Processes Forming Volcanic Topography at Atla Regio, Venus

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Summary: Venus and Earth share a similar size, mass and density, however Venus' high surface temperatures, pressures, dehydrated crust and absence of subduction zones, makes the surface unique. One area which shares similar volcanic landforms to the Earth's is the equatorial highland Atla Regio, which extends from 170°E-220°E 30°N-10°N. The similarity of these volcanic features to Earth based examples, as well as an understanding of the processes which produced them, can help to constrain the deformational mechanisms that may have operated at Atla Regio. This study will aim to compare volcanic topography at Atla Regio to topography observed within the Hawaiian and Cape Verde Archipelagos, Earth to determine if similar deformational processes occurred within Atla Regio. On Earth, three main processes act to produce volcanic topography and include: 1) uplift associated with a mantle plume impinging on the underside of the lithosphere; 2) rifting and volcanism associated with the mantle plume and; 3) volcanic loading, where the extra volcanic mass is compensated by flexure of the lithosphere. An understanding of the processes that gave rise to topography at Atla Regio is fundamental to determine a suitable rheological profile within this area.

Keywords: Atla Regio, Venus, Hawaii, Cape Verde, mantle plume, volcanism, lithospheric flexure

Introduction

Venus consists of three main types of geological landforms, including highland areas, lowland volcanic plains and lowland plains regions, which encompass 8%, 27% and 65% of Venus' surface, respectively [1]. One highland located along the equator is Atla Regio, which is marked by three large volcanoes, Ozza Mons in the south-east, Maat Mons in the south-west and Sapas Mons in the north-west, as well as three rifts, Ganis Chasma, Dali Chasma and Parga Chasma that form a triple junction at Ozza Mons (Figure 1) [2]. Due to the presence of these rifts and volcanoes, in conjunction with this region's broad domal morphology, gravity high, extensive lava flows, high elevation (~3 km) and absence of compressional features, Atla Regio is considered to be a volcanic rise [2]. Atla Regio shares similar characteristics to Earth's hot-spots, where volcanism is associated with mantle plumes [1, 2]. On the basis of the localised topography and gravity observed at Ozza Mons and Sapas Mons, each of these areas may be sites of mantle upwellings, implying Atla Regio may have two mantle plumes acting to produce the topography and gravity anomalies within this area [2].

Volcanic topography on Earth can result when an upwelling mantle plume impinges onto the base of the lithosphere causing volcanism and uplifted topography forming a volcanic rise [3] (Figure 2). Generally these areas also have large dimensions > 1000 km, suggesting that the topography is not solely supported by lithospheric strength, but by a deeper source such as a mantle plume [4]. Two currently active, Earth hot-spots are Hawaii, situated along the Hawaiian-Emperor Seamount chain and Cape Verde, located west of Senegal, Africa [5, 6]. Currently Hawaii and the Cape Verde Islands are located above mantle plumes, which cause
uplifted topography and a volcanic rise [5, 6, 7, 8].

Figure 1 A 3D map showing the topography at Atla Regio. This area is dominated by three large volcanoes; Ozza and Maat Mons to the south and Sapas Mons to the north. A triple junction rift, comprising Ganis, Parga and Dalia Chasma meet at Ozza Mons.

Figure 2 Schematic showing the three processes that can give rise to volcanic topography. Red line is an upwelling mantle plume, which can produce uplift, volcanism and the production of volcanic topography. The volcanoes produced can act as loads and cause the lithosphere to bend (green line).

Aside from uplift and volcanism, the presence of a mantle plume can be further constrained by an excess temperature, since mantle plumes may represent extra heat, which is superimposed on the background temperature associated with mantle convection [8, 9]. These temperatures rely on a stagnation distance ($r_s$), which is the distance that separates the upwelling plume material from the normal asthenosphere and a plume channel thickness ($A$), which is the diameter of the asthenospheric conduit that connects the plume from the core-
mantle boundary to the base of the lithosphere [8, 9]. Sleep [8] calculated an average excess temperature of 230-300 °C for stagnation distances of 350-450 km and a plume channel thickness of 100 km for the Hawaiian swell.

The origin and structure of mantle plumes on Venus is highly unconstrained with some authors proposing that plume channels may not extend throughout the entire depth of the mantle [10]. However, if a plume channel exists below Atla Regio it should be thicker (~200±100 km), than those estimated for Hawaii (~100 km±16 km), due to Venus' hotter mantle [8, 11, 12]. Also coronae are possible surface expressions of mantle plumes, therefore their range of diameters 200-600 km could represent a range of possible stagnation distances [13].

The presence of a mantle plume can also be associated with a buoyancy flux, where a thermal mass is upwelling to the lithosphere per second, by a plume [8, 9, 12]. Since plume buoyancy is proportional to heat flow, the excess heat flow associated with this plume can also be calculated [8, 9, 12].

For the Earth, the calculation of buoyancy flux relies on the knowledge of the absolute velocity of a hot-spot, which is defined as the motion of the hot-spot relative to the plate it resides on [8, 9]. Therefore, lower heat fluxes associated with the mantle plume component of heat flow are on the order of 10-20 mW m⁻² at slower-moving hot-spots, such as Cape Verde and Bermuda, compared to the higher mantle plume heat flux obtained at faster-moving hot-spots, such Hawaii [8, 9, 14].

Turcotte & Schubert [12] used a plate velocity of 90 mm yr⁻¹ for the Hawaiian hot-spot with a cross-sectional area of 1.13 km², to calculate a buoyancy flux of 7.4 Mg s⁻¹, which represents a mantle plume heat flux of 3x10¹¹ W or less than 1% of total global heat flow of the Earth at this location. Venus, however, has a stationary lithosphere and the closest Earth analog for these features may be for a slow-moving plate, since these areas may be closer to thermal equilibrium [14, 15]. Cape Verde is one example of a hot-spot located on a slow-moving plate [8, 14, 15]. The buoyancy flux, and in turn, the heat flow due to a mantle plume within this area is 1.6 Mg s⁻¹ and 20 mW m⁻² [8, 14]. Models by Smrekar & Parmentier [16] suggest that buoyancy fluxes, and in turn the extra heat fluxes from a mantle plume, which are most representative of Venus hot-spots, may not be large [14, 15].

Volcanoes associated with the impingement of a mantle plume can also initiate rifting and volcanism, contributing to the topography observed [17]. Hawaii is dominated by shield volcanoes, which are characterised by their broad shapes, low elevations and undulating slopes [17]. This morphology is also consistent with Venusesian volcanoes, however, these volcanoes generally have larger basal diameters and flatter profiles in comparison to those found at Hawaii [18]. The larger basal diameters of Venusesian volcanoes could be caused from an absence of plate motion, since a mantle plume would produce volcanism for longer time frames on a stationary lithosphere [1]. The lower elevations and shallower slopes of Venusesian volcanoes could result from: 1) high temperatures within the crust and at the surface that may have acted to slow cooling of rising magma; 2) low lava viscosities; 3) highly effusive, large volumes of magma; 4) extensive lava flow formation from lava tubes; 5) slow cooling of lava flows due to high atmospheric temperatures [18, 19, 20].

Modelling of gravity anomalies over Alta Regio can be useful to determine a suitable density distribution and therefore, geological structure below the highland [13]. The highland topography associated with volcanism at Alta Regio should correspond to a high free-air gravity anomaly, high geoid anomaly and a low bouguer anomaly [2]. Free-air gravity
anomalies at short-wavelengths correspond well to local topography [12]. Bouguer anomalies, however, tend to exhibit structure associated with long-wavelength, isostatically compensated topography [12]. At sites of large-scale topography a negative bouguer anomaly is usually observed since the excess mass of the highland is isostatically compensated by a low-density crustal “root” or a low-density partial melt and magma chamber associated with volcanism [12]. The geoid anomaly, defined as the elevation difference between a reference equipotential surface (geoid) and the measured geoid, also reflects deeper density anomalies [12, 21, 22].

The presence of these volcanoes on the surface can act as loads which force the lithosphere downwards at the center of the load, whilst the surrounding area bulges upwards (Figure 3) [12]. By modelling this flexural response the elastic lithosphere thickness, which is the portion of the lithosphere that is rigid enough to sustain elastic stresses for long periods, can be determined [12]. This elastic lithosphere thickness can be used to determine a suitable temperature at the base of this layer, and in turn a thermal gradient and heat flux through an area [23].

![Figure 3 An east-west profile showing the upwarping of the lithosphere associated with a volcanic load at Atla Regio. The distance between the center of the load (x) and the maximum amplitude of the forebulge (x_b) can be used to determine elastic lithosphere thickness (h). The red line is the inferred line of constant depth.](image)

Previous authors have determined a range of elastic lithosphere thicknesses for Atla Regio to reside between 20-52 km [14, 15, 24]. Even though these authors used spectral analysis of gravity and topography to determine elastic lithosphere thickness, their estimates are valuable for comparison purposes [14, 15, 24]. Phillips [15] used spectral analysis and Monte Carlo modelling to suggest a range of elastic lithosphere thicknesses between 40-50 km and a mean elastic lithosphere thickness of 45 km. Using this elastic lithosphere thickness as a guide, a thermal gradient of 7-10 K km\(^{-1}\) was found from moment-curvature relationships [15]. The elastic thickness estimate for Atla Regio was consistent with those obtained over the uplifted portion of the slow-moving, East African hot-spots (43-49 km), which experience a total heat flow (from mantle convection and a possible mantle plume) of 20-50 mW m\(^{-2}\) and a temperature gradient of 5-12.5 K km\(^{-1}\) [15, 25].

Phillips [15] inferred Atla Regio to share a similar thermal environment as the East African hot-spots and used an Earth-scaled heat flux of 74 mW m\(^{-2}\) as a guide to determine total heat flow within this area. Assuming a lower bound total heat flow of 80 mW m\(^{-2}\) for Atla Regio and a temperature gradient of 20 K km\(^{-1}\), an elastic lithosphere thickness of 20 km was found [15]. However, this elastic lithosphere thickness is significantly lower than spectral analysis suggested, which led Phillips [15] to propose that Atla Regio should have a lower total heat flow (i.e. < 80 mW m\(^{-2}\)) than what is ascertained from Earth scaling alone.
Phillips [15] also used the mean elastic lithosphere thickness and thermal gradient within Atla Regio to propose a 100 km thick thermal lithosphere, which comprises the crust and top-most portion of the upper-mantle, to occur below this area. A thin thermal lithosphere below Atla Regio, compared to the plains regions (~350 km) could suggest that the area experiences possible heating from the tail of a mantle plume [15].

Turcotte & Schubert [12] estimated the elastic lithosphere thickness below the Hawaiian Archipelago, by taking the distance from the center of the Island of Oahu to the surrounding arch. Using this distance (250 km), a flexural parameter and rigidity of 80 km and $2.4 \times 10^{23}$ N m, respectively, an elastic lithosphere thickness of 34 km was found for this area [12]. The total heat flux at the Hawaiian swell was found from measurements to be $\sim 52.9 \pm 2.3 \text{ mW m}^{-2}$ by Von Herzen et al. [26].

Spectral admittance studies suggest that the thickness of the elastic lithosphere may vary between 20-29 km below the Cape Verde Islands [6]. Best fit surface/subsurface loading models, however, suggest an elastic lithosphere thickness closer to 29 km [6]. Heat flow measurements taken across the Cape Verde Rise were found to increase over the middle of the swell up to $16 \pm 4 \text{ mW m}^{-2}$ above the $45.5 \pm 3.4 \text{ mW m}^{-2}$ heat flow of normal 125 Myr old crust [27].

**Methodology**

Parameters used to calculate the excess temperature associated with a mantle plume at Atla Regio, Oahu and Cape Verde are listed in Table 1.

*Table 1 Parameters used to calculate excess temperature, buoyancy flux and heat flow.*

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Atla Regio (Venus)</th>
<th>Oahu (Earth)</th>
<th>Cape Verde (Earth)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cross-sectional area (km$^2$)</td>
<td>W</td>
<td>3056*</td>
<td>1188*</td>
</tr>
<tr>
<td>Mantle density (kg m$^{-3}$)</td>
<td>$\rho_m$</td>
<td>3330&quot;</td>
<td>3330*</td>
</tr>
<tr>
<td>Surface density (kg m$^{-3}$)</td>
<td>$\rho_s$</td>
<td>67*</td>
<td>1030†</td>
</tr>
<tr>
<td>Stagnation distance (km)</td>
<td>$r_s$</td>
<td>200-600‡</td>
<td>350 – 450‡</td>
</tr>
<tr>
<td>Plume channel thickness (km)</td>
<td>A</td>
<td>200</td>
<td>100‡</td>
</tr>
<tr>
<td>Thermal expansion coefficient (x10$^{-5}$K$^{-1}$)</td>
<td>$\alpha$</td>
<td>3.10#</td>
<td>3.00†</td>
</tr>
<tr>
<td>Specific Heat (x10$^3$ J kg$^{-1}$ K$^{-1}$)‡</td>
<td>$c_p$</td>
<td>1.25</td>
<td>1.25</td>
</tr>
</tbody>
</table>

* Values measured from topographic profiles.
** Value from [28].
† Values from [7].
‡ Density of Venus’ atmosphere at the surface. Value from [29].
†† Density of water. Value from [7].
‡‡ Values from [13].
# Value from [30].
These parameters, as well as Equation [1] can be used to estimate the excess temperature ($T$) due to a mantle plume:

$$T = \left( W(\rho_m - \rho_s) \right) / \left( \pi r_s A \rho_m \alpha \right)$$  \[1\]

where $W$ is the cross-sectional area, $\rho_m$ is mantle density, $\rho_s$ is surface density, which is considered to be atmosphere for Venus (67 kg m$^{-3}$) and water for Earth (1030 kg m$^{-3}$) and $\alpha$ is the coefficient of thermal expansion \[8, 9\].

The excess mass produced due to the presence of the plume can be represented by the buoyancy flux ($B$) delivered to the rise by the plume:

$$B = [(\rho_m - \rho_s) W v]$$  \[2\]

where $W$ is the cross-sectional area and $v$ is the velocity of the hot-spot relative to the lithosphere. This equation follows from Sleep \[9\] and the terminology in this equation is not standardised. Venus has a stationary lithosphere, therefore the closest Earth-based hot-spot analogue would be for a slow-moving plate \[14\]. The $v$ values used for Venus are the lower limit plate velocity values of the slow-moving Cape Verde hot-spot (12 mm yr$^{-1}$) \[27\]. Hot-spot velocities relative to the lithosphere for the Hawaiian swell range from ~83 mm yr$^{-1}$ to 96 mm yr$^{-1}$ \[4, 9, 31, 32\]. A value of 90 mm yr$^{-1}$ was used, similar to Turcotte & Schubert \[12\], since this value is closer to the average plate velocity for the values from \[4, 9, 31, 32\].

The heat flux ($Q_H$) of the mantle plume could then be found using:

$$Q_H = (c_p B) / \alpha$$  \[3\]

The properties of volcanoes, which were calculated from topographic profiles over Atla Regio and the Hawaiian and Cape Verde swells, included basal diameter, height, average slope, root mean square (RMS) slope and volume. Topographic profiles across the edifice of the volcano were observed and the basal extents defined where the edifice joined the surrounding volcanic flanks. The heights, volumes, average slope and RMS slope of the volcano were then measured above the edifice base.

Modelling of the lithospheric structure below Atla Regio was completed using the finite-difference code GEO3Dmod \[21\]. GEO3Dmod is an interactive 3D forward modelling software, which calculates the thermal, pressure and density structure for a given model, as well as elevation, free-air gravity, bouguer gravity, geoid and heat flow \[21\]. By fitting these calculated values to observables, the models are more tightly constrained than by singularly fitting each observable \[21, 22, 33\]. GEO3Dmod assumes a conductive thermal lithosphere, comprising the crust and the conductive portion of the upper mantle \[21, 22, 33\]. See \[21, 22, 33\] for more details.

Lithospheric flexure calculations were also used to determine the thermal gradient and in turn heat flow through an area. The half-width of the depression ($x$) was located beneath the load and the distance ($d$) to the forebulge ($x_b$) calculated (Figure 3).

This distance ($d$) was then used to determine the flexural parameter ($F$) using:

$$F = d / \pi$$  \[4\]

Using the flexural parameter ($F$), the flexural rigidity ($D$) of the plate was found:
\[ D = F^4 (\rho_m - \rho_s) g \]  

where \( g \) is the acceleration due to gravity (9.81 m s\(^{-2}\) for the Earth and 8.87 m s\(^{-2}\) for Venus).

The flexural rigidity was then used to estimate the elastic lithospheric thickness \((T_e)\) from:

\[ T_e = [D x 12 (1 - \nu^2) / E]^{1/3} \]

where \( \nu \) is Poisson's ratio (0.25) and \( E \) is Young's modulus (70 GPa), similar to the values used by [12].

By using a reference lithosphere and moment-curvature relationships, Phillips et al. [14] determined a suitable equation to determine a linear thermal gradient \((dT/dZ)\) for elastic lithosphere thicknesses below certain Venusian features:

\[ dT/dZ = 9.54 (T_e / 30)^{-0.817} \]

Using this temperature gradient \((dT/dZ)\) and the thermal conductivity \((k)\) at the base of the lithosphere, the total heat flux \((q)\) of the area was determined by using the equation:

\[ q = k -dT/dZ \]

Estimates of the thermal conductivities at the base of the lithosphere were \(~3.3\) W m\(^{-1}\) K\(^{-1}\) [30] for Venus and \(2.55\) W m\(^{-1}\) K\(^{-1}\) for Hawaii [34, 35]. It should also be noted that the total heat flux calculated by Equation [8] includes both a heat flux from a possible upwelling mantle plume and the background heat flux associated with mantle convection [14].

**Results**

From Equation [1] the excess temperature associated with a mantle plume below Atla Regio was found to range from 85-769°C for stagnation distances between 200-600 km. Venus has been proposed to have a stationary lithosphere (plate velocity = 0 mm yr\(^{-1}\)). However, due to the dependency of buoyancy flux and heat flow equations on a plate velocity, a hot-spot located on a slow-moving plate was considered to be the closest Earth-based analogue [9, 14].

By using a lower-limit velocity (12 mm yr\(^{-1}\)) of the Cape Verde hot-spot, relative to the lithosphere [31], and Equation [2-3], a buoyancy flux of 3.79 Mg s\(^{-1}\) and a mantle plume heat flux of 1.53x10\(^{11}\) W at Atla Regio was obtained. Therefore, the range of excess heat associated with a mantle plume on Venus would be less than 1.53x10\(^{11}\) W.

The excess temperature associated with Oahu, a fast-moving hot-spot, and was found to range from 193-249°C for stagnation distances between 350-450 km. Based on a plate velocity of 90 mm yr\(^{-1}\) [11], a buoyancy flux of 7.79 Mg s\(^{-1}\) and a mantle plume heat flow of 3.25x10\(^{11}\) W was found for Oahu. The excess temperature of Cape Verde, a slow-moving hot-spot, was found to be 206 K for a stagnation distance of 390 km. Using a plate velocity of 12 mm yr\(^{-1}\), a buoyancy flux of 0.96 Mg s\(^{-1}\) and a heat flow of 4.0x10\(^{10}\) W was found.

Results for calculating the basal diameter, height, average slope and RMS slope for the three volcanoes within Atla Regio and comparing these to Hawaii and Pico de Fogo (within the Cape Verde Archipelago), can be observed in Table 2. The volcanoes located at Atla Regio have larger basal diameters (405-630 km), and except for Sapas Mons, larger volumes (30.5-42.3x10\(^4\) km\(^3\)), than the Island of Hawaii and Pico de Fogo, which have a basal diameters of
200 km and 63 km and volumes of $12.5 \times 10^4$ km$^3$ and $1.46 \times 10^4$ km$^3$, respectively.

All the volcanoes located within Atla Regio also have smaller heights (2.34, 3.85 km, 6.8km for Sapas, Ozza and Maat Mons, respectively) when compared to Hawaii (8.48 km). However, only Ozza Mons and Sapas Mons have smaller heights when comparing to Pico de Fogo (6.28 km). The RMS slope of each volcano was also calculated, however, since long slope-frequency distributions will bias the RMS slope to greater values, an average of the slopes measured for each volcanic edifice were also calculated [36, 37]. Shallower RMS slopes and average slopes were obtained for the volcanoes within Atla Regio (0.646°-2.19° and 0.537°-1.66°), compared to Hawaii (5.60° and 5.01°), and Pico de Fogo (13.55° and 12.06°). Since Pico de Fogo is a stratovolcano it has steeper slopes compared to the volcanoes within Hawaii and Atla Regio, which are shield volcanoes [17, 18, 38].

### Table 2 Volcano characteristics

<table>
<thead>
<tr>
<th>Volcano</th>
<th>Basal Diameter (km)</th>
<th>Height (km)</th>
<th>RMS Slope (degrees)</th>
<th>Average Slope (degrees)</th>
<th>Volume ($x10^4$ km$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sapas Mons</td>
<td>630</td>
<td>2.34</td>
<td>0.646</td>
<td>0.537</td>
<td>10.9</td>
</tr>
<tr>
<td>Maat Mons</td>
<td>405</td>
<td>6.8</td>
<td>2.19</td>
<td>1.66</td>
<td>30.5</td>
</tr>
<tr>
<td>Ozza Mons</td>
<td>495</td>
<td>3.85</td>
<td>0.719</td>
<td>0.764</td>
<td>42.3</td>
</tr>
<tr>
<td>Hawaii</td>
<td>200</td>
<td>8.48</td>
<td>5.60</td>
<td>5.01</td>
<td>12.5</td>
</tr>
<tr>
<td>Pico de Fogo</td>
<td>63</td>
<td>6.28</td>
<td>13.55</td>
<td>12.06</td>
<td>1.46</td>
</tr>
</tbody>
</table>

A possible lithospheric model below Sapas Mons is shown in Figure 4. The topographically high Sapas Mons volcano between 1000-2300 km, corresponds to a bouguer low, reflecting a possible lower-density partial melting/magma chamber. This compares to the topographically low plains region surrounding Sapas Mons (0-1000 km) which has a thinner crust (~20 km). A thinner (~70 km) thermal lithosphere (the crust and upper mantle layers) and higher surface heat flow (~38 mW m$^{-2}$), occurs below Sapas Mons and Atla Regio, which could be associated with the presence of a mantle plume.

Flexural modelling at Atla Regio and Oahu can be observed in Figures 3 & 5. The distance between the half-width of the depression ($\chi$) and the maximum amplitude of the forebulge ($\chi_b$) were found to be 324 km, 310 km and 212 km for Atla Regio, Oahu and Cape Verde, respectively.

Using Equations [4] and [5] a flexural parameter of 103 km, 98.7 km and 67.5 km and a flexural rigidity of $8.19 \times 10^{23}$ N m, $5.34 \times 10^{23}$ N m and $1.17 \times 10^{23}$ N m were found for Atla Regio, Oahu, and Cape Verde, respectively. Substituting these flexural rigidity estimates into Equation [6], an elastic lithosphere thickness of 50.9 km, 44.1 km and 26.6 km were found for Atla Regio, Oahu and Cape Verde, respectively. The thermal gradient for Atla Regio was calculated from Equation [7] and found to be 6.19 K km$^{-1}$. Using this thermal gradient with Equation [8] a total heat flux of 20.4 mW m$^{-2}$ for basal lithospheric thermal conductivities of 3.3 W m$^{-1}$ K$^{-1}$ was found.

Since Equation [7] is specific of Venus, the thermal gradient at Oahu was found from a simple oceanic geotherm [12] and yielded a geothermal gradient of 23 K km$^{-1}$ and 20 K km$^{-1}$ for Oahu and Cape Verde, respectively. This gradient and a thermal conductivity of 2.55 W m$^{-1}$ K$^{-1}$...
was used to determine a total heat flux of 59 mW m\(^{-2}\) and 51 mW m\(^{-2}\) for Oahu and Cape Verde, respectively.

Figure 4 Profile across Sapas Mons at 8° latitude, showing the crust (yellow and red) and conductive upper mantle (orange) that comprise the thermal lithosphere. The adiabatic sub-lithosphere is shown in purple. Note: The plot of heat flow has no observable. Scale bar: 10° Longitude = 1090 km

Figure 5 North-South topographic profile across Oahu, Hawaii. The red line is the constant-depth line, facing arrows infer an elastic lithosphere of thickness (h).

**Discussion/Conclusion**

The excess temperature due to a mantle plume on Venus was found to range from 85-769 °C for unconstrained stagnation distances between 200-600 km, compared to the excess temperature of 193°-249°C and 206 K for more constrained stagnation distances of 350-450 km and 390 km for the Hawaiian and Cape Verde swells, respectively. One reason for a large range of excess temperatures at Atla Regio compared to Oahu and Cape Verde is that this area has a larger cross-sectional area ~3056 km\(^2\) compared to Oahu (1188 km\(^2\)) and Cape Verde.
Due to Venus' lack of water, a higher density contrast between mantle and air also produced higher excess temperatures compared to the smaller density contrast arising from mantle and water on Earth. The equation for excess temperature also relies on a stagnation distance, which is calculated from the radial velocity of the plume, the average unperturbed asthenospheric velocity and the half plate velocity at the hot-spot (Equation 8 in [9]). The stagnation distances for the Hawaiian and Cape Verde swells are also better constrained, than those for Atla Regio, since Venus has a stationary lithosphere, with zero plate velocities. The closest approximation of stagnation distances on Venus may be the diameter of coronae, due to their possible mantle plume origin (200-600 km) [13]. The existence of an asthenospheric channel that extends throughout the entire mantle is questionable on Venus [10], however our calculations of excess temperature assumed an asthenospheric channel thickness. This thickness was more tightly constrained for Hawaii (~100±16 km), compared to Atla Regio (~200 ± 100 km) [8, 12].

The buoyancy flux and heat flow values obtained also rely on the velocity of the hot-spot relative to the lithosphere [8, 9]. For Oahu the hot-spot velocity ranged from 83 mm yr^{-1} to 96 mm yr^{-1} [4, 9, 31, 32], with ~90 mm yr^{-1} being chosen in this study, similar to the value used by [12]. This yielded a mantle plume heat flux of 3.25x10^{11} W at Oahu, slightly higher than the mantle plume heat flux of 3x10^{11} W obtained by [12] at a similar location.

Buoyancy flux and heat flow estimates for a mantle plume at Atla Regio, however, have a greater uncertainty than values obtained for Hawaii, since Venus has a stationary lithosphere [1]. The closest Earth-based hot-spot analogue to Venus are those located on slow-moving plates [14]. By taking the lowest estimate of the plate velocity of the Cape Verde hot-spot (12 mm yr^{-1} [27]), and taking the heat flux obtained as an upper bound, heat flux from a mantle plume through Atla Regio should range from 0 W to 1.53x10^{11} W. This compares to the heat flux of 4.0x10^{10} W found for Cape Verde. Even though the same plate velocities were used for Atla Regio and Cape Verde, the differences in cross-sectional area and surface density contrasts caused the variation in heat flow estimates for each of these areas. Despite the high uncertainties and large range of estimated values for excess temperature, buoyancy flux and heat flow, this study has showed that dynamic support from a mantle plume may be contributing to the topography at Atla Regio.

The volcanoes at Atla Regio have larger basal diameters and volumes (except Sapas Mons), but lower elevations (except Maat Mons) than Hawaii and Pico de Fogo. This is consistent with the general trend that volcanoes on Venus have larger basal diameters and volumes than shield volcanoes on Earth [1, 18]. This difference can be attributed to Venus' lack of plates and the associated velocities over hot-spots [1]. On Venus, volcanoes would remain over the hot-spot for longer time intervals compared to Earths' volcanoes, where a plate would move away from the hot-spot over time [1]. Volcanoes within Atla Regio also tend to have shallower slopes than Hawaii and Pico de Fogo, which could be due to: 1) higher temperatures within the crust and surface; 2) lower-viscosity lavas; 3) larger volumes of magma erupted at higher effusion rates; 4) propagation of lavas by lava tubes and; 5) higher surface pressures [18, 19, 20].

The thin thermal lithosphere, high heat flow and presence of a low-density root below Sapas Mons could reflect current volcanism associated with an upwelling mantle plume. The excess heat associated with a mantle plume would act to thin the thermal lithosphere, producing a geoid high, and causing partial melting below this area. The associated bouguer low could represent either this partial melt or a possible low-density magma chamber.
The elastic lithosphere thickness of 50.9 km, obtained at Atla Regio from flexural modelling, is close to the upper limit of the range (40-50 km) of elastic lithosphere thicknesses from Phillips [15], who used spectral analysis and Monte Carlo inversions. The elastic lithosphere thickness obtained at Oahu (44.1 km) is higher than that obtained by Turcotte & Schubert [12] over the Hawaiian Archipelago (34 km). The elastic lithosphere thickness at the Cape Verde Archipelago was found to be 26.6 km, within the range found by [6] from spectral admittance studies, but lower than the 29 km found from best-fit surface/subsurface loading models. These differences in estimates of elastic lithosphere thicknesses may be caused from the difficulties associated with determining the maximum amplitude of the forebulge due to interference from the surrounding topography that acts to obscure the flexural upwarp, similar to what is observed on the left-side of the profile in Figure 4.

The thermal gradient of 6.19 K km\(^{-1}\) for Atla Regio, obtained indirectly from flexural modelling, was also within the range of thermal gradients [7-10 K km\(^{-1}\)] outlined by Phillips [15]. Phillips [15] suggested Atla Regio should have a total heat flux less than what is obtained from Earth-scaling (~80 mW m\(^{-2}\)). From flexural modelling a total heat flux of 18.6-24.8 mW m\(^{-2}\) was found for Atla Regio, which is significantly lower than that obtained by Earth scaling. The total heat flux value obtained from flexural modelling at Oahu is slightly higher (58.7 mW m\(^{-2}\)) than those obtained from total heat flux measurements taken at Hawaii (52.9±2.5 mW m\(^{-2}\) [26]), whilst the obtained heat flux value for Cape Verde is is slightly less (51 mW m\(^{-2}\)) than measured values (61±7.4 mW m\(^{-2}\) [27]). Differences in the heat flow estimates obtained from flexural modelling to measured values are most likely caused from comparing these areas to a globally averaged oceanic geotherm.

The results of this study suggests that volcanic topography at Atla Regio could be produced by three processes, thermal, volcanic and flexural. The impingement of a mantle plume on the lithosphere could have initiated rifting and volcanism, resulting in volcanic loading and flexure of the lithosphere. These processes act in a similar way to those observed on Earth, but yield different results due to Venus’ unique surface conditions. Future work aims to constrain the stagnation distances through finite-element modelling and to produce a plausible rheological structure below Atla Regio, which will be used to constrain these processes further.

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References

3. Phillips, R.J., Grimm, R.E., & Malin, M.C. “Hot-spot evolution and the global


