Thermorheological Modeling of Venusian Canali

by

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Abstract

Venusian canali are long, narrow channels on the Venusian plains that show distinctive fluvial characteristics. Many mechanisms have been put forward to explain their formation, with the most likely candidates being thermomechanical erosion by a low viscosity lava. Using the FLOWGO thermorheological modeling program, flow distances of rhyolite, tholeiite, and komatiite lavas were tested under Venusian conditions over a range of flow depths and bulk densities. Rhyolite is too viscous to flow any considerable distance. Komatiite is found to cool too quickly to travel the distances of Venusian canali. At high effusion rates, tholeiite basalt is capable of traveling the distances of Venusian canali under laminar flow conditions. Turbulent flow is generally considered a requirement to achieve helical flow conditions necessary to form fluvial morphological features. However, it is unclear if tholeiite is capable of eroding canali through laminar flow, or if tholeiite is capable of flowing the same distances under turbulent flow conditions. Carbonatite may still be the strongest candidate for canali formation.

Keywords: Venus; canali; thermorheological modeling; Baltis Vallis; lava flows; channels
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Introduction

Magellan synthetic aperture radar (SAR) mapping revealed numerous types of channel features distributed across the surface of Venus. One type of channel have been termed “canali” and are characterised by their long (few 100s to over 7000 km), narrow (1-5 km wide), sinuous nature, and are predominantly found on the smooth volcanic plains (Baker et al., 1992). These channels show distinctively fluvial features, like meanders, point bars and oxbows (Williams-Jones et al., 1998). Diverse formation mechanisms have been put forward for canali construction, including thermomechanical erosion from komatiite (Baker et al., 1992) or tholeiite basalt lavas (Baker et al., 1992; Komatsu et al., 1992), constructive emplacement of tholeiites (Gregg and Greeley, 1993), mechanical erosion from carbonatite lava (Baker et al., 1992; Kargel et al., 1994; Williams-Jones et al., 1998), liquid sulphur (Baker et al., 1992) and sediment gravity flows (Jones and Pickering, 2003; et al., 2008) or thermal erosion from a subsurface fluid flow via lava piping (Lang and Hansen, 2006). None of these hypotheses have been conclusively ruled out, and pieces of morphological evidence support differing interpretations. The measured meander wavelengths of canali support the interpretation that they were formed from a turbulently flowing, low viscosity fluid (Bray et al., 2007).

At 7818 km in length, Baltis Vallis is the longest canali on Venus (Bray et al., 2007). Any fluid capable of eroding Baltis Vallis must be able to remain fluid long enough to travel this distance. The purpose of this study is to determine if any silicate lavas are capable of remaining fluid over these distances. To accomplish this, I use the FLOWGO program, which calculates thermological and rheological parameters for a given flow at set increments down a pre-existing channel.

1.1 Venusian Geology

Venus is a highly dynamic planet, but it is difficult to study. Unlike Mars, or the Moon, its surface is covered in a veil of thick clouds, rendering visible light observations impossible. Instead, surface mapping is performed based on gravitational anomalies,
SAR, and altimetry, which are used to observe and study topography and geological processes at play. While there is an absence of the tectonic plates akin to those observed on Earth, there is a plethora of active volcanism, plateau uplift, and extensional and compressional stresses, providing a wealth of unique tectonism that is not plate driven. The most recent mapping mission by the spacecraft Magellan (1990-1994), provided SAR coverage of ~98% of the planet. Through these data, the analysis of physical features, such as plateau, lava flows, ridges and shield volcanos and other features becomes possible.

Venus is often considered Earth’s ‘twin planet’ due to similarities in size, but it is physically quite different. The mass of Venus is 4.87 x 10^{24} kg – 0.815 of Earth’s mass. Similarly, Venus’s surface gravity at the equator is 0.905 that of Earth, at 8.87 m/s^2. The difference between Venus/Earth mass and gravity can be attributed to Venus’ smaller radius of 6051.8 km at the equator.

One of the most striking features of Venus compared to Earth are the atmospheric conditions. Average atmospheric density at mean surface level (6051.5 km determined from average surface distance from Venus’s centre of mass) is 67 kg/m^3, and the surface pressure is 9.3 MPa: equivalent to being almost a kilometre deep on one of Earth’s oceans. The increased pressure is due to the composition of the atmosphere. The Venusian atmosphere is composed of 96.5 mol% CO_2, followed by 3.5% N_2. In comparison, Earth’s atmosphere is 78.1% N_2 and 20.9% O_2, with the remaining as traces of other gases. The Venusian atmosphere contains minor traces of SO_2, COS, H_2O, CO, helium, neon, argon, and krypton in decreasing order, measured in parts per million. Although atmospheric clouds travel upwards of 100 m/s, wind speeds at the surface are significantly slower, between 0.3-1 m/s (Basilevsky and Head, 2003). Whereas the upper cloud layer comprises sulphuric acid, the composition of the middle and lower cloud decks are less certain.

Unlike Earth, which is generally hot at the equator (avg. 303 K) and cold at the poles (avg. 242 K at North Pole), and warm during daytime and cool at night, Venus experiences very little variation in temperature through latitude and orientation to the Sun. These variations are 2 degrees Kelvin at most. However, as on Earth, temperature varies with elevation. Temperatures range from 650 K at the highest summits of Maxwell Montes to 755 K at the lowest depths of trenches, with a gradient of
approximately 8 K/km (Fig. 1) (Seiff et al., 1980; Hedin et al., 1983; Basilevsky and Head, 2003). In comparison, the U.S. Standard Atmosphere documents a 6.5 K/km average environmental lapse rate for Earth (Sissenwine et al., 1962). At the planetary datum (mean surface level), Venusian temperatures are approximately 740 K (Basilevsky and Head, 2003).

![Venus Magellan SAR Mosaic](image)

*Figure 1: Global Map of Venus from Magellan SAR (Harrington and Treiman, 2015). Venusian topography correlates to radar brightness. Radar-bright areas (white) correspond with highlands, whereas the radar-dark areas are volcanic plains (orange).*

Most of Venus, approximately 70%, is covered in smooth plains, ~10% of the surface is highland, mountainous regions, and ~20% are lowland depressions. The volcanic plains surface is made up of series of lava flows, with younger flows visibly covering older flows in undeformed areas. The lava flows from the volcanically active planet leads to resurfacing, such that the surface of Venus is relatively young on a geological scale (Taylor, 2006). A major resurfacing event is thought to have occurred on Venus ~700 Ma, which transitioned the planet from a period of thin lithosphere, to thick lithosphere, via uncertain mechanisms (Phillips and Hansen, 1998); although this age has been recently challenged and the plains surface could be as young as 200 Ma (Smrekar et al., 2016).
The Soviet Venera 8 lander recorded the density of the surficial rock at 2.7-2.9 g/cm$^3$, with U, Th, and K levels in the rock comparable to terrestrial basalt, (Taylor, 2006; Aitta, 2012). No other rock types have been identified with certainty, but the morphology of some lava flows suggest a more viscous behaviour than those typically seen in basalt (Nimmo and McKenzie, 1998). Venera 13 revealed rocks similar in composition to olivine nephelinite and olivine leucite (Surkov et al., 1983; Baker et al., 1992; Komatsu, 1993). Potassium levels at the Venera 8 landing site are also consistent with an alkali, silica-undersaturated basalt (Keldysh, 1977; Baker et al., 1992).

1.2.1 Volcanism

Volcanoes cover ~4% of Venus’s surface (Basilevsky and Head, 2003), and most Venusian surface features can be linked in some way to volcanism (Taylor, 2006). Unlike Earth, volcanoes on Venus are not distributed in chains (such as the Pacific ring of fire) which are related to volcanic arcs over subduction zones, mid-ocean ridges, and continental rifts. Instead, in the absence of subduction zones, they are widely distributed around the planet. This is due to their volcanic origin from mantle plume hotspots rather than continental boundaries (Nimmo and McKenzie, 1998). Various types of volcanoes have been identified, with many being forms of shield or dome volcanoes (Solomon et al., 1992; Taylor, 2006). Some of the largest volcanos on Venus are shield volcanoes, which range up to 1000 km in diameter, and have lava flows which can extend for hundreds to thousands of kilometers, (Nimmo and McKenzie, 1998; Taylor, 2006).

Because Venus has a higher ambient temperature, lava flows lose less heat from radiation. In contrast, because the Venusian atmosphere is denser than Earth’s, and CO$_2$ at Venusian temperatures has a higher heat capacity than N$_2$ at terrestrial temperatures, surface winds will be more efficient at removing heat from lava flows via convection (Bridges, 1997). We might expect to see quicker cooling for Venusian flows than for terrestrial equivalents.
1.2.2 Volcanic Plains

Venus’ global history has been divided into two main periods; the first was a time of a thinner lithosphere, until about 200-700 Ma, when a global resurfacing event occurred, followed by a time of thickened lithosphere, which extends to the present (Phillips and Hansen, 1998; Smrekar et al., 2016). One of the features formed in the thin-lithosphere period and resurfacing event are volcanic plains. 65-70% of the planet's surface is covered in volcanic plains, formed from multiple lava flow events due to the convection of upwelling mantle erupting at surface. According to the resurfacing hypothesis, hot mantle plumes rise and undergo pressure-release melting under the lithosphere, which can breach the surface (Phillips and Hansen, 1998). The eruptions which formed these volcanic plains were extremely large, equivalent to the Large Igneous Provinces on Earth, making them comparable to the Deccan Traps and Columbia River flood basalts in scale and likely even greater. Recent analysis of impact crater degradation has suggested that the volcanic plains may be younger than previously thought, and possibly formed from multiple smaller events (Smrekar et al., 2016).

1.3 Venusian Canali

The Magellan mission revealed unique channel-like features on the Venusian plains (Fig. 2). These features, termed canali are narrow, long, channels of constant width that meander across the volcanic plains showing distinct fluvial characteristics (Fig. 3,4). Canali are differentiated from other channel features, like sinuous rilles, by their greater lengths and width-to-depth ratios (Baker et al., 1992). Canali widths range from 0.25 to 3.8 km, while their lengths range from a few hundred kilometres up to 7181 km (Bray et al., 2007) (Fig. 4). The average width to depth ratios of the canali, Baltis Vallis, is ~47, making it greater than terrestrial meandering rivers with floodplains (which have width to depth ratios <12), and more comparable in morphology to entrenched meandering riffle/pool channels without floodplains (>12) (Rosgen, 1994; Oshigami and Namiki, 2007). Canali sinuosity is highly variable, from 1.01 to 3.75, making it difficult to
compare with terrestrial rivers that are characterized by having high sinuosity if sinuosity >1.4, low sinuosity if <1.2, and moderate values in between. (Rosgen, 1994; Bray et al., 2007). The source regions for canali are indistinct at the resolution of Magellan SAR. Although canali are globally distributed across Venus, they are primarily developed on the lowland volcanic plains (Fig. 2).

Figure 2: Distribution of canali across Venus. Modified after Williams-Jones et al., (1998). Fig. 3 indicated by the star. Figs. 4 and 5 locations indicated by boxes north of Aphrodite Terra.

Some fluvial-like characteristics demonstrated by canali include meanders, cutoffs, oxbows, and abandoned paleochannels. However, meanders within Venusian canali have higher radius of curvature and wavelengths per channel width than terrestrial rivers (Baker et al., 1992). Meander wavelengths of the canali range from 7 to 372 km, with typical ranges between 11 to 77 km. Canali have depth to width ratios around 0.0013, and bank slopes of approximately 6°, which is comparable to alluviually bedded, low gradient rivers on Earth (Bray et al., 2007).
Figure 3: A canale showing meander oxbows, cut-offs, and a delta-like feature. Canale is located on Sedna Planitia is located at approximately (45°N, 019°E) See Fig. 2 for location. Black arrows indicate fluvial-like features. The leftmost arrow could be a kipuka if a silicate melt formed the canali. The northmost arrow could be either an oxbow or a levee structure. Modified after Williams-Jones et al., (1998).
Figure 4: Segment of Baltis Vallis, the >7000 km canale north of Aphrodite Terra. Arrow indicates what appears to be a channel bar or levee, where Baltis Vallis appears to anastomose. Image centre at approximately (042°N, 161°E). See Fig. 2 for location.
1.3.1 Baltis Vallis

At over 7181 km in length, longer than the Nile River, Baltis Vallis is the longest channel identified on Venus (Bray et al., 2007). Baltis Vallis is a canali-type simple channel, located north of Rusalka Planitia and Aphrodite Terra (Figs. 1, 2, 5). Measurements of Baltis Vallis range from 17.3-105.4 m in depth, averaging 46 m deep. Its channel width ranges between 0.9-3.4 km, averaging 2.2 km, but is fairly constant throughout its length (Bray et al., 2007; Oshigami and Namiki, 2007). The exact terminus and source are obscured in radar imagery (Oshigami and Namiki, 2007), but its source is located at 44.5°N, 185°E and it terminates at 11.5°N, 167°E. Although uncertain, the source has been suggested to be a 150 km diameter volcanic formation at the aforementioned coordinates. The terminus is obscured in a radar-dark patch which is likely a volcanic deposit formed from channel materials (Baker et al., 1992). There are no other channels in proximity to Baltis Vallis, and the canali itself comprises a single channel that meanders, and possible anastomoses (Fig 4). Baltis Vallis is found on the volcanic plains, which generally range in gradient from 0.032-0.159°. However, more recent tectonic deformation has resulted in segments of the channel appearing to flow uphill at 1.5°, so modern slopes are not reliable along all intervals of the channel. In terms of relative age, Baltis Vallis cross-cuts the volcanic plains and coronae (100s of km wide, deformed oval-shaped features) and in turn is disrupted by meteorite impact craters (Baker et al., 1992).

The highest concentration of levee structures is in the first quarter of the channel from the source, whereas there is stronger evidence of erosion farther down the channel. The distribution of levees indicates that construction played a stronger role in formation of the upper channel, and then waned to predominately erosional processes downstream. The highest concentration of intra-channel ridges (similar to terrestrial river bars) is in the second quarter of the channel, which suggests that sediment deposition influenced channel morphology during this section. All segments of Baltis Vallis are dominated by groove-like morphology (Oshigami and Namiki, 2007). Baltis Vallis also meanders and shows fluvial characteristics such as anastomosing channels (Fig. 4).
Figure 5: Geological map showing extent of Baltis Vallis. Greyscale elevations are from a reference planetary radius of 6051 km. t = tessera terrain, pwr = plains with wrinkle ridges, ptr rb = fractured and rigid plains, ridge belts. Modified after Oshigami and Namiki (2007).
1.3.2 Canali Formation Mechanisms

Some estimates have claimed that the scale of lava eruption required to produce flows 100s to 1000s of km long can only be explained by planetary-scale volcanism (Komatsu, 1993), such as the global resurfacing event 200-700 Ma years ago (Philips and Hansen, 1998; Smrekar et al., 2016). Because the canali cut into the volcanic plains, the canali must have formed either syn-eruptively, or after this resurfacing event. If the canali are syn-eruptive, then they may represent low volumetric discharge outlets for a type of low-viscosity lava. Multiple hypotheses have been put forward to explain the origin of the canali. These hypotheses can be split into two groups: mechanical erosion, and joint thermomechanical erosion. Different fluids have been suggested as being able to mechanically abrade and erode the channels like water in terrestrial rivers, from ultramafic silicate lavas to more exotic melts like sulphur flows and carbonatite lavas (Baker et al., 1992).

Silicate melts hotter than the melting point of the tholeiitic basalt (tholeiitic lava, komatiite) would be able to partially melt the volcanic plains, and entrain plains rock into the melt. A few case studies of thermomechanical erosion have been observed on Earth, including Kilauean tube-fed lavas. It was recorded that erosion rates were highest after pauses in the eruption, such that thermal erosion was supplemented by mechanical erosion of debris that had collected in the channel, with average rates of 10 cm/day (Fagents and Greeley, 2001). Thermomechanical erosion by Archean komatiites has been documented in the Norseman-Wiluna greenstone belt of Western Australia. This komatiite flow eroded a channel up to 150 m deep in a felsic substrate (Williams et al., 2001). Purely thermal erosion has been observed via carbonatite flowing laminarly over a carbonatitic substrate, but the cooler temperatures of carbonatite melt would not be able to melt a basaltic substrate (Fagents and Greeley, 2001). The ability for a melt to thermomechanically erode into deep channels depends on the slope, effusion rate, duration of flow, temperature, volatile content, rheology, and compositions of the lava and abraded material (Williams et al., 1999; Schenk and Williams, 2004).

Thermomechanical erosion is hypothesised to have created channels on Mars, the Moon, and Io. Tawhaki Vallis, a channel on Jupiter’s moon, Io, is projected at being over 250 km long, between 0.5-6 km wide, and 40-65 m deep (Schenk and Williams,
2004). Erosion and iterative cooling modeling has shown that lava is certainly capable of rapidly eroding lunar sinuous rilles and could have formed some types of Martian channels (Carr, 1973). The channels emerging from Elysium Mons on Mars are particularly suitable candidates for the thermal erosion model (Wilson et al., 2001). Although lunar rilles have lower meander wavelengths and radius of curvature than Venusian canali, and the Martian channels from Elysium Mons are considerably shorter than the canali (<20 km vs. 100s-1000s of km), these examples nonetheless demonstrate the ability of lavas to perform thermomechanical erosion at sustained durations to form deep channel features. In discussing the formation of the Tawhaki Vallis on Io (250 km long, 40-65 m deep), abrasion by silicate lava over a silicate country rock would require an eruption lasting days to months, whereas a silicate or sulphur flow over a frozen SO₂ substrate would only require hours to days to form (Schenk and Williams, 2004). However, the basaltic nature of Venusian plains makes a frozen SO₂ substrate unlikely, and thus a longer duration of channel formation would be expected.

Liquid water cannot be conclusively ruled out as the erosional agent (Baker et al., 1992). Today, water constitutes only 30 ppm of Venusian clouds, but the high deuterium/hydrogen ratio in Venusian water suggests that early Venus conditions may have had more water than at present. Venus could have had enough atmospheric water to completely cover its surface 4-115 m deep if condensed on the surface within the past billion years. High water contents could be explained by a massive volcanic outgassing event, which was then lost through vapourization in the greenhouse atmosphere (Donahue and Russell, 1997). Modern Venusian surface temperatures are too hot for liquid water, and at 90 bar pressure at Venus’s surface, water requires temperatures less than 580°K to be stable (Baker et al., 1992). Water is not a good fit for the higher abundance of constructional levees along the first quarter of Baltis Vallis. In terrestrial rivers, levees tend to be more concentrated in the downstream segments of a river, rather than near the channel source (Oshigami and Namiki, 2007). Furthermore, the association of canali with volcanic deposits and structures better supports a volcanic origin for the canali (Baker et al., 1992).

Early modeling by Baker et al. (1992) shows that lava with tholeiitic composition is unlikely to have remained fluid and erosive for the lengths required to erode Baltis Vallis. Without an insulation mechanism, at depths of 1-5 m, flow velocities <10 m/s,
and thermal diffusivity of $10^6 \text{ m}^2/\text{s}$, tholeiitic lavas are projected to travel a maximum flow distance of 926 km before solidifying (Baker et al., 1992). Tholeiite flow modeling by Komatsu et al. (1992) achieved flow distances of 6800 km using flow thicknesses of ~80 m, which is greater than the present depth of Baltis Vallis, but we do not know if the canale is filled with a solidified lava of any known thickness. Olivine leucitite and nephelinite on Venus, as analysed by Venera 13, would have a viscosity lower than terrestrial basalt, and may be able to travel longer distances in a channel (Kargel et al., 1993; Kargel et al., 1994) At Venus surface temperatures of 745.9 K, it is unlikely that any silicate lava would remain fluid for long enough to erode the entire 7181 km length of Baltis Vallis. Tholeiitic basalt erupting at 1473 K is able to completely crystallize below 1270 K (Schmincke, 2004) and would experience glass transition at ~903 K (Giordano et al., 2008; Chevrel et al., 2014), while komatiite will crystallize at even higher temperatures, ranging from 1473 up to 1700 K (Huppert and Sparks 1985; Baker et al., 1992). The glass transition for a komatiite at eruption temperatures of 1873 is ~915 K (Huppert and Sparks, 1985; Giordano et al., 2008). Whereas komatiite lavas have very low viscosities, they have higher melting temperatures and would solidify more rapidly, making them unlikely to be capable of forming the long canali. Early calculations regarding lava flows on Venus calculated that lavas would cool approximately 10% slower on Venus than for lava of the same composition on Earth because of the smaller $\Delta T (T_{\text{eruption}} - T_{\text{ambient}})$ (Head and Wilson, 1986). In contrast, more felsic, siliceous lavas have lower melting temperatures, but are considerably more viscous, and are expected to be unable to flow fluidly to form a fluvial-shaped channel.

Sulphur volcanism has been suggested on Venus due to the high abundance of sulphurous gases in the Venusian atmosphere, as well as the weight percent of S found in surface samples by the Venera landers. The viscosity of sulphur is highly variable, depending on temperature and degree of polymerization. Fluid behaviour of sulphur can range from that of andesitic melts to considerably more fluid and water-like. A highlight of the sulphur hypothesis is that liquid sulphur has a crystallization temperature of 393 K under Venusian conditions, which is well below Venusian surface temperatures (Williams-Jones et al., 1998). However, sulphur can volatilize between 750 and 1262 K, and Venusian ambient temperatures are around 750 K, meaning that the sulphur might evaporate (Baker et al., 1992; Williams-Jones et al., 1998). It is possible that sulphur
erupting above ambient conditions would lose most of its volume from vapourization before cooling below the condensation temperature (Baker et al., 1992). Even if Venusian ambient conditions are below the evaporation point for sulphur, the fact that sulphur will not crystallize under Venusian temperatures shows that it cannot form the levees found at some points along Venusian canali like Baltis Vallis (Oshigami and Namiki, 2007).

Carbonatites are another exotic lava type that has been put forward. They are an attractive suggestion because their solidification point of 764 K is just slightly higher than Venusian ambient surface temperatures (Krafft and Keller, 1989; Kargel et al., 1994). On Earth, they generally form from the partial melting of carbonated ultramafic peridotite and immiscible segregation between carbonatic fluids and alkaline silicate melts. The latter mode of genesis is consistent with a sample from the Venera 13 lander, which has a similar elemental signature to olivine nephelinite (Baker et al., 1992). Some models suggest that Venusian crustal rock could have high percentages of carbonate and sulphate minerals, which could be melted from endogenic or impact heat addition. The idea of subsurface carbonatite aquifers could explain multiple Venusian morphological features, including not only canali and their floodplains (from large, long duration eruptions), but also crater outflows, sapping vallys, outflow channels, and chaotic terrain (Kargel et al., 1994).

### 1.4 Terrestrial Analogues

No documented terrestrial lava flows come close in length to the >7000 km exhibited by Baltis Vallis. However, the longest and largest lava flows on Earth are quite substantial. The Rajahmundry Trap lava flows, extensions of the late Deccan volcanism, have been recorded at approximately 1000 km in length (Self et al., 2008). These flows were confined to some form of pre-existing channels (in this case, the Krishna valley drainage system), like the Venusian canali if the mechanical erosion hypothesis is correct. The Rajahmundry Trap is composed of lobes up to 25 m in thickness. Downflow cooling rates have been modeled from virtually no internal cooling at 25 m thickness, to -0.1 °C/km for lobes 10 m thick (Self et al., 2008). The Deccan Traps provide the highest endmember for terrestrial volcanism. Flood basalt eruptions on
Earth can have effusion rates anywhere from 370 to 37,000 m$^3$/s (Bryan et al., 2010). The 1783 Laki eruption in Iceland had the highest effusion rate in recorded history, with the peak volumetric output estimated to be at a maximum of 6600 m$^3$/s (Thordarson, 2003).

1.5 Purpose of Study

The purpose of this study is to attempt to constrain which fluid types are able to mechanically erode Venusian canali by modeling different lava flow properties under Venusian conditions. Whereas the Rajahmundry Trap lava flow lengths were inhibited by terrain morphologies, and terminated on reaching the ocean, Venusian canali would only be limited by cooling rates and lava viscosities as well as the extent of the pre-existing channel if they formed from incremental erosion (Self et al., 2008). Baltis Vallis, the longest canali on Venus, is used as an endmember to help determine which types of melts would be able to remain fluid for long enough and flow far enough in order to mechanically erode this, and other, canali on Venus. By testing different hypothesised melts (rhyolite, basalt, komatiite) using more comprehensive thermorheological modeling, we can determine if any silicate melts could be viable candidates for canali construction.
Methodology

2.1 FLOWGO Modeling

The model used for this experiment, named FLOWGO, was developed by Harris and Rowland (2000) for kinematic thermorheological flow modelling for channelized lava. The aim of FLOWGO is to determine how far channelized and leveed lava is able to flow, for applications in studying the thermal and rheological properties of confined lava. The model works by calculating core cooling and crystallization within an aliquot of lava, and its subsequent changes in temperature, rheology and velocity as it flows down slope in the channel, until the lava slows to a complete stop (Harris and Rowland, 2001). The model has been adapted for use in Microsoft EXCEL, which is the platform used here. FLOWGO works through four main steps: determining initial velocity; determining initial rheology; calculating cooling and crystallization rate; and calculating the changes in the control lava volume as it evolves down the channel. Velocity and shear strength are both calculated as functions of viscosity, which itself is a function of temperature, initial eruptive viscosity, and degree of crystallization. The cooling and crystallization rate determine how temperature changes down channel. The model is executed until the either the core temperature reaches the stop temperature, or the velocity reaches zero, causing the lava to stop flowing (Harris et al., 2015).

Rowland et al. (2004) extended the use of the FLOWGO program beyond terrestrial conditions. They studied the effects of gravity, ambient temperature, and atmospheric conditions on channelized lava flows on Mars compared to terrestrial analogues. They observe that the lower gravity and the thin CO₂ atmosphere has the greatest influence on increasing basaltic flow lengths on Mars compared to Earth. With Venus’ slightly smaller gravity than Earth, and a thick CO₂ atmosphere, we may expect to see similar results, although our primary goal is to test different silicate lava types under Venusian conditions. Nonetheless, in our approach, we aim to employ the methods of Rowland et al. (2004) by comparing Venusian and terrestrial conditions, to evaluate how ambient atmospheric conditions and gravity will affect flows.
2.2 Parameters Used

Gravitational acceleration at mean Venus radius is 8.87 m/s² (NASA, 2016). This is slightly lower than that on Earth, and over twice the gravity of Mars. Gravity is one of the few parameters that will remain constant throughout the study.

There are areas along the profile of Baltis Vallis along which it appears to run uphill. This is likely the result of post-formational uplift, indicating that present-day slopes may not represent reliable parameters (Baker et al., 1992; Komatsu and Baker, 1994). Present slopes on the volcanic plains with canali are very low, ranging from 0.032 to 0.159°, or 0.5 m/km to 2.5 m/km (Williams-Jones et al., 1998). Using stereotopography from Herrick (2012), some segments along Baltis Vallis ostensibly seem to have slopes as high as 1.5° in the downslope direction. The volcanic plains of Baltis Vallis lie below the mean planetary radius of 6051.0 km. In the areas we measured, elevations ranged between ~100 to ~1350 m below the mean planetary radius, with average elevation at ~740 m. We sourced elevations from Herrick (2012) stereotopography, which does not contain the entire extent of Baltis Vallis. Therefore, the actual range in elevation is likely to be slightly greater, and the average elevation may differ. This range in elevation for the volcanic plains is likely due to post-canali tectonic deformation, as some segments of the channel appear to be flowing uphill. Nonetheless, for modelling purposes, variation in elevation produces minor variation in surface temperature which is expected to have minimal effect on lava flow distances. Similarly, plains topography was likely smoother at time of channel formation, reducing topographic effects of flow velocity, turbulence and rheology. At mean planetary radius, Venus has a surface temperature of ~740 K. The temperature gradient at surface is ~8 K km⁻¹, giving temperature ranges of 740.8 – 750.8 K, for Baltis Vallis, with an approximate average of 745.9 K.

Channel widths and depths for 120 points along Baltis Vallis have been provided from the analysis undertaken by Oshigami and Namiki (2007).

In the first stage of the study, volumes of the Laki eruption were input into the model for tholeiitic lava to see if terrestrial scale outputs are capable of forming canali under varying stages of terrestrial to Venusian conditions. Calculations using terrestrial
conditions have suggested that basaltic lava flows greater than 100 km in length could be produced through either rapid emplacement (quickly moving a’a flows) or through slower moving, insulated flows (with the development of a full crust). For the rapidly emplaced flows, effusion rates of 3100 to 11,000 m$^3$/s, velocities of 4-12 m/s, and flow thicknesses (depths) of 3-17 m are required to attain these lengths. An insulated flow could have lower velocities of 0.2-1.4 m/s at higher thicknesses of 6-23 m, and effusion rates anywhere from 50-7100 m$^3$/s (Keszthelyi and Self, 1998). For one of our modeling trials, we use the upper limits of the “rapidly emplaced flow” conditions to determine the maximum flow length attainable using reasonable parameters.

Previous basalt lava modeling by Harris et al. (2015) on Earth used crystal parameters of 10.4% phenocrysts in eruptive lava (which have calculated increases downflow), and insulating lava crustal cover of 60% on the flow surface. In studying a basaltic flow on Mars, Rowland et al. (2004) used 0% phenocrysts and 100% crustal coverage. For our study, we use the terrestrial parameters.
Table 1: Input parameters for tholeiite basalt in Venus simulations (modified after Harris et al., 2015).

<table>
<thead>
<tr>
<th><strong>VENUS INPUT PARAMETERS</strong></th>
<th><strong>THERMAL PARAMETERS</strong></th>
<th><strong>THERMAL PARAMETERS</strong></th>
</tr>
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<tr>
<td><strong>Channel Dimensions</strong></td>
<td><strong>Terupt</strong></td>
<td>1199.85 C 1473 K</td>
</tr>
<tr>
<td>Width (Initial Channel)</td>
<td>200 m</td>
<td></td>
</tr>
<tr>
<td>Depth</td>
<td>25.00 m</td>
<td>Variable in different trials</td>
</tr>
<tr>
<td>Down-flow increment</td>
<td>2000 m</td>
<td>Arbitrary</td>
</tr>
<tr>
<td><strong>Tcrust</strong></td>
<td></td>
<td>500 C 773.15 K</td>
</tr>
<tr>
<td><strong>Buffer</strong></td>
<td></td>
<td>140 K Δ(Tcore - Tsurface)</td>
</tr>
<tr>
<td><strong>Velocity Constants</strong></td>
<td></td>
<td>d (a constant) -7.56E-03</td>
</tr>
<tr>
<td>g</td>
<td>8.87 m s-2</td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>3</td>
<td></td>
</tr>
<tr>
<td><strong>Viscosity and Yield Strength (YS) Parameters</strong></td>
<td><strong>Density and Vesicularity</strong></td>
<td></td>
</tr>
<tr>
<td>Fluid Viscosity</td>
<td>800 Pa s</td>
<td>Fixed (Lava viscosity changes down flow)</td>
</tr>
<tr>
<td>a</td>
<td>0.04 K-1</td>
<td></td>
</tr>
<tr>
<td>b</td>
<td>0.01 Pa</td>
<td></td>
</tr>
<tr>
<td>c</td>
<td>0.08 K-1</td>
<td></td>
</tr>
<tr>
<td><strong>Stefan-Boltzmann Constant</strong></td>
<td>5.67E-08 W m-2 K-4</td>
<td></td>
</tr>
<tr>
<td>emissivity (e)</td>
<td>0.9</td>
<td></td>
</tr>
<tr>
<td><strong>Radiation Parameters</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Wind Speed</strong></td>
<td>0.3 m s-1</td>
<td>based off Venus surface</td>
</tr>
<tr>
<td><strong>CH (Wind Friction Factor)</strong></td>
<td>0.0036</td>
<td></td>
</tr>
<tr>
<td><strong>Tair</strong></td>
<td>472.75 C 745.9 K</td>
<td></td>
</tr>
<tr>
<td><strong>Air Density</strong></td>
<td>67 kg m-3</td>
<td>CO2 at Venusian T-P</td>
</tr>
<tr>
<td><strong>Air Specific Heat Cap</strong></td>
<td>1148 J kg-1 K-1 CO2 at Venusian T-P</td>
<td></td>
</tr>
<tr>
<td><strong>Convection Parameters</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Crystal Parameters</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Phenocrysts</td>
<td>0.104</td>
<td></td>
</tr>
<tr>
<td>Post eruption xtabs</td>
<td>0.896</td>
<td></td>
</tr>
<tr>
<td>Cooling Range</td>
<td>123 K</td>
<td>Δ(Terupt - Tsolidus)</td>
</tr>
<tr>
<td>Rate of crystallization</td>
<td>0.007284553</td>
<td>fractional xtablization per K</td>
</tr>
<tr>
<td>Latent Heat of xtablation</td>
<td>4.20E+05 J kg-1</td>
<td></td>
</tr>
<tr>
<td>R</td>
<td>1.51</td>
<td></td>
</tr>
</tbody>
</table>
Table 2: Input parameters for tholeiite basalt in Earth simulations (modified after Harris et al., 2015).

<table>
<thead>
<tr>
<th>EARTH INPUT PARAMETERS</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Channel Dimensions</strong></td>
</tr>
<tr>
<td>Width (Initial Channel)</td>
</tr>
<tr>
<td>Depth</td>
</tr>
<tr>
<td>Down-flow increment</td>
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<tr>
<td><strong>Thermal Parameters</strong></td>
</tr>
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<td>Terupt</td>
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<tr>
<td>Tcrust</td>
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<tr>
<td>Buffer</td>
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<tr>
<td><strong>Velocity Constants</strong></td>
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<tr>
<td>g</td>
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<tr>
<td>B</td>
</tr>
<tr>
<td><strong>Viscosity and Yield Strength (YS) Parameters</strong></td>
</tr>
<tr>
<td>Fluid Viscosity</td>
</tr>
<tr>
<td>a</td>
</tr>
<tr>
<td>b</td>
</tr>
<tr>
<td>c</td>
</tr>
<tr>
<td><strong>Density and Vesicularity</strong></td>
</tr>
<tr>
<td>DRE Density</td>
</tr>
<tr>
<td>Vesicularity</td>
</tr>
<tr>
<td>Bulk Density</td>
</tr>
<tr>
<td><strong>Convection Parameters</strong></td>
</tr>
<tr>
<td>Wind Speed</td>
</tr>
<tr>
<td>CH (Wind Friction Factor)</td>
</tr>
<tr>
<td>Tair</td>
</tr>
<tr>
<td>Air Density</td>
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<tr>
<td>Air Specific Heat Cap</td>
</tr>
<tr>
<td><strong>Conduction Parameters</strong></td>
</tr>
<tr>
<td>Thermal Conductivity</td>
</tr>
<tr>
<td>Tbase</td>
</tr>
<tr>
<td>Base</td>
</tr>
<tr>
<td>hbase</td>
</tr>
<tr>
<td><strong>Crystal Parameters</strong></td>
</tr>
<tr>
<td>Phenocrysts</td>
</tr>
<tr>
<td>Post eruption xts</td>
</tr>
<tr>
<td>Cooling Range</td>
</tr>
<tr>
<td>Rate of crystallization</td>
</tr>
<tr>
<td>Latent Heat of xtalization</td>
</tr>
<tr>
<td>R</td>
</tr>
</tbody>
</table>
2.3 Baltis Vallis Channel Morphology

To test the flow capabilities of various lavas, we modeled the lavas flowing through Baltis Vallis in order to determine what melts would be capable of traversing its length under modern Venusian conditions. Oshigami and Namiki (2007) created a series of cross-sectional profiles of the Baltis Vallis canali. The original Magellan altimetry data has a coarse horizontal resolution of 10-20 km, and vertical resolution of 50-100 m (Herrick et al., 2012) - significantly greater than the morphology of canali. Oshigami and Namiki (2007) circumvented this spatial limitation by producing new, short-wavelength topography by applying Muhleman's backscattering function to Magellan SAR. This new topography has a vertical error of only 5 m for a channel width of 1 km. The authors produced channel cross-sections for 120 positions along a ~6000 km extent of Baltis Vallis.

Dr. Oshigami cordially provided us with their channel morphology data from the 2007 study. For our first trial, we use 120 points along a 5732 km extent of Baltis Vallis, at which distance along channel, channel width, and channel depth have been measured. We are able to input these data into the FLOWGO model, using specified distance measurements as opposed to generating 10 m space intervals like previous authors. However, using the actual channel dimensions yields unreasonable results suggesting that it is unlikely that the flows filled the channel completely, and channel formation is likely the result of long-duration sustained flow, or multiple pulses of output.

Rowland et al. (2004) comment on how the FLOWGO model varies channel width downflow in order to keep effusion rate constant as flow velocity decreases. Subsequently, as flow velocity reaches zero, FLOWGO models channel widths expand exponentially. Changes in channel width affects the amount of heat lost through radiation and convection, and so results towards the end of the channel may not be reliable. To counteract these uncertainties, Rowland et al. (2004) “trim” their modeled flows at the point where channel width reaches ten times the initial channel width. For our purposes, we trim our data at the point where channel width reaches 4000 m (twenty times initial width of 200 m) which is the maximum channel width recorded for Baltis Vallis (Oshigami and Namiki, 2007).
2.4 Lava Compositions

For our project, we test the composition of multiple lava types in order determine most probable candidates for canali erosion. We start with silicate lavas, including rhyolite, basalt, and komatiite. Although rhyolite is expected to be far too viscous to travel the distances required to form the canali, they are included to show comparisons between felsic lava flows on Earth and Venus. Basaltic lavas are the most common effusive lava on Earth, and are expected to be predominant on Venus based on compositional data from the Venera 8 and 13 landers. Whereas basalt has been proposed as a candidate for canali formation based on the discovery of abundant basalt at surface, it is hypothesised that the high crystallization temperature of the melt will prevent basalt from remaining fluid long enough to travel the length of the canali. After testing komatiite and the more siliceous lavas, we attempted to model more exotic fluids. Carbonatites are exceedingly uncommon on Earth, but nonetheless are extremely fluid, and are hypothesized to behave in a water-like way to possibly form the fluviatile-like canali. However, due to limitations within the FLOWGO model, we were unable to accurately model low viscosity fluids like carbonatite, as the program output unreasonably high starting velocities and effusion rates.

In order to test different lava flows under Venusian conditions, we need to set parameters for each of the lava types. “R” is a value indicating $1/(\varphi_{\text{max}})$, where $\varphi_{\text{max}}$ is the maximum crystallinity of a lava before the viscosity is too high for lava flow movement. $\varphi_{\text{max}}$ for different lavas is variable with regards to the types of phenocrysts and their crystal shapes (Saar et al, 2001). Instead, general values have been used from the literature based on the approximate fluidity of each lava type: $R = 1.35$ is used for fluid liquids and is estimated for carbonatite (Pinkerton and Stevenson, 1992); $1.51$ for Basalt (Harris et al., 2015); $1.67$ for Rhyolite (Pinkerton and Stevenson, 1992); and the value for komatiites estimated as between tholeiite and carbonatite at $R = 1.4$.

It is expected that Venus’ high atmospheric pressures will inhibit exsolution of volatiles in lava. Because volatiles are less able to exsolve and escape the melt, Venusian lavas have lower vesicularity, and higher bulk densities than terrestrial lavas of the same composition. The effects of pressure on lava have been observed in seafloor
lavas on Earth. We have selected the vesicularity vol % values used by Bridges (1997) for rhyolite and basaltic lavas on Earth and Venus. Comparitively, the terrestrial lavas have vesicle volume of 30%, whereas the Venusian lavas will have 10% vesicles by volume. The inability for lavas to lose volatiles through vesiculation will in turn reduce viscosities of the lavas, due to increased volatile contents. It is expected that a rhyolite on Venus would have only 10% of the viscosity of a terrestrial rhyolite (Bridges, 1997).

The parameter of “cooling range” is the difference between extrusion temperature and the lava’s solidification temperature. It has been calculated for each of the lavas in the FLOWGO Excel sheets using the temperatures given in Table 3. The temperatures at the base of the flows are less certain.

Because of the higher ambient temperature conditions, the surrounding basaltic substrate is hotter on Venus than on Earth. It is therefore expected that Venusian lava flows would lose less heat from convection than identical flows on Earth. Rowland et al. (2004) calculated that differences in ambient temperatures on flows to minimally affect flow length for flows between Earth and Mars. To partially account for the differences between Earth and Venus, ΔT_{base} (the difference in temperatures between the base of the flow and its core) has been reduced by 200 K for our basalt and komatiite flows. The values for the carbonatites is the difference between the ambient temperature and core flow, which is a minimal 27.3 K. As predicted from Rowland et al.’s (2004) observations, changing the temperature of the flow base has very minimal effect on flow length.
Table 3: Physical properties used for modeling different lavas.

<table>
<thead>
<tr>
<th></th>
<th>Unit</th>
<th>Rhyolite&lt;sup&gt;a&lt;/sup&gt;</th>
<th>Tholeite&lt;sup&gt;ac&lt;/sup&gt;</th>
<th>Komatiite&lt;sup&gt;abd&lt;/sup&gt;</th>
<th>Alkali Carbonatite&lt;sup&gt;e&lt;/sup&gt;</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Solidus temperature</strong></td>
<td>K</td>
<td>973</td>
<td>1350</td>
<td>1473.15</td>
<td>763</td>
</tr>
<tr>
<td><strong>Extrusion temperature</strong></td>
<td>K</td>
<td>1073</td>
<td>1473</td>
<td>1873</td>
<td>773</td>
</tr>
<tr>
<td><strong>Fluid Viscosity</strong></td>
<td>Pa s</td>
<td>Earth: 1 x 10&lt;sup&gt;8&lt;/sup&gt;</td>
<td>Earth: 1000</td>
<td>0.1 – 10</td>
<td>1 x 10&lt;sup&gt;9&lt;/sup&gt; – 10&lt;sup&gt;1&lt;/sup&gt;</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Venus: 1 x 10&lt;sup&gt;7&lt;/sup&gt;</td>
<td>Venus: 800</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>DRE Density</strong></td>
<td>kg/m&lt;sup&gt;3&lt;/sup&gt;</td>
<td>2200</td>
<td>3000</td>
<td>3200&lt;sup&gt;d&lt;/sup&gt;</td>
<td>2200</td>
</tr>
<tr>
<td><strong>Vesicularity (vol %)</strong></td>
<td></td>
<td>Earth: 30%</td>
<td>Earth: 30%</td>
<td>Earth: 30%</td>
<td>Set to 0</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Venus: 10%</td>
<td>Venus: 10%</td>
<td>Venus: 10%</td>
<td></td>
</tr>
<tr>
<td><strong>Bulk Density</strong></td>
<td>kg/m&lt;sup&gt;3&lt;/sup&gt;</td>
<td>Earth: 1540</td>
<td>Earth: 2100</td>
<td>Earth: 2240</td>
<td>2200</td>
</tr>
<tr>
<td><strong>emissivity</strong></td>
<td></td>
<td>*0.9</td>
<td>*0.9</td>
<td>*0.9</td>
<td>*0.9</td>
</tr>
<tr>
<td><strong>Latent Heat of Crystallization</strong></td>
<td>J/kg</td>
<td>2.9 x 10&lt;sup&gt;5&lt;/sup&gt;</td>
<td>4.2 x 10&lt;sup&gt;5&lt;/sup&gt;</td>
<td>8 x 10&lt;sup&gt;5&lt;/sup&gt;</td>
<td>Earth: 44900 (gas free) Venus: 54800 (gas rich)</td>
</tr>
<tr>
<td><strong>Thermal conductivity</strong></td>
<td>W/m/K</td>
<td>2.5</td>
<td>2</td>
<td>1</td>
<td>1.8</td>
</tr>
<tr>
<td><strong>R</strong></td>
<td></td>
<td>*1.67</td>
<td>*1.51</td>
<td>1.4</td>
<td>*1.35</td>
</tr>
</tbody>
</table>

<sup>a</sup>Bridges (1997)<br><sup>b</sup>Pinkerton & Stevenson (1992)<br><sup>c</sup>Murase and Mc Birney (1970); Basaltic Volcanism Study Project (1981); Scarfe et al. (1983)<br><sup>d</sup>Huppert et al. (1984), Huppert and Sparks (1985), de Bremond d’Ars et al. (1999)<br><sup>e</sup>Glasstone and Lewis [1963]; Janz et al. [1979]; Treiman and Schedl [1983]; Fink et al. [1983]; Naranjo [1985]; Treiman [1989]; Dawson et al. [1990]; Norton et al. (1990) by DSC<br><sup>f</sup>Harris et al. (2015)
Results

3.1 Rhyolite Flows

As predicted, rhyolite flows move extremely slowly on the low slopes of the Venusian plains (0.159°). At channel width of 200 m (to be consistent with later basalt and komatiite trials) and high thicknesses of 150 m, FLOWGO predicts that a rhyolite flow will have a starting velocity of only 0.0226 m/s, and the resulting effusion rates necessary to produce such a flow would be only 679 m³/s. The thickness of 150 m is deeper than the maximum depth of Baltis Vallis, and was modeled to see if the rhyolites were capable of flowing at all. A flow of these parameters is expected to travel only 12.5 km before stopping. Upon stopping, the rhyolite is expected to have a final viscosity of 2.81 x 10⁸ Pa·s, a core temperature of 1035.4 K, and 45.2% crystallinity. A flow of the same composition on Earth (with higher initial viscosity due to exsolved volatiles) would flow less than a kilometer on the same slope.

3.2 Effects of Flow Depth and Bulk Density on Tholeiitic Basalt

Different variables were tested to see how parameters would affect flow lengths using FLOWGO. Two of the most important parameters found to affect the lengths of basalt flows are bulk density (which is set by lava composition and vesicularity) and depth of lava (thickness of flow) (Fig. 6b). One of the longest lava flows we modeled was obtained using bulk density equal to dense rock equivalent (DRE) of 3000 kg/m³ for tholeiitic basalt. Under Venusian conditions, using a flow depth of 25 m (the maximum height of terrestrial Rajahmundry Trap lava lobes, Self et al., 2008) this flow reached 1806 km in length. The effusion rate for this flow is calculated at 62,206 m³/s, which is approximately twice the effusion of the Deccan Traps. If a flow of the same bulk density, fluid viscosity, and depth were to occur on Earth on the same slope (0.159°) and with Venusian atmospheric wind speeds (0.3 m³/s), the flow would extend further to 1932 km. Starting velocities for such a terrestrial flow are 11.01 m/s, with effusion rates of 55,027 m³/s. Such high effusion rates and absence of vesicularity is not expected to occur on
Earth. The complete absence of volatiles is not predicted for Venusian lavas either. Comparisons of core temperature, viscosity, and crystallinity over distance for each Venusian lava flow are shown in Figures 7, 8 and 9 respectively. The final crystallinity for this Venusian flow is 44.6% at the end of the run, and a viscosity of $1.26 \times 10^4$ Pa·s. The minimum core temperature attained at the end of the Venusian flow is 1426.5 K.

Using a slightly less extreme estimate of 10% vesicularity, the tholeiite has a bulk density of 2700 kg/m³. Applying the same flow depth of 25 m, this flow travelled 1488 km down the uniform slope of 0.159°. The initial modeled velocity of this flow is 11.2 m/s. Effusion rates are constant at 55,933 m³/s. The final crystallinity of the Venusian flow is 44.0% and a viscosity of $1.19 \times 10^4$ Pa·s. The minimum core temperature attained at the end of the Venusian flow is 1427 K. A flow of the same depth and bulk density on Earth would flow 1588 km. The effusion rates for a terrestrial flow are lower, at 49,482 m³/s. Starting flow velocities are 9.9 m/s. Since terrestrial lavas have higher vesicularity than the Venusian flow (as expected owing to less atmospheric pressure), vesicularity of 30% was tested under terrestrial conditions. Such a flow would have an effusion rate of 38,393 m³/s, starting velocity of 7.68 m/s, and would travel 982 km. A Venusian tholeiite flow, 25 m deep, would travel approximately 51.5% farther than a 22% less dense terrestrial lava.

At shallower flow depths of 17 m (consistent with depths projected by Kezthelyi and Self, 1998 for generating rapidly emplaced basalt flows > 1000 km long on Earth), and 10% vesicularity, Venusian flows will travel 515 km. Starting velocities are lower than suggested by Kezthelyi and Self (1998) at 5.15 m/s, likely due to the low inclination of Venusian plains. The effusion rate is 17,510 m³/s, which is almost triple the Iceland Laki eruptions, but well within the range of terrestrial flood basalts. This effusion rate is slightly higher than those projected by Kezthelyi and Self (1998), likely owing to the channel starting width of 200 m used across all simulations. If we apply a flow depth of 17 m to an terrestrial lava flow (again at 30% vesicularity), the flow travels 317 km, which is less than predicted by Kezthelyi and Self (1998). This flow has a starting velocity at 3.53 m/s, and effusion rates of 12,010 m³/s. These starting velocities are lower than described in Kezthelyi and Self (1998), which we ascribed to the low slope angle of the Venusian plains applied in our modeling. In order to quickly test the effects of slope on initial flow velocity, we increased the slope from 0.159 to 0.5° for the terrestrial flow. The
initial velocity increased considerably from 3.53 m/s to 11.23 m/s. The maximum crystallinity in the Venusian flow is 41.4% with a viscosity of 9.12 x10³ Pa·s. The minimum core temperature attained at the end of the Venusian flow is 1430.7 K.

If higher vesicularity is used, flow length decreases further as expected. A density of 2430 kg/m³ at a flow depth of 17 m produces flows with starting velocities of 4.63 m/s and effusion rates of 15,735 m³/s (still over double the historically observed Laki eruption, but within the range of ancient terrestrial flood basalts). This flow would reach 416 km in length. The maximum crystallinity in a Venusian flow is 40.8% and a viscosity of 8.55 x10³ Pa·s. The minimum core temperature attained at the end of the Venusian flow is 1431.6 K.

One of the first tests aimed to replicate Laki-eruption scale effusion rates for lava flows on both Venus and Earth. For this trial we again used 30% vesicularity for Earth and 10% for Venus (2100 and 2700 kg/m³, respectively). Unlike other trials, the depths were set to attain effusion rates as close to 6600 m³/s as possible (that estimated for the Laki eruption, Thodarson, 2003). The depths were set at 14 m for Earth (resulting in constant effusion of 6685 m³/s), and 12.33 m for Venus (effusion of 6646 m³/s). The results are a 178 km flow on Earth (initial starting velocity of 2.39 m/s), and a 198 km flow on Venus (initial starting velocity of 2.69 m/s) (Figure 6a). The maximum crystallinity in the Venusian flow is 39.5% and a viscosity of 7.31 x 10³ Pa·s. The minimum core temperature attained at the end of the Venusian flow is 1434 K.

If we run the model with extreme depths of 50 m (approximately the average depth of Baltis Valis, assuming it was completely filled with lava), with 10% vesiculation and tholeiite bulk densities of 2700 kg/m³, the tholeiite flows run 7,560 km on Venus. The initial velocities are 44.96 m/s, with effusion rates of 449,562 m³/s. Final core temperature is 1423 K, viscosity is 1.69 x 10⁴ and crystallinity is 46.9%. The effusion rates for this trial are likely to be unrealistically high, as this value is approximately 12 times the estimated effusion rates of the Deccan Trap flows.
### Table 4: Results from tholeiite basalt flow trials

<table>
<thead>
<tr>
<th>Planet</th>
<th>Flow Depth (m)</th>
<th>Density (kg/m³)</th>
<th>Fluid Viscosity (Pa·s)</th>
<th>Effusion Rate (m³/s)</th>
<th>Flow Length (km)</th>
<th>Flow Time (Hours)</th>
<th>Initial Velocity (m/s)</th>
<th>Final Tcore (K)</th>
<th>Final Viscosity (Pa·s)</th>
<th>Final Crystallinity (%)</th>
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</thead>
<tbody>
<tr>
<td>Venus</td>
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<td>2700</td>
<td>800</td>
<td>449,562</td>
<td>7,560</td>
<td>149</td>
<td>45</td>
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<td>3000</td>
<td>800</td>
<td>62,206</td>
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<tr>
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<td>800</td>
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<td>1427.3</td>
<td>1.19E+04</td>
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</tr>
<tr>
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<td>800</td>
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<tr>
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<td>1000</td>
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<td>2100</td>
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<td>2.69</td>
<td>1434</td>
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</table>
Figure 6: Tholeiite flow velocities and travel distances under terrestrial and Venusian conditions using $p = 3000$ kg/m$^3$ for tholeiite DRE $d =$ depth (m), $p =$ density (kg/m$^3$) a) at effusion rates comparable to the Laki eruption b) for various flow depths and bulk densities.
Figure 7: Core temperature cooling of modeled Venusian tholeiite lavas through time, $d =$ flow depth (m), $p =$ density (kg/m$^3$).

Figure 8: Rheological changes in viscosity of modeled Venusian tholeiite lavas through time, $d =$ flow depth (m), $p =$ density (kg/m$^3$).
Figure 9: Crystallization in modeled Venusian tholeiite lava through time for varying depths and bulk densities, \( d \) = flow depth (m); \( p \) = bulk density (kg/m\(^3\)).
3.3 Komatiite Flows

Komatiites have a wider range of viscosities than tholeiite from 0.1-10 Pa·s (Huppert and Sparks, 1985). At the lower range, the difficulties experienced modelling carbonatites (unreasonably high velocities and effusion rates) also applied to the komatiite flows. Because of their lower viscosities, komatiites flow at significantly higher velocities than tholeiite lavas, even at shallow flow depths and low slope angles. As a result, the FLOWGO model projects very high (Deccan Traps scale) effusion rates for our few trials.

Our first trial explored the upper viscosity limits (10 Pa·s), which are approximately one order of magnitude smaller than tholeiite basalt, and more suited for the FLOWGO model than more fluid komatiite lavas. With an eruptive temperature of 1873 K and at flow depths of 5 m, a Venusian flow would have an effusion rate of 38,130 m$^3$/s, starting velocity of 38.13 m/s, and would be expected to flow 390 km. Final core temperatures reach 1780 K, viscosity of 42.1 Pa·s, and crystallinity of 31.3%. A terrestrial flow of the same depth and composition would require an effusion rate of 32,523 m/s$^3$, would have a starting velocity of 32.5 m/s, and is expected to flow 328 km (Fig. 10a). Even though the core temperatures have not reached the komatiite solidus, flow velocity reaches zero yield strength increases, momentum diminishes, and gravitational pull is no longer able to move the lava on such a flat slope.

At moderate komatiite viscosities (1 Pa·s), for the same flow depth of 5 m, velocities and effusion rates increased by approximately an order of magnitude. For the Venusian trial, the initial velocity is 381 m/s, with effusion rates of 381,301 m/s$^3$. This flow is modelled to travel a distance of 1131 km. Final core temperatures reach 1780 K, viscosity of 4.21 Pa·s, and crystallinity of 31.3%. A terrestrial flow of the same composition would be expected to travel 894 km with initial starting velocities of 325.2 m/s and effusion rates of 325,227 m$^3$/s (Fig. 10b). The initial velocities of the 1 Pa·s at 5 m flow depth is unreasonably high. If flow depth is decreased to 2 m, then flow length reaches 117 km, and starting velocities and effusion rates are substantially lower, at 57 m/s and 22,744 m$^3$/s respectively (Fig. 11). Final core temperature, viscosity, and crystallinity are 1814 K, 2.74 Pa·s, and 23.8%, respectively (Fig. 12).
Table 5: Results from komatiite flow trials

<table>
<thead>
<tr>
<th>Planet</th>
<th>Flow Depth (m)</th>
<th>Density (kg/m³)</th>
<th>Fluid Viscosity</th>
<th>Effusion Rate (m³/s)</th>
<th>Flow Length (km)</th>
<th>Flow Time (Hours)</th>
<th>Initial Velocity (m/s)</th>
<th>Final Tcore (K)</th>
<th>Final Viscosity (Pa-s)</th>
<th>Final Crystallinity (%)</th>
</tr>
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<tbody>
<tr>
<td>Venus</td>
<td>5</td>
<td>2880</td>
<td>10</td>
<td>38,130</td>
<td>390</td>
<td>7.6</td>
<td>38.1</td>
<td>1780</td>
<td>42.1</td>
<td>31.3</td>
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<tr>
<td>*Venus</td>
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<td>2880</td>
<td>1</td>
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<td>1131</td>
<td>3.3</td>
<td>381</td>
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</tr>
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<td>2.7</td>
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<td>1822</td>
<td>2.48</td>
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<tr>
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<td>894</td>
<td>2.2</td>
<td>325.2</td>
<td>1786</td>
<td>3.88</td>
<td>30.1</td>
</tr>
</tbody>
</table>

*Flow trials that are discounted due to unreasonably high initial flow velocities
Figure 10: Komatiite flow velocities over flow length for komatiites on Earth and Venus, with a depth of 5 m, viscosities of a) 10 Pa·s. and b) 1 Pa·s.
Figure 11: Comparing effects of depth and viscosity on Venusian Komatiite flows through time, $d = \text{depth (m)}$, $n = \text{viscosity (Pa}\cdot\text{s)}$.

Figure 12: Comparing effects of depth and viscosity on Venusian komatiite flows over time, $d = \text{depth (m)}$, $n = \text{viscosity (Pa}\cdot\text{s)}$. 

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Discussion

4.1 Variations in FLOWGO Input Parameters

Modelling with the initial channel dimensions from Oshigami and Namiki (2007) yielded improbable results. If we assume that Baltis Vallis was completely filled with lava during peak flow, the model produces flow velocities peaking above 100 m/s, with effusion rates variable, but often >1,000,000 m$^3$/s. If we instead suppose that the canali are filled just above the average depth of Baltis Vallis at 50 m, and an initial flow width of only 200 m, then the starting flow velocity is ~45 m/s. Whereas this is not an entirely unreasonable flow velocity, it is still accompanied by excessively high effusion rates of 449,562 m$^3$/s. It is therefore unlikely that Venusian canali were ever filled to the brim with fluid, and were likely carved through a lower volume input, either sustained through a long duration, or through multiple pulses, resulting in channels which resemble terrestrial canyons. Sustained erosion can also account for the great width of the canali (~2 km average for Baltis Vallis) allowing for lower effusion rates than if the entire width was filled at any given time. Our tests always considered an initial channel width of 200 m, as opposed to 2000 m, with channel width increasing as velocity decreases over the course of the flow duration.

The force of gravity plays an important role in determining lava flow velocity, and ultimately how far lava will flow before crystallizing. However, the density of the Venusian atmosphere, and subsequent suppression of volatile exsolution and vesiculation, seems to offset Venus’ lower gravitational acceleration, enabling lavas to flow at starting velocities higher than terrestrial equivalents. Before accounting for vesiculation, terrestrial rhyolite, basalt, and komatiite flows for a specific depth had higher starting velocities than the Venusian flows, and traveled greater distances. The latent heat of crystallization for the different lavas also had strong control on flow length.
4.2 Venus vs. Earth Tholeiite Flows

The modelled tholeiite flows show that identical tholeiite lava compositions will flow greater distances under Venusian conditions than under terrestrial conditions, after bulk density is adjusted for vesiculation. The conditions required for terrestrial flows to travel over 1500 km are unreasonable, and not expected to ever occur on Earth. These results require the total absence of, or very low exsolution of volatiles creating pore spaces and reducing bulk density. Additionally, the effusion rates required to achieve these flow lengths are approximately 1.5 times more than those estimated for the Deccan Traps, which was the largest known volcanic event in Earth’s history. In comparison the longest flows on Earth, the Rajahmundry Trap lava flows, travelled ~1000 km at Deccan Traps effusion rates (~37,000 m³/s). This length and effusion rate fits well with our modelled terrestrial flow of similar parameters (982 km, 25 m deep, Q = 38,393 m³/s) Whereas the Rajahmundry Trap lava flows have recorded lobe heights of 25 m, this is a maximum depth, and not the average across the length of the flow.

It is worth noting that the present minimum depth of Baltis Vallis as reported by Oshigami and Namiki (2007) is 17.3 m. Although their average reported depth is 46 m, with 114 out of 119 measured points having depths >25 m, this minimum depth shows that a lava flow in Baltis Vallis could not have been greater than 17.3 m at this interval, at present depths, lest it spill over the sides of the channel to form levees. Constructional levees are present along Baltis Vallis, particularly in the quarter of the channel proximal to the source area. It is also possible that channels were deeper prior to flow emplacement, and have partially filled with cooled lava. However, a depth restriction shows that our postulated parameters and conditions for achieving flow lengths >1000 km on Venus may not fit the empirical evidence of what such flow channels actually look like.

The high effusion rates required to make a lava flow travel the reported distances make such flows highly improbable on Earth, but not necessarily on Venus. As a planet, Venus is characterized by widespread volcanism, and it is suggested that Venus underwent a global resurfacing event 200-700 Ma (Phillips and Hansen, 1998; Smrekar et al., 2016). Analysis of canali meander wavelengths has suggested that the canali require peak discharges up to $6.6 \times 10^6$ m³/s which is an order of magnitude greater than
terrestrial rivers, and greater than our highest modeled effusion rate for basalt (Bray et al., 2007). Accomplishing effusion rates that high from basalt would require a massive eruptive event. Previous studies of Venustian canali have even claimed that the scale of lava eruption to produce flows 100s to 1000s of km long can only be explained by planetary-scale volcanism (Komatsu, 1993). The scale of such a resurfacing event extends beyond all terrestrial analogues, and with such a staggering precedent it is not unreasonable to consider extreme effusion rates for Venustian lava flows. Quicker flow rates at lower volumes of a less viscous fluid, like a carbonatite, might be more capable of more regularly reaching the effusion rates described by Bray et al. (2007) to achieve canali meander geometries.

Our trials to match flow lengths >1000 km as predicted by Kezthelyi and Self (1998) for “rapidly emplaced lavas” were hampered by the gentle slopes of Venustian plains. We used their maximum suggested flow depth of 17 m, but the resulting initial velocities were lower (5.15 m/s on Venus, 3.53 for lower-density Earth lava) than their projected flows, which had predicted velocities of 4-12 m/s, and lower effusion rates of 3100 to 11000 m$^3$/s. Our higher effusion rates can be attributed to our initial channel width of 200 m. The flow distance achieved with these parameters was 515 km under Venustian conditions and 317 km under terrestrial conditions: over half of the predicted distance of 1000 km. An increase in slope from 0.159° to 0.5° changes the terrestrial starting velocity from 3.53 m/s to 11.23 m/s, showing the extreme sensitivity to slope in the FLOWGO model.

The majority of Venustian canali are significantly shorter than Baltis Vallis, although most are longer than 500 km (Komatsu and Baker, 1994). Our modelling has shown that at Laki eruption-scale effusion rates, Venustian tholeiite flows are capable of travelling ~200 km. This distance increases as effusion rates and flow depths increase for a channel of a given starting width, with loosely defined “reasonable” effusion rates traveling over 500 km. We were able to achieve the distance of Baltis Vallis using flow depths of 50 m. In contrast, modeling done by Komatsu et al. (1992) using tholeiite lava achieved flow distances of 6800 km (the previously measured length of Baltis Vallis before new measurements of 7181 km) but at depths of ~ 80 m, which is significantly deeper than the 46 m average depth of Baltis Vallis. Even though the shallowest segment of Baltis Vallis is measured at 17.3 m, the results we attain using flow depths of
25 m serve as a reasonable average flow depth, because very few measured points are shallower than this (Oshigami and Namiki, 2007). It is possible then, that basaltic eruptions can produce flows of similar length as the canali, but Baltis Vallis represents either a different erosion mechanism, or a massive planetary-scale eruptive event.

### 4.3 Komatiite Flows

The komatiite flow trials were affected by the same issues concerning the carbonatite trials with regards to flow viscosity, velocity, and effusion rates. We tested the maximum viscosity of komatiite lavas (10 Pa·s) as well as an intermediate value (1 Pa·s). At 5 m depth, the 1 Pa·s komatiites gave both unrealistic velocities and effusion rates; an order of magnitude higher than those of the 10 Pa·s lava, owing to the direct proportional causal effect viscosity on velocity and effusion rates in the FLOWGO set up. Because the flows modeled at 5 m and 1 Pa·s show effusion rates in range of 10 times the Deccan Traps, we question both the plausibility of these results, as well as the effectiveness of the velocity equation for modelling lower viscosity fluids.

At shallower depths of 2 m, the flows at both 10 and 1 Pa·s appear far more plausible. At 2 m depth, 1 Pa·s flows have effusion rates less than the Deccan Traps, and the 10 Pa·s komatiites have effusion rates less than Laki, and comparable to modern volcanic eruptions. In contrast, these flows do not travel nearly as far as their basaltic counterparts. The 5 m deep, 10 Pa·s trial was the farthest flowing komatiite within reasonable effusion rates and velocities that we modelled, and travelled ~400 km. Despite higher initial starting velocities, the 2 m, 1 Pa·s trial travelled a shorter distance, which can be attributed to quicker cooling. The shallower lava has less insulation, and core temperatures dropped quicker than the thicker flow trials. It is unlikely that any komatiite flow, under these conditions, could flow long enough distances to erode channels like Baltis Vallis, at 1000s of km.

Another observation regarding the komatiite flow trials is that their velocities reach zero before the lava core temperatures reach komatiite solidus temperature. This can be explained by the incremental increases in crystallinity, which would be occurring at the edges of the flow, and not the core. Increases in crystallinity increase flow
viscosity, and internal yield strength. Velocity is approaching zero when the gravitational pull on the lava down the slope is no longer able to overcome the increased yield strength of the cooling flow. This is occurring more quickly than expected because of the almost negligible slope of the Venusian plains. The forces pulling the lava are less than observed on steeper slopes.

4.4 Erosive Capabilities of Laminar vs. Turbulent Flows

Although we have shown that basaltic lava can flow the lengths of the canali, it remains unresolved if basalt is capable of forming fluvial-like channels. One of the prevailing arguments against basaltic lavas as the erosive mechanism for Venusian canali is the viscosity of the lava. Tholeiite classically flows in a laminar fashion, which does not explain the meander geometry of the Venusian canali, assuming flat and planar topography. Cut-banks in terrestrial rivers are eroded via the helical nature of river flows. Water at the air-surface interface of a river experiences less frictional resistance, and moves at a higher velocity than the centre of the flow. When the water encounters a bend in the river, the higher inertia of the surface water drives it down the bank in a corkscrew motion, eroding into the cut-bank and across the bottom of the river channel. This water slows down as it moves up the opposite point bar, and deposits sediment load at these lower velocities (Dalrymple et al., 2010). Venusian canali have bends and meanders, and even oxbow cut-offs, showing clear evidence of channel migration (Williams-Jones et al., 1998). Even though we have shown that tholeiite flows can travel the 100s to 1000s km lengths of the canali, we have not proven that they are capable of achieving helical flow to mechanically eroding meandering landforms, to the point of forming oxbows.

Reynold’s numbers are used to describe laminar and turbulent flow behaviour for Newtonian fluids. Although basalt is not a Newtonian fluid, we calculate the Re values for our thicker lava flows to see if the tholeiite could potentially be showing turbulent behaviour. The Re number for our Venusian tholeiite flow 25 m thick, at 2700 kg/m$^3$ is 614 at the start of its flow, whereas our flow 17 m thick at 2700 kg/m$^3$ would have a Reynold’s number of 192 (Eq. 1).
\[ Re = \frac{\rho dv}{\eta} \]  
(Eq. 1)

Komatsu et al., (1992) state that the laminar-turbulent flow boundary for basalt is \( Re = 500 \), whereas Carr et al., explain that the turbulent-laminar transitional boundary is between 500-2000, with turbulent flow being \( Re > 1000 \). Our 25 m thick flow surpasses the lower limit of the laminar-turbulent flow boundary, and may be capable of flowing turbulently, or showing some turbulent character. FLOWGO assumes laminar flow for its cooling calculations, and we incorporated a fraction of crustal coverage into our modeling, which would not be maintained under turbulent flow conditions. Therefore, while our thicker flows might be capable of flowing turbulently, they would not travel as far as we predicted, losing heat more rapidly than their laminar counterparts. Additionally, most canali are considerably shorter than Baltis Vallis, but also show meandering patterns. Even if the thicker flows could show turbulent flow characteristics, the shallower flows that form shorter canali would not be showing turbulent behaviour.

Komatiite lavas are considerably less viscous, behaving much closer to water and would be more inclined to flow turbulently, although again this was not taken into consideration during the modelling stage. It is mechanically possible for a komatiite flow to be erode a meandering channel, but our modelling shows that komatiites cool too quickly to travel the distances for Venusian canali, even if flows are laminar only. Carbonatite is known to flow turbulently, and remains a strong candidate for fluids capable of mechanically eroding meanders and fluvial-like channels.
Conclusions

Only our 50 m tholeiite flow under Venusian ambient conditions reached the >7000 km required to match the extent of Baltis Vallis. Rhyolite is far too viscous to travel the distances required, and the thickness used (150 m) to get the rhyolite to flow at all on the Venusian plains is far deeper than the average depths of Baltis Vallis. Under Venusian conditions, flows for any set chemical composition are able to flow greater distances because of higher bulk densities due to lower vesiculation. Even though no trial with “reasonable” effusion rates ran over 7000 km, we were able to generate tholeiitic lava flows capable of traveling 1806 km assuming zero vesiculation. If 10% vesicularity is considered, then Venusian tholeiite can flow 1488 km at plausible effusion rates and velocities. Both of these distances are significant, and well beyond the length of most Venusian canali, which are only a few 100s of km in length. Lower effusion rates in the range of 6,600-30,000m³/s may be capable of flowing the lengths of shorter canali <1000 km. Although basaltic lava may be able to flow the lengths of Venusian canali, it is uncertain if tholeiitic flows will regularly flow turbulently to carve meanders into channel cutbanks. If our 25 m thick flow is capable of flowing turbulently, and would lose heat more rapidly than modeled owing to absence of crustal coverage, and would not flow as far as predicted. Turbulent flow behaviour would also affect the viscosity of the lava.

Testing lower viscosity lavas did not function properly in the FLOWGO model, as evidenced by the unrealistically high starting velocities. The results we attained for komatiite flows at 10 and 1 Pa·s showed that shallow komatiite flows cool too quickly to travel the long distances of the canali, travelling less than 400 km each. Further modelling is necessary to determine if extremely low viscosity komatiites or carbonatities are capable of flowing equal or greater distances than tholeiitic lavas. Based on this work, carbonatities are still the most likely fluid capable of eroding the Venusian canali based on their low viscosity, enabling water-like behaviour and capacity for turbulent flow, along with their crystallization temperature being near the ambient surface temperature of the Venusian.
5.1 Future Modelling Considerations

Something not considered for komatiite flows is the contamination from underlying substrate from thermal erosion. Because komatiites have higher extrusion temperatures than the liquidus of basalt, komatiite flows would partially melt the underlying basaltic substrate. Silicic melt would be incorporated into the komatiite lava, contributing to the increase in viscosity of the flow over time. Basaltic contamination could be as high as 10% (Huppert and Sparks, 1985).

Additionally, the FLOWGO program models channel width as increasing down flow, as lava lobes fan out. This is not consistent with observations of Venusian canali, which have fairly constant widths, and actually taper slightly down channel (Bray et al., 2007).

Finally, the FLOWGO model calculates heat loss assuming laminar flow conditions. Because our 25 m flow may be showing some turbulence, it is likely to have no crustal coverage, and would lose heat more rapidly than we initially calculated. Future studies should take turbulent flow conditions into consideration when calculating heat loss.
References


