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MEGAPLUMES ON VENUS. W. M. Kaula, University of California, Los Angeles.

The geoid and topography heights of Atla Regio and Beta Regio, both peaks and slopes, appear explicable as steady-state plumes, if non-linear viscosity $\eta(T, \dot{\epsilon})$ is taken into account. Strongly constrained by the data are an effective plume depth of about 700 km, with a temperature anomaly thereat of about 30°, leading to more than 400° at the plumehead. Also well constrained is the combination $Q\eta/s_0^4$ =(volume flow rate)Xviscosity/(plume radius)$^4$: about 11 Pa/m/sec. The topographic slopes $dh/ds$constrain the combination $Q/A$ , where $A$ is the thickness of the spreading layer, since the slope varies inversely with velocity. The geoid slopes $dN/ds$ require enhancement of the deeper flow, as expected from non-linear viscosity. The Beta data are best fit by $Q$ =500 m$^3$/sec and $A$ = 140 km; the Atla, by $Q$ =440 m$^3$/sec and $A$ =260 km. The dynamic contribution to the topographic slope is minor.

A major controversy about solid Venus is whether its contemporary rate-of-heat-loss (on a 100 My timescale), and thence its level of volcanism & tectonism, is (1) much less than the rate averaged over about 1 Gy, due to an oscillatory character arising from material properties– most obviously, density differences of constituents– or (2) near this averaged rate (within a factor of two), as appears to be true for the Earth. Rather moot between hypotheses (1) and (2) is the important finding of Pioneer Venus, confirmed by Magellan, of a much higher ratio of gravity-to-topography than Earth's, which requires a much stiffer upper mantle. Supporting hypothesis (1) are (a) a considerably smaller rms variation in topographic elevation about the mean than exists on Earth; (b) the absence of the topographic signature of a plate tectonic spreading system [1]; and (c) a consistency of the horizontal distribution of craters with randomness [2]. To claim hypothesis (2) against property (a) is to say that Venus's averaged level of activity is lower than Earth's in recent Gy; against property (b), to say that convective heat transfer on Venus is appreciably more regional than Earth's. Property (c) is discussed in another abstract to the effect that consistency does not entail necessity [3]. Supporting hypothesis (2) more directly is the extraordinary magnitude of some extremes in the topography and gravity of Venus, such as (d) the steep slopes of the western front of Maxwell Montes, and (e) the great geoid highs over Beta Regio and Atla Regio: 20 percent higher than the Earth's maximum. We explore whether the data (e) are compatible with hypothesis (2): are Atla and Beta explicable as steady-state flow features? For this, it is probably necessary to take into account the greater stiffness of Venus's upper mantle, mentioned above.

The model is stimulated by Sleep's for sub-oceanic plumes on Earth [4] . A significant difference from Sleep's concern is that the Venus plume is a primary phenomenon in its heat transfer, rather than being secondary to lithospheric spreading from a rise elsewhere. Hence in a Venus-relevant model the overlayer is a passive crust, through which heat transfer is conductive, and the axisymmetric flow from the plume is analogous to to the bilateral flow from the rise, in that the topography and geoid are controlled by conductive cooling. However, an important difference of the spreading from a plume is that conservation of volume flow leads to an inverse correlation of velocity with distance from the rise, and hence, in the simplest case of a steady-state cooling half-space, a decrease in height linear with distance, rather than with the square-root thereof. The reduction in strain-rate with velocity as well as the cooling act to increase the effective viscosity. Hence there may be a dynamic contribution to the topographic and geoidal slopes. To drive this horizontal flow there must be a pressure generated by the buoyancy of the plume, $P = \rho g \alpha \int_0^K \Delta T dz$, where $K$ is the depth of the plume and the other symbols have their usual meanings.

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In an actual planet, the density anomaly $\Delta T$ will vary with depth $z$, since the plume will have a temperature gradient $dT/dz$ closer to adiabatic than its surroundings. We divide the problem in two parts. Firstly, from the peaks in geoid height $N$ and topographic height $h$ we can infer, as an inverse problem, the temperature anomaly $\Delta T$ and the depth of the plume $K$. These lead to the pressure $P$ and its gradient. Then the product $Q\eta/s_0^4$, as defined in the first paragraph, can be inferred from the Poiseuille formula [5]: $Q\eta/s_0^4 = \pi(dP/dz)/8 = \pi\rho g\alpha\Delta T/8$. To separate $Q\eta$ from the plume radius $s_0$, an assumption about the ratio $s_0/K$ must be made. But the high power to which $s_0$ appears in the constrained combination limits it plausible range. The product $Q\eta$ has the dimensions of energy, like the flexural ridigity: first-order tectonic geometry and dimensions can always be fit by steady state viscous flow as well as by elastic flexure. To separate $Q$ from $\eta$, a rheology must be assumed. A form $\eta = B\exp(CT_M/T)/\dot{\epsilon}^m$ is used; values corresponding to the results of Karato et al [6] are $B$ =24200, $C$ =8.91, $m$ =0.71. This is a weak point not only in the extrapolation from the laboratory, but also in the effective temperature– or corresponding depth $z(\eta)$– which probably is near the plume head within the spreading layer, but must be assumed.

However, from the radial topographic slope $dh/ds = (dh/dt)/v$, due to cooling, there is a different constraint from conservation of the volume flow rate: $Q = 2\pi sAv$. To compute the thermal decrease $dh_T/dt$, the conductive formula for $dT(z)/dt$ for one boundary at a fixed temperature and the other at zero heat flow [7] was used. For the dynamic slope $dh_D/ds$, we need the pressure gradient $dP/ds = -\Psi Q\eta/\pi A^3 s$, where $\Psi$ is 6 for a fixed boundary and 2 for a free boundary: more realistic, and always used. After a complete set of temperatures and topographic heights are calculated, the geoid heights $N(s)$ are obtained by a two-dimensional FFT: the most time-consuming step. To get geoid slopes $dN/ds$ as steep as observed, it was necessary to assume that the non-linear rheology led to a deeper skewing and confinement of the velocity: i.e., terms proportionate to $z^3$and $z^4$ in $v(z)$.

The principal defining parameters, and values explored, were thus the observed peaks of 3.5 km in topography and 110 m in geoid height, and slopes $dh/ds = 0.0017$ and $dN/ds = 0.03 - 0.04 m/km$; the viscosity parameter $B$ : 24200 – 242000; plus ratios $A/K : 0.2 - 0.5; s_o/K : 0.1 - 0.2; z(\eta)/K : .05 - 0.2; v_3/v : 0 - 1; v_4/v : 0 - 2;, C/K : .02 - .03$; and $D/K : 0 - 0.25$. $C$ is the thickness of the passive crust, and $D$ is the "dynamic" part of the plume, leading to an apron around the peak, $h(s)$, $s < s_0$, which must be included in the mass balance. Values corresponding to a best fits were $B$= 80000, $v_3/v$ =1, $v_4/v$ =4, $z(\eta)/K$ =.05, $C/K$ =.03, $D/K$ =0 for both plumes; $A/K$ = 0.2 and $S_0/K$ = 0.175 for Beta; $A/K$ = 0.4 and $S_0/K$ = 0.2 for Atla.

The steepness of the dropoffs is the main constraint on the volume flows $Q$. The total of about 950 $m^3$/sec is about one-third the *minimum*for the Earth's spreading rises (from area/year X crustal thickness X cumulate factor = 3 X 7 X 4 $km^3$/yr).

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