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CANALI-TYPE CHANNELS ON VENUS: SOME GENETIC CONSTRAINTS

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Abstract. Canali-type channels on Venus are unique because of their great lengths (up to 6800 km) and nearly constant channel cross sectional shapes along their paths. A simple model incorporating channel flow and radiative cooling suggests that common terrestrial-type tholeiite lava cannot sustain a superheated and turbulent state for the long distances required for thermal erosion of canali within allowable discharge rates. If canali formed mainly by constructional processes, laminar tholeiitic flows of relatively high, sustained discharge rates might travel the observed distances, but the absence of levees would need to be explained. An exotic low temperature, low viscosity lava like carbonatite or sulfur seems to be required for the erosional genesis of canali.

Introduction

The canali-type channels discovered on Venus are unique in their morphology [Baker et al., 1992]. They are typically longer than 500 km, and the longest ones extend for distances from a few thousand km up to 6800 km. In addition to their great length, these channels are unique for their nearly constant width throughout their entire length. The channels lack obvious levees, with the exceptions of a few localized, narrow ones observed on Magellan SAR images.

Because canali occur in plains regions which probably consist of mostly tholeiitic rocks similar to MORB [Surkov et al., 1983], we initially hypothesize the formation of canali by open-channeled flow and erosion by a tholeiitic lava. Our approach is to assume that turbulence is required for the lava to retain its capacity to thermally and mechanically erode a channel.

Dimensions of three major canali are shown in Table 1. A depth was measured for channel #1 using radar foreshortening measurements at the deepest appearing position where the measurement method can be applied (Table 1). The result indicates that the depth is less than 50 m, most likely a few tens of meters. We note that the accuracy of depth measurement for such shallow channels like canali (a few tens of meters) is questionable. Because the canali shown in Table 1 do not have distinct sources or termini, their total lengths may be longer than indicated.

Travel Distance

Turbulent flow above eutectic temperature. Turbulence is a key factor in lava-channel erosion, and it will significantly enhance thermal erosion if the lava temperature is near or above the liquidus melting point of the substrate [Hulme, 1982]. Most terrestrial lavas are erupted near or slightly below their liquidus, as indicated by the usual occurrence of phenocrysts. Such lavas commonly do not have significant thermal erosional power. In rare cases, lavas are superheated above the liquidus, and, in other cases, near-liquidus mafic or ultramafic lavas have temperatures well above the liquidus of a more silicic substrate. In either case, significant thermal erosion may occur as long as the flow remains turbulent. However, because of rapid mixing, turbulent flows tend to cool more rapidly than laminar flows. The Reynolds Number indicates the degree of turbulence and can be expressed in terms of the flow thickness (h), velocity (v), density (ρ) and dynamic viscosity (η), or in terms of a 2-dimensional discharge rate (Q_{2D}) and kinematic viscosity ($\nu = \eta/\rho$), as follows:

$$Re = \rho hv/\eta = Q_{2D}/\nu. \tag{1}$$

In the turbulent flow regime, mixing of lava exposes the hot interior core to the atmosphere to enhance radiative cooling. In an ideal fully mixed tholeiitic lava, the heat flux from the lava by radiation (F_R) is determined from the lava temperature T [Hulme, 1973] by

$$F_R = \varepsilon\sigma(T^4 - T_a^4), \tag{2}$$

where emissivity (ε) is 0.9, Stefan's constant (σ) is 5.67×10^{-8} $Wm^{-2}K^{-4}$, and the venusian ambient temperature (T_a) is 750 K. The cooling rate can be estimated by the equation

$$(-\rho C_p h)(dT/dt) = F_R, \tag{3}$$

where a typical tholeiite density (ρ) is 2700 kg/m^3 and specific heat (C_p) is 1217 J kg^{-1} K^{-1} [Touloukian and Buyco, 1970]. Equations (2) and (3) were solved semi-numerically. The rate of temperature change (dT/dt) was calculated at each T and derived time intervals were used to compute elapsed time. The eruption temperature was assumed to be 1500°C. In terrestrial lava eruptions, the observed lava temperature tends to be near the liquidus which is about 1200-1300°C for tholeiitic lava (1200°C for Hawaiian tholeiite [Shaw, 1969], 1274°C for diopside-anorthite eutectic [Cox et al., 1979]). 1500°C is much hotter than we would expect for tholeiite eruption on either Venus or Earth, but is assumed to allow for the extreme case of unusually hot eruptions. Such high temperatures could result from exceptionally rapid, adiabatic ascent from very deep levels in Venus leading to massive decompressive melting, or, alternatively, from heating of basaltic crust by a large volume of intruded komatiite magma. This temperature was used to optimize the eruption condition.

The temperature-dependent viscosity (in poise) of an eutectic tholeiite-like mixture, approximated as 58% diopside + 42% anorthite (eutectic temperature = 1274°C [Cox et al., 1979]), is given by

$$\mathrm{Log}\,\eta = (A+B/(T - T_o)), \tag{4}$$

where A = -1.587, B = 2304 and T_o = 656.5°C [Scarfe et al., 1983]. To convert the time to travel distance, the velocity of the tholeiite flow is determined, for the highly turbulent regime, by the Chezy equation,

$$v = (2gh\alpha/C_f)^{1/2}, \tag{5}$$

where α is a regional slope and C_f is a friction coefficient.

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TABLE 1. Size Dimensions of Three Major Canali-Type Channels
(N/A: Not Available)

	1	2	3
Location	Lat. 53°N-11.5°N, Long. 159.5°E-185°E	Lat. 59.5°S-50.5°S, Long. 168°E-205°E	Lat. 18.5°N- 8°N, Long. 145.5°E-151.5°E
Length	6800 km	3000 km	1800 km
Width	1-3 km	1-2 km	1-2 km
Depth	< 50 m	N/A	N/A

Morphological similarities between venusian channels and terrestrial fluvial channels [Baker et al., 1992] imply a similarity of flow-resistance properties for canali lavas and channelized water. This similarity allows us to assume $C_f = 0.001$, which is a typical friction coefficient for a natural river bed. The Venus gravity (g) is 8.87 m/s^2. The regional slope (α) is assumed to be 0.001, since most canali occur in lowland plains. For a fixed 2-dimensional discharge rate (Q_{2D}), the velocity and the depth of the flow can be calculated simultaneously because $Q_{2D} = vh$ (Table 2). We calculated the temperatures of the flow at the given travel distance for Q_{2D} = 1, 10, 100, and 1000 m^2/s (Figure 1). The calculations (Table 2, Figure 1) show that tholeiitic lava cannot travel the maximum observed distance (6800 km) as a turbulent flow before reaching eutectic temperature, regardless of the following optimizations: 1) No heat loss by either atmospheric convection or by conduction to the ground. 2) Ignoring melting of ground material. 3) Use of a high eruption temperature (1500°C) relative to the liquidus temperature (1274°C). 4) Assuming the largest Q_{2D} was 1000 m^2/s, corresponding to the depth of about 38.34 m at a velocity of 26.08 m/s. This is about the deepest limit compared with the nominal depth of channel #1 (Table 1). The depth of the channel is an upper limit for the lava flow, unless lava solidified and deposited on the channel floor. Since channel #1 is the longest, widest, and presumably the deepest of all canali, $Q_{2D} = 1000$ m^2/s is probably the largest reasonable discharge. 5) C_f may actually be as high as 0.03, which Huppert and Sparks [1985] used to compute velocities of komatiite ascending through fractures. This high C_f would not change the travel distance for each discharge rate because the increased velocity is cancelled by the lowered cooling rate in a thicker flow. However, the increased thickness of a high discharge flow is less realistic because of the relatively shallow extant channel depths. 6) By using an eutectic composition, lava takes a longer time to initiate crystallization and transition to laminar flow. 7) Use of a low critical Reynolds Number (Re = 500) for the turbulent-laminar transition. The Reynolds Number of the fully turbulent flow would be much higher.

Note that the assumed discharge rate makes a large difference in the distance which the flow can reach (Table 2). For the low discharge rate $Q_{2D} = 1$ m^2/s, the flow changes to laminar even before it cools to the eutectic temperature.

TABLE 2. Distance Turbulent Flows of Various Discharge Rates Can Travel Before Reaching Eutectic Temperature

Discharge Q_{2D} (m^2/s)	1	10	100	1000
Velocity (m/s)	2.61	5.62	12.11	26.08
Flow thickness (m)	0.38	1.78	8.26	38.34
Distance (km)	1.20	20.33	203.3	2032

Period of crystallization. A crystallizing flow can travel a considerable distance before completely solidifying. As noted above, most terrestrial lava flows have already started crystallizing even before they erupt. A fluid with suspended solidified particles (a slurry) will exhibit a rapid increase of viscosity as the fraction of crystals increases. We calculated the distance this slurry would travel before its viscosity increased to cause a transition to laminar flow, thereby ceasing effective channel erosion. Our best-fit equation to viscosity data for Hawaiian basalt [McBirney and Murase, 1984] is

$$\mu_r = \eta_{slurry} / \eta_{liquid} = \exp\,[13\phi/(1-1.35\phi)], \qquad (6)$$

where μ_r is a dimensionless relative viscosity, η_{slurry} is the viscosity of the slurry, η_{liquid} is the viscosity of the pure intergranular liquid, and ϕ is the crystal fraction. This equation, valid for ϕ up to 0.3, and strictly valid only for Newtonian behavior, describes the sharp dependence of viscosity on crystal fraction as observed in real lavas. Similar relationships hold for a wide chemical variety of other silicate lavas, including dacite and for ammonia-water slurries [Kargel et al., 1991]. Equation (6), in particular the coefficient (13), implies considerable physical interaction among distinctly nonspherical grains.

For our simplified test case of a eutectic tholeiite-like composition, the temperature, liquid composition, and η_{liquid} do not change as crystallization ensues. Thus, from equation (4) $\eta_{liquid} = 13.9$ Pa-s. For the flow to be turbulent, the Reynolds Number has to be > 500. The viscosity of the slurry at Re = 500 and η_{liquid} is used with equation (6) to estimate ϕ at the critical point when the flow becomes laminar. The crystal fraction, ϕ, increases with time as the flow cools, and has the following relationship:

$$d\phi/dt = F_R/L\rho h, \qquad (7)$$

where L is the latent heat of fusion of the eutectic = 4.36×10^5 J kg^{-1}. Since the eutectic temperature is constant, the emitted flux F_R also is constant. Hence, $d\phi/dt$ is constant until a solid crust develops. It follows that the time (t) during which the flow remains turbulent, after reaching the eutectic temperature, is given by

$$t = \phi_c L\rho h/F_R, \qquad (8)$$

where ϕ_c is the crystal fraction for onset of laminar flow. The distance traveled by the lava during this period is

$$X = vt, \qquad (9)$$

and results are shown in Table 3. The results suggest that flows at a very high discharge rate may travel over 1000 km after reaching eutectic temperature and before becoming laminar. Such flows may mechanically erode a channel, but significant thermal erosion is not possible if the substrate has a composition similar to the lava.

Laminar flow. In a laminar flow, a solid crust may develop which causes a lower rate of heat loss since the surface radiates at a lower temperature than the pure liquid. The velocity of a laminar flow at low slopes (α) is calculated by the equation [Murase and McBirney, 1970]:

$$v = \rho g \alpha h^2/3\eta. \qquad (10)$$

At the laminar-turbulent flow boundary (Re = 500), viscosity

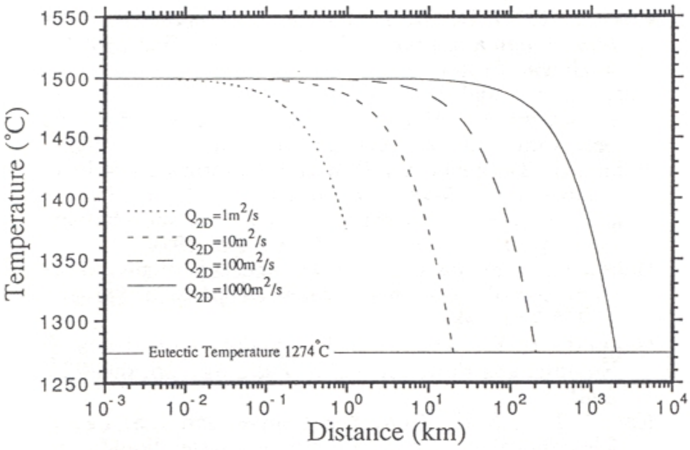

Fig. 1. Traveling distance of turbulent tholeiitic lava for four 2-dimensional discharge rates before reaching eutectic temperature. The flow of low discharge rate (1 m^2/s) transits to laminar flow even before reaching eutectic temperature. Regardless of optimizations, tholeiitic lava does not travel the observed distance before reaching its eutectic temperature.

(η) is determined for each 2-dimensional discharge rate (Q_{2D}) = vh, allowing calculation of v and h for tholeiite. Assuming velocity is constant, the distance this flow can travel before a solid crust forms and attains half the flow thickness (assumed to be sufficient to stop the flow advance) is estimated by the following equation [Head and Wilson, 1986]:

$$X = h^2 v / \Lambda \kappa, \tag{11}$$

where Λ is a non-dimensional Gratz number determined to be 270 for venusian lava, and κ is the thermal diffusivity of the solid basalt crust at a mean crustal temperature near 900°C (0.6 mm^2/s). Results are shown in Table 4. Laminar flows may travel an extremely long distance, if the flow is heavily crusted and the discharge rate is high. However, channel erosion generally cannot occur, and the flow will tend to fill in any channel through which it passes.

Discussion

Erosional origin by tholeiite lava. For a lava to thermally erode, its temperature has to be higher than the eutectic temperature of the substrate, which is assumed to have a similar eutectic to that of the lava. We found that for the lava erupted at 1500°C, a discharge rate of over 3000 m^2/s, corresponding to a flow thickness of ~ 80 m, is required to travel 6800 km before reaching its eutectic temperature under the optimized conditions. This would only be possible if the flow was much deeper than the observed present depth of the channel. If the channel was once deeper and continued eruption provided lava to be solidified on the channel floor, or degradational processes have modified the channel structure, the measured depth may not reflect the real depth of formation. However, there are morphological arguments against such a high discharge flow. Assuming a channel width of 1 km, 3000 m^2/s corresponds to a 3-dimensional discharge rate of about 3×10^6 m^3/s. This far exceeds eruption rates of a terrestrial flood basalt event, which is about 1 km^3/day/km (equivalent to 11.6 m^2/s [Swanson et al., 1975]). The rate of 3×10^6 m^3/s is equivalent to the water discharge rates responsible for the origin of terrestrial and Martian outflow channels [Baker, 1982]. Such a high discharge rate is expected to cause a braided pattern as we observe in outflow channels. The more obvious results of lowering temperature and the Reynolds Number as the flow travels are shallowing and narrowing of the channel in the downstream direction. These are primary characteristics of erosional lava channels on Earth, Mars, and the Moon, and certain types of channels on Venus, but have not been observed for venusian canali [Baker et al., 1992].

TABLE 3. Distance Turbulent Flows of Various Discharge Rates Can Travel After Reaching Eutectic Temperature

Discharge Q_{2D} (m^2/s)	10	100	1000
η_{slurry} (Re = 500)	54	540	5400
$L\rho h/F_R$	7589	35215	163454
ϕ_c	0.0915	0.2040	0.2833
t(s)	694	7184	46307
Distance (km)	3.90	87.0	1208

Large effusions of lava may travel some distance as turbulent flows after reaching their eutectic temperature (Table 3) and additional great distances during the laminar stage of flow (Table 4). However, once reaching the eutectic temperature, thermal erosion is ineffectual, and, once the lava enters the laminar regime, it becomes ineffective both as a thermal and as a mechanical erosional agent. Moreover, the expected longitudinal change of flow rheology as crystal content increases and as a solid crust thickens should be reflected in channel morphology. This transition is not observed in canali.

Constructional origin. Turbulence is not required if the canali originated mainly as constructional channels. As suggested in Table 4, the discharge rate of tholeiitic lava does not necessarily have to be over 100 m^2/s nor must eruption occur at very high temperature for travel over 6800 km. Nevertheless, lava generally ceases to flow when the supply of lava is cut off, which usually happens long before the lava solidifies [Hulme, 1982]. The total canali lengths of thousands of kilometers indicate that the flow durations must have been abnormally long. Although the shallow depths of canali may not require incision into the substrate, the lack of obvious canali levees is a clear contrast with many well-defined lava levee/channel morphologies that are widespread on the venusian surface [Baker et al., 1992]. According to Hulme [1974], the levee width (w_b) is expressed by

$$w_b = Y / 2 g \rho \alpha^2 \tag{12}$$

where Y is a yield strength. For a range of Y for the terrestrial basalts 10^2-10^4 N/m^2 [Moore et al., 1978], the width is about 2-200 km. These are observable scales on Magellan images. It is possible the levees are not observed because the channel-forming lava had a very low yield strength resulting in a very narrow levee. Alternatively, a very low-viscosity of lava might not produce a rough surface. A degradational and/or burial processes may have obliterated levees.

Roofing. A thickened crust of laminar flow could eventually lead to the formation of a solid, and fixed roof. If canali originally formed as lava tubes, roofing would insulate lava from the atmosphere and reduce the cooling rate. According to Oberbeck et al. [1969], the maximum width of uncollapsed roofs of terrestrial tubes is about 30 m. For similar gravity and higher surface temperature on Venus, the

TABLE 4. Distance Laminar Flows of Various Discharge Rates Can Travel

Discharge Q_{2D} (m^2/s)	1	10	100	1000
Velocity (m/s)	1.14	2.45	5.29	11.39
Flow thickness (m)	0.88	4.08	18.90	87.80
Distance (km)	5.45	252	11664	5×10^5

maximum roof width would not be very different from that on Earth. For the wide channels, as soon as the support by flow is removed, the roof has to collapse. The canali-type channels' floors appear to be generally flat or ridged, and do not show clear evidence of collapsed roofs. So it is difficult to prove this origin morphologically.

Other types of lava. Baker et al. [1992] proposed alkaline or ultramafic silicate lavas, carbonatite and sulfur as candidate channel-forming materials. These are known to have low viscosities at their melting temperatures. Alkaline and ultramafic silicate lavas have higher eruption temperatures than that of tholeiite, which may lead to marginally longer cooling times, but higher solidification temperatures tend to counterbalance this advantage.

Carbonatite lavas have melting points lower than those of silicate lavas, and viscosities orders of magnitude lower. Alkali-rich carbonatites have melting points comparable to or slightly greater than Venus ambient temperatures, potentially allowing considerable travel distance and mechanical erosional capacity. Sulfur's melting temperature is lower than the ambient temperature, so it remains molten under current venusian conditions. The viscosity of sulfur under Venus surface conditions is comparable to that of water on Earth. Carbonatite and sulfur may be satisfactory candidates in terms of the travel distance. Although these lavas cannot thermally erode ground materials, their advantage is that they can travel long distances even at very low discharge rates. Also, because the melting temperatures of these lava candidates are close to or lower than Venus ambient conditions, temperature-dependent rheologies remain nearly constant all the way. These advantages make many morphological characteristics easier to understand, and reinforce morphologic and physical analogies to the case of terrestrial channels mechanically eroded by water [Baker et al., 1992]. Both erosional and constructional origins by carbonatite and sulfur are plausible. Sulfur has one limitation not suffered by carbonatite, namely that the low humidity of sulfur in the Venus atmosphere causes it to evaporate rapidly, whereas the CO_2 pressure of the Venus atmosphere is near carbonate saturation.

Although carbonatite and sulfur are rheologically satisfactory, production of these lavas must be explained. The total amount of lava required to scour channel #1 in Table 1 is estimated to be about 10^{12} m^3, assuming width is 1 km, depth is 20 m and ratio of lava to eroded material is 10. No terrestrial examples of carbonatite and sulfur volcanism are known to match this scale. If venusian canali were eroded by such exotic lavas the scale would be unique among the terrestrial planets.

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References

Baker, V. R., *The Channels of Mars*, 198 pp., University of Texas Press, Austin, Texas, 1982.

Baker, V. R., G. Komatsu, T. J. Parker, V. C. Gulick, J. S. Kargel, and J. S. Lewis, Channels and valleys on Venus: Preliminary analysis of Magellan data, *J. Geophys. Res.*, *in press*, 1992.

Cox, K. G, J. D. Bell, and R. J. Pankhurst, *The Interpretation of Igneous Rocks*, 450 pp., George Allen & Unwin, Boston, 1979.

Head, J. W., and L. Wilson, Volcanic processes and landforms on Venus: Theory, predictions, and observations, *J. Geophys. Res.*, *91*, 9407-9446, 1986.

Hulme, G., Turbulent lava flow and the formation of lunar sinuous rilles, *Modern Geology*, *4*, 107-117, 1973.

Hulme, G., The Interpretation of lava flow morphology, *Geophys. J. R. Astr. Soc.*, *39*, 361-383, 1974.

Hulme, G., A review of lava flow processes related to the formation of lunar sinuous rilles, *Geophysical Surveys*, *5*, 245-279, 1982.

Huppert, H. E., and R. S. J. Sparks, Komatiites I: Eruption and flow, *Journal of Petrology*, *26*, 694-725, 1985.

Kargel, J. S., S. K. Croft, J. I. Lunine, and J. S. Lewis, Rheological properties of ammonia-water liquids and crystal-liquid slurries: Planetological applications, *Icarus*, *89*, 93-112, 1991.

McBirney, A. R., and T. Murase, Rheological properties of magmas, *Ann. Rev. Earth Planet. Sci. 1984*, *12*, 337-357, 1984.

Moore, H. J., D. W. G. Arthur, and G. G. Schaber, Yield strength of flows on the Earth, Mars, and Moon, *Proc. Lunar Sci. Conf. 9th*, 3351-3378, 1978.

Murase, T and A. R. McBirney, Viscosity of lunar lavas, *Science*, *167*, 1491-1493, 1970.

Oberbeck, V. R., W. L. Quaide, and R. Greeley, On the origin of lunar sinuous rilles, *Modern Geology*, *1*, 75-80, 1969.

Scarfe, C. M., D. J. Cronin, J. T. Wenzel, and D. A. Kauffman, Viscosity-temperature relationships at 1 atm in the system diopside-anorthite, *Amer. Mineral.*, *68*, 1083-1088, 1983.

Shaw, H. R., Rheology of basalt in the melting range, *Journal of Petrology*, *10*, 510-535, 1969.

Surkov, Yu. A., L. P. Moskalyeva, O. P. Shcheglov, V. P. Kharyukova, O. S. Manvelyan, V. S. Kirichenko, and A. D. Dudin, Determination of the elemental composition of rocks on Venus by Venera 13 and Venera 14 (preliminary results), Proc. 13th Lunar and Planet. Sci. Conf., *J. Geophys. Res.*, *88* (supplement), A481-A493, 1983.

Swanson, D. A., T. L. Wright, and R. T. Helz, Linear vent systems and estimated rates of magma production and eruption for the Yakima Basalt on the Columbia Plateau, *Amer. J. Sci.*, *275*, 877-905, 1975.

Touloukian, Y. S. and E. H. Buyco, *Thermophysical Properties of Matter*, 5: Specific Heat, Nonmetallic Solids, 1649 pp., IFI Plenum, New York, 1970.

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