibility that magma sources are located
at systematically shallower depths on Venus than
on Earth also has a bearing on the observation
that large-scale, steady explosive eruptions appear to be rare. Table 3 shows that carbon
dioxide has a role as a potential source of magmatic
activity unless sources have depths between 35 and
80 km. However, water (or any volatile having a
similar pressure dependence of solubility) could
produce explosive activity in magmas ascending
from depths in excess of 1-3 km. It may be,
therefore, that the paucity of extensive explosion
products, if confirmed by future investigations to
be planet-wide, can be taken as direct evidence of
the current depletion of such volatiles in the
Venus interior.

6.3.3. Volcanic calderas and lithospheric
structure. Depressions identified as volcanic
calderas on Venus (e.g., Colette, Sacajawea, and
the depression of Theia Mons) range in size from
90 to about 200 km, a factor of 10 larger than the
sizes of most terrestrial examples [Wood,
1984]. On Earth there is evidence that the diame-
ters of calderas are approximately equal to the
diameters of high-level magma reservoirs [Ryan et
al., 1981, 1983], the draining of which during one
or more eruptions is the main reason for caldera
formation. Additionally, the shapes of the stress
fields generated in the crust by the inflation of
magma chambers during preeruption periods [Mogi,
1958] are such that the diameters of the resulting
calderas are comparable to the depths of the magma
bodies. This relationship breaks down, on Earth,
for relatively deep magma reservoirs, however.
Evacuation of such bodies is more likely to lead
to formation of a broad, regional downwarp
[Walker, 1980] than a caldera. The reason is pre-
sumably related to the fact that the amplitudes of
stresses induced at the surface by deep, inflating
magma chambers are not great enough to cause the
fracture patterns required for caldera collapse.

If we attempt to take the sizes of the larger
calderas on Venus as representing the depths to
the corresponding magma bodies, we obtain depths
which are of the order of 5 times as great as current
estimates of the thickness of the thermal
lithosphere [Solomon and Head, 1982, 1984]. A
simple way of resolving this paradox is again to
assume that the Venus lithosphere is unusually
thin, at least in the immediate vicinity of the
calderas, and that caldera formation is controlled
directly by the stress field induced by magma
accumulation in a large region which amounts to an
upwarp of the asthenosphere. This assumption has
the added attraction that it provides a reason for
the presence of the large magma volumes (of the
order of 100 km\(^3\)) which must be removed (either by
eruption to the surface nearby or by injection
into the crust as intrusions) to justify the calder-
a volumes.

6.4. Observations From Future Venus Missions

There are a number of critical observations
relating to volcanic eruption styles and the rela-
tionships between volcanic and tectonic processes
on Venus that could be made from future missions.
1. It is important to establish whether the
apparent near-absence of large-scale explosive
activity seen in those parts of Venus examined so
far is true of the whole surface of the planet.
If it is, this will imply a severe depletion of
all high-solubility volatiles in the Venus
interior for the last several hundred million
years and will reinforce the arguments that we
have given about the length of time that the present
atmosphere has existed.

2. Information is required about the small
cone and dome structures at higher resolution than
that currently available. It is important to
establish the frequency of occurrence and sizes of
summit pits and craters on these features and the
presence or absence of related lava flows or
pyroclastic deposits, since this will help to
reduce the ambiguities in assigning their forma-
tion mechanism.

3. More detailed morphological information is
needed for the features identified as lava flow
deposits. Regrettably, it is very hard to see how
current radar techniques will be able to yield
information on the thicknesses of lobes. However,
widths can be established for levees and central
channels and if the lengths of individual flow
lobes can be established unambiguously, more
information can be deduced about eruption rates
and lava rheological properties and hence likely
caldera compositions.

4. Higher-resolution images of shield volca-
noes are needed to establish vent locations for
summit and flank eruptions, the detailed struc-
tures of the summit calderas, and the time order
in which these features formed. This information
would shed light on the evolution of the stress fields within the volcanoes, which are in turn related to the state of the lithosphere via magma supply rates and rates of deformation of the volcanic edifices.  

5. Systematic measurements of the variation of radar roughness characteristics between different parts of volcanic edifices and individual flow units would aid in the definition of morphologically distinct features; though it will not be possible to obtain unique interpretations of radar roughness signatures without more experimental radar studies of terrain types on Earth.  

6. Additional information (from automated landers, sample return missions, or other techniques) on the composition of the atmosphere and the nature of short-term changes there, as well as the bulk composition and volatile content of surface rocks, is of fundamental importance to further understanding of Venus volcanic processes.

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