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Mantle potential temperature estimates of basalt from the surface of Venus

J. Gregory Shellnutt*

National Taiwan Normal University, Department of Earth Sciences, 88 Tingzhou Road Section 4, Taipei 11677, Taiwan

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ABSTRACT

The crater density and distribution of Venus indicates the average surface age is younger (≤1 Ga) than most terrestrial planets and satellites in the Solar System. The type and rate (i.e. equilibrium, catastrophic or differential) of volcanism associated with the stagnant lid tectonic system of Venus is a first order problem that has yet to be resolved but is directly related to the thermal conditions of the mantle. The calculated primary melt composition of basalt at the Venera 14 landing site is high-Mg basalt to picrite with a mantle potential temperature estimate similar to terrestrial ambient mantle (1370 ± 70 °C). The calculated accumulated fractional melting curves indicate the olivine compositions from the melt have Mg# of 89–91. The results show that the thermal regime required to generate the primary melt composition of the Venera 14 basalt was not anomalously high (i.e. mantle-plume system) but rather consistent with a lithospheric tensional rift system. The juxtaposition of high thermal regime structures (e.g. Beta Regio) and 'ambient' mantle potential temperature estimates of the Venera 14 basalt suggests that the relatively young surface of Venus is the result of volcanism from a combination of thermal systems that resurfaced the planet at variable rates.

1. Introduction

The surface of Venus is dominated (70–80%) by featureless lava plains that lie within ±1 km of the mean planetary radius (Fegly., 2003; Hansen and Young, 2007). The rest of the surface comprises median elevation (1–2 km) mesolands and highland (>3 km) regions. The mesolands have features such as coronae and chasmata (troughs) whereas the highlands consist of crustal plateaux, tesserae terrane, volcanic edifices and regional-scale compression-related mountains (Suppe and Connors, 1992; Nimmo and McKenzie, 1998; Phillips and Hansen, 1998; Hansen and Young, 2007). Unlike the Moon, Mercury and the southern hemisphere of Mars, the surface of Venus is not heavily cratered suggesting there is a dynamic process of planetary-wide resurfacing in the absence of Earth-like plate tectonics (Nimmo and Mckenzie, 1998; Smrekar et al., 2007). The impact crater density and distribution on Venus indicate the mean surface age is ≤1 Ga (Schaber et al., 1992; Strom et al., 1994; Basilevsky and Head, 2002; Hansen and Young, 2007). In order to explain the relatively young surface of Venus, two end-member models are proposed: (1) a catastrophic resurfacing within 100 Ma that was related to high rates of volcanism and a hot mantle (i.e. mantle plumes and/or insulated mantle), or (2) a more gradual resurfacing over a period of 500 Ma related to equilibrium rates of volcanism associated with ambient mantle temperatures and lithospheric rifting (Schaber et al., 1992; Phillips et al., 1992; Strom et al., 1994; Lancaster et al., 1995; Hauck et al., 1998; Reese et al., 2007; Romeo and Turcotte, 2010; Smrekar et al., 2010; O'Rourke et al., 2014; Kreslavsky et al., 2015). A third model suggests differential resurfacing where periods of elevated volcanism occurred independently of or concurrently with periods of equilibrium volcanism (Hansen and Young, 2007; Reese et al., 2007; O'Rourke et al., 2014; Ivanov and Head, 2015). An advantage of the differential resurfacing model is that it incorporates the spatial pattern of volcanically embayed craters rather than using a random volcanic distribution that was used for the equilibrium and catastrophic resurfacing models (Ivanov and Head, 2015).

The morphology and structure of lava flows on Venus are variable and testify to different lava viscosities, thermal regimes and possibly different compositions (Head et al., 1992; Pavri et al., 1992; Lancaster et al., 1995; Nimmo and Mckenzie, 1998; Byrnes and Crown, 2002; Shellnutt, 2013). The prevalence of pahoehoe with some a'a basalt flows suggests volcanic effusion rates were similar to Earth, but there is evidence of lower viscosity flows that may have ultramafic compositions (Kargel et al., 1993; Lancaster et al., 1995). The identification of giant radiating dyke swarms, large igneous provinces and large volcanic edifices suggest that mantle-plume-related eruptions were common on Venus and contributed significantly to planetary cooling and resurfacing

* Tel.: +886 277346386.
E-mail address: jgshelln@ntnu.edu.tw, gshellnutt@hotmail.com

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Table 1
Surface compositions of Venus and estimates on primary melt compositions and mantle potential temperatures.

Sample	Venera 13[a]	Venera 14[a]	Vega 2[a]	N-MORB[a]	Venera 14 (model 1)	Batch melt	AFM	Venera 14 (model 2)	Batch melt	AFM	AFM	Venera 14 (model 3)	Batch melt	AFM
SiO_2 (wt%)	45.1 ± 3.0	48.7 ± 3.6	45.6 ± 3.2	48.77	48.7	49.11	49.13	48.7	46.41	46.47	46.66	48.7	48.06	48.13
TiO_2	1.59 ± 0.45	1.25 ± 0.41	0.2 ± 0.1	1.15	1.25	1.26	1.26	1.25	1.07	1.08	1.12	1.25	1.19	1.20
Al_2O_3	15.8 ± 3.0	17.9 ± 2.6	16 ± 1.8	15.9	17.9	17.97	18.01	17.9	15.26	15.44	15.99	17.9	17.05	17.22
Fe_2O_3						0.63	0.63		0.53	0.54	1.12		0.60	0.60
FeO						6.62	6.61		9.0	9.0	8.44		7.57	7.56
FeOt	9.3 ± 2.2	8.8 ± 1.8	7.7 ± 1.1	9.82	7.1			9.7				8.1		
MnO	0.2 ± 0.1	0.16 ± 0.08	0.14 ± 0.12	0.17	0.16	0.16	0.16	0.16	0.15	0.15	0.16	0.16	0.16	0.16
MgO	11.4 ± 6.2	8.1 ± 3.3	11.5 ± 3.7	9.67	9.9	10.21	10.11	11.3	15.56	15.14	13.91	11.1	12.61	12.24
CaO	7.1 ± 0.96	10.3 ± 1.2	7.5 ± 0.7	11.16	11.4	11.44	11.47	11.5	9.82	9.94	10.29	10.8	10.29	10.39
Na_2O	2.0 ± 0.5[b]	2.4 ± 0.4[b]	2.0[b]	2.43	2.4	2.41	2.42	2.4	2.04	2.07	2.14	2.4	2.29	2.31
K_2O	4.0 ± 0.63	0.2 ± 0.07	0.1 ± 0.08	0.08	0.2	0.20	0.20	0.20	0.17	0.17	0.18	0.20	0.19	0.19
SO_3	1.62 ± 1.0	0.88 ± 0.77	4.7 ± 1.5											
Cl	<0.3	<0.4	<0.3											
Total	98.1	98.7	99.89	99.15	99.01	100	100	103.11	100	100	100	100.61	100	100
Pressure (bar)						100	100		100	100	100		100	100
FeO (source)						8.02	8.02		8.02	8.02	8.02		8.02	8.02
MgO (source)						38.12	38.12		38.12	38.12	38.12		38.12	38.12
Fe_2O_3/TiO_2						0.5	0.5		0.5	0.5	1.0		0.5	0.5
Mole fraction Fe^{2+}/Fe^*						0.92	0.92		0.94	0.94	0.88		0.93	0.93
% ol addition						0.6	0.3		13.0	11.8	8.2		4.3	3.3
F (%)						0.06	0.06		0.11	0.11	0.09		0.11	0.11
Temperature (°C)						1240	1240		1360	1350	1330		1300	1290
T_P (°C)						1310	1310		1450	1440	1410		1370	1360

Error of the Venera 13, Venera 14 and Vega 2 data reported at the 1σ level. AFM = accumulated fractional melt. F = melt fraction. The model compositions above are entered in PRIMELT3. The software automatically normalizes the data to 100% for the calculation.
[a] Venera 13, Venera 14, Vega 2 and normal mid-ocean ridge basalt (N-MORB) (Surkov et al., 1984, 1986; Fegly, 2003).
[b] The Na_2O content is calculated for the Venera 13, 14 and Vega 2 data (Fegly, 2003).
[*] Denotes total Fe ($Fe^{2+} + Fe^{3+}$).

(Ernst et al., 1995; Hansen, 2007; Basilevsky and Head, 2007; Smrekar et al., 2010). Rapid eruption rates of mafic and ultramafic lavas and high (>1550 °C) mantle potential temperatures (T_P) are characteristic of terrestrial mantle-plume-derived mafic large igneous provinces (LIPs) (Campbell, 2007; Bryan and Ernst, 2008; Ali et al., 2010). Consequently the eruption durations of terrestrial LIPs are typically ≤10 million years, in some cases <1 million years, and may be analogous to volcanic features and mantle conditions that existed on Venus during a catastrophic resurfacing event.

The surface composition of Venus was analyzed by the Venera 13, Venera 14 and Vega 2 landing probes (Surkov et al., 1984,1986; Kargel et al., 1993; Fegly, 2003). The compositions of the rocks are basaltic but the Venera 13 site appears to be more alkaline than the compositions found at Venera 14 and Vega 2 (Table 1). In comparison to terrestrial samples, the rocks at the Venera 14 and Vega 2 sites are similar to normal-mid-ocean ridge basalt (N-MORB) and thus may have originated from a thermal regime similar to spreading centers on Earth (Mckenzie et al., 1992; Kargel et al., 1993; Fegly, 2003).

The primary melt composition of basalt can be deduced from its bulk composition providing it has only experienced olivine loss (Herzberg et al., 2007; Herzberg and Asimow, 2008, 2015). Based on forward modeling of dry peridotite, the primary melt composition can constrain the mantle potential temperature (T_P) required to produce the melt and thus provide the initial melting pressure because the adiabatic temperature-pressure melting path is very similar to the olivine liquidus (Herzberg et al., 2007; Herzberg and Asimow, 2008, 2015). Ambient mantle potential temperatures on Earth at mid-ocean ridge settings range from 1300 °C to 1400 °C whereas some ocean island basalts and continental flood basalts record mantle potential temperatures ~200 °C to ~300 °C above ambient temperatures (Farnetani and Richards, 1994; Campbell, 2007; Herzberg and Asimow, 2008; Ali et al., 2010). Higher mantle potential temperatures (>1550 °C) at some non-ridge within-plate settings on Earth are considered to be an artifact of a mantle-plume (Campbell, 2007; Herzberg et al., 2007; Herzberg and Asimow, 2008, 2015; Ali et al., 2010). Thus constraining the T_P of basalt can help to determine the thermal regime of its formation and provide support in favor of a passive rift setting or a mantle-plume setting. In this paper, the primary melt composition and mantle potential temperatures of venusian basalt are calculated in order to assess the likely thermal regime required to produce the melt and to determine if the conditions favor a particular resurfacing model (i.e. equilibrium, catastrophic or differential).

2. PRIMELT3 calculations

The primary melt and mantle potential temperature are calculated using PRIMELT3 (Herzberg and Asimow, 2015). The calculations used in PRIMELT3 are based on experimental results of terrestrial mantle compositions. The estimated mantle composition of Venus suggests it is likely to be analogous with terrestrial mantle but it is possible that the bulk FeOt may differ from Earth by a few percent (Fegly, 2003; Smrekar et al., 2007; Treiman, 2009; Filiberto, 2014; Rubie et al., 2015). The most important parameters for the calculation of the primary melt and T_P values are FeOt and MgO because they are the constituent components of olivine. The variability of CaO is also important because it is an indicator of clinopyroxene and/or plagioclase removal from the liquid. Therefore the amount of CaO must be flexible such that the model composition does not produce a clinopyroxene (i.e. augite) fractionation warning. The extreme values of MgO, FeOt and CaO ($\pm 1\sigma$ error) were sought so that the maximum and minimum T_P can be determined. The remaining elements (i.e. TiO_2, Al_2O_3, MnO, CaO, Na_2O and K_2O) are not major components of olivine and thus their variability will not significantly influence the estimates. The poor precision of the venusian basalt data are a hindrance to the certainty of the calculations as they do not allow for a single unique result (Treiman, 2007). However the end-member melt compositions and T_P will constrain the maximum range that is likely given the uncertainty of the available data.

The peridotite source model uses FeO and MgO contents equal to terrestrial values (i.e. FeOt = 8.02 wt%; MgO = 38.12 wt%) and the pressure equal to the surface of Venus (i.e. ~100 bar). The