LAVA FLOWS ON VENUS: ANALYSIS OF MOTION AND COOLING. Bruce A. Campbell and James R. Zimbelman, *Center for Earth & Planetary Studies, National Air & Space Museum, Washington, DC, 20560.*

Introduction. Effusive volcanism is a major mechanism for resurfacing on Venus. The large edifice complexes are marked by lava flow fields which extend hundreds of kilometers from their apparent source vents. Earlier work on the emplacement of lava predicted few major differences in flow length between Venus and the Earth (for similar magmas) based primarily on assessment of the relative rates of heat loss to their respective atmospheres [1]. This analysis, combined with the obvious presence of very long lava flows on Venus, has led to suggestions that basalt magmas on Venus may be of generally lower viscosity or that magma effusion rates exceed terrestrial norms. If correct, these hypotheses might imply systematically different chemical and dynamic characteristics for volcanic processes on Venus, and a somewhat catastrophic history for many of the large edifices and flow fields. In this abstract, we re-examine the mechanics of magma motion and cooling, and discuss some of the issues involved in analyzing flows on Venus.

Flow Motion. Magma motion can be described by two expressions for the flow of a Bingham plastic material. The first could be termed "gravity-driven flow", and the flow per unit area through an interface is [2]:

$$q = (1/\nu)[h^3\rho g\sin\theta/3 - yh^2/2 + y^3/(6\rho^2 g^2 \sin^2\theta)]$$

where ν is viscosity, ρ is density, g is acceleration due to gravity, y is yield strength, h is the height of the flow, and θ is the slope of the underlying ground. The second mechanism can be called "pressure-driven flow", where the $\sin\theta$ term in the above equation is replaced by $\tan\phi$, ϕ being the slope of the upper magma surface. This latter case becomes most important when the topographic gradient is very small, since it expresses the tendency of a flow to spread out over a flat surface due to its internal pressure gradient. Flow will not commence until the depth reaches a critical height, which is different for the two cases:

$$\text{Gravity: } h_c = y/\rho g\sin\theta \qquad \text{Pressure: } h_c = y/\rho g\tan\phi$$

If the yield strength is zero, which is expected for melts very close to the liquidus, then the above expressions reduce to those for Newtonian flow, and the critical height disappears. It is evident that the gravity-driven flow mechanism will act slowly on shallow slopes unless the yield strength is relatively low. The pressure-driven case may flow much more readily over flat terrain by exploiting the slope along the front of the liquid body (which may reach 30 degrees or more).

Flow Cooling. The loss of heat from a lava flow predominantly takes place through the upper surface, where energy is lost by radiation and convection to the atmosphere. Soon after eruption, basalts form a cooler crust at this top margin. Heat is then lost by conduction through the crust, whose lower surface temperature slows heat transfer to the atmosphere, and by more efficient energy transfer from hot core material exposed along cracks [3]. Terrestrial experience suggests that heat lost by a lava flow is typically accomodated by increases in the thickness of the crust rather than by major changes in core temperature.

The growth of a crust on a lava flow has been modeled previously with the assumption of a fixed upper surface temperature [4]. We carried out a similar analysis but permitted the crust surface temperature to vary with time so as to remain in thermal equilibrium with the heat flux from below and the radiative/convective losses to the atmosphere. By balancing the heat flux at top and bottom, we can at any moment calculate the temperature of the crust surface, and thus iteratively model the growth of the crust as a function of time. Our results support those of [1] in that flow crusts form much more rapidly on Venus, but after about 1 day crustal thickness will be greater on terrestrial flows (the crossover thickness is ~50 cm).

The above discussion of flow cooling is appropriate only for flows which move over the ground surface transporting an integral crust. If the lava becomes sufficiently crusted over

to form tubes, then heat loss from the magma may become negligible over distances of 10's to 100's of km. Only at the flow front or in breakouts along the tube will magma rapidly lose heat to the atmosphere. Tube formation is generally limited to low to moderate effusion rate flows, since at higher discharge rates tubes will either overflow or never form due to rapid recycling of the surface crust.

Practical Applications. Using the models for flow motion and cooling discussed above to analyze real flows is difficult. Laboratory data for basalt viscosity and yield strength are very limited, and in all cases use remelted material. Field measurements have been used to estimate magma rheology, but these approaches must assume a particular flow law in order to extract the parameters. Lab analysis shows that melts close to the liquidus can have nearly zero yield strengths. Even when held at a fixed temperature, a basalt melt will increase in viscosity and yield strength until equilibrium is reached [5,6].

Why a lava flow stops moving is also very much in debate: it can occur when the crust thickens to a point at which the central hot core no longer exceeds the critical depth for flow, when the flow core cools sufficiently to raise the yield strength and viscosity, or when the magma supply ceases and the flow halts at its critical depth.

Lava Flows on Venus. The basic issue we address is the presence of long (100's of km) lava flows, many of which are characterized by surface roughness comparable to that of terrestrial pahoehoe surfaces [7], emplaced on slopes of less than 1 degree. The great length of these flows is often cited as evidence for rather high effusion rates or low magma viscosity, but the lack of significant surface roughness argues for low magma effusion rates based on terrestrial experience [8].

Several general statements can be made for lavas on Venus: 1) magma viscosity will be increased over equivalent terrestrial basalts due to the lack of water, 2) eruption temperatures will be slightly higher, causing reduced viscosity, 3) volatile exsolution is suppressed by the atmosphere, delaying the onset of crystallization and lowering the viscosity further, and 4) the more rapid formation of a crust on Venus shields the core from the atmosphere, and is likely to lessen the total heat loss by a significant amount [1,5,6]. The degree to which these mechanisms operate is uncertain, but they may permit a lower average viscosity and a more favorable cooling regime for basaltic magmas on Venus.

We propose that most lava flows on the major Venus highland edifices are similar to terrestrial basalts, with common effusion rates of 1-10 m^3/s, and that such low-volume eruptions form the vast aprons and mottled fields characteristic of these constructs. The radar-bright flows are most likely isolated cases of higher-volume eruptions (on the order of 100-1000 m^3/s) which produce a'a surface textures. The great lengths of the smooth flows are not necessarily indicative of larger eruption rates or lower viscosity; tube-fed flows on Earth might travel equal distances were they not stopped by the ocean or regional topography. Eruptions in the lowlands of Venus may be of much higher volume flow rates since magma is not expected to stall within the crust prior to eruption [9].

Numerical Models. We are investigating the differences in lava flow emplacement on Venus and the Earth using numerical simulations based on the magma motion and cooling models discussed above. This work treats each small region of the flow as a discrete element, and solves for local flow rates and heat loss to the atmosphere or ground [10]. Early results are encouraging [11], and we are expanding the range of behaviors to better simulate the emplacement process.

References. [1] Head, J., and L. Wilson, (1986), *JGR*, 91, 9407-9446; [2] Dragoni, M., et al., (1986), *J. Volc. & Geotherm. Res.*, 30, 305-325; [3] Crisp, J., and S. Baloga, (1990), *JGR*, 95, 1255-1270; [4] Turcotte, D., and G. Schubert, (1982), Geodynamics; [5] McBirney, A., and T. Murase, (1984), *Ann. Rev. Earth Planet Sci.*, 12, 337-357; [6] Asimow, D., and J. Wood., (1992), *JGR*, 97, 13643-13666; [7] Campbell, B. A. and D. B. Campbell (1992) *JGR*, 97, 16293-16314; [8] Rowland, S., and G. Walker, (1990), *Bull. Volc.*, 52, 615-628; [9] Head, J. et al., (1992), *JGR*, 97, 13153-13198; [10] Ishihara, K. et al., (1990), *IAVCEI Proc.*, 2, 174-207; [11] Zimbelman, J., et al., (1993), this volume.