MODELS FOR THE CRUSTAL STRUCTURE OF IO: IMPLICATIONS FOR MAGMA DYNAMICS. W. L. Jaeger[1] and A. G. Davies[2], [1]Astrogeology Research Program, U.S. Geological Survey, 2255 N. Gemini Dr., Flagstaff, AZ 86001, wjaeger@usgs.gov, [2] Jet Propulsion Laboratory, California Institute of Technology, 4800 Oak Grove Dr., Pasadena, CA 91109.

Introduction: Io, the innermost of the Galilean satellites, is the most volcanically active body in our solar system. While its volcanic eruptions can be studied directly, our knowledge of the underlying magmatic processes rests wholly on indirect inferences. Here we combine constraints from magma ascent and crustal density models to gain insight into the plumbing system underlying the Prometheus volcanic center.

Magma Ascent Model: Prometheus was selected for focused study because it is the best known example of a long-lived flow field on Io. It also exhibits many similarities to the ongoing eruption of Kilauea Volcano, Hawaii [1]. However, despite their comparable eruptive styles, Prometheus and Kilauea have magma fluxes that differ by about an order of magnitude. During one ~6-month period (1996-1997), Prometheus erupted ~0.8 km^3 of lava, with an estimated effusion rate peaking at ~70 m^3 s^{-1}, whereas Kilauea erupted 0.08 km^3 over the same period.

Building upon the work of [2-4], it is possible to relate the observed effusion rate to the subsurface plumbing system. For a cylindrical conduit, the effusion rate (Q_F) can be found by

$$Q_F=(\pi/4)\,[D^5\, g\, (\rho_c-\rho_m)/(2\, f\, \rho_m)]^{1/2} \qquad (1)$$

where D is the conduit diameter, g is gravitational acceleration (1.8 m s^{-2}), ρ_c is the bulk crust density, ρ_m is the magma density, and f is a friction factor (~0.01). As Equation (1) shows, the difference in the effusion rates of Prometheus and Kilauea could be explained if the former has a larger conduit diameter or a greater density contrast. However, the values for ρ_c and ρ_m suggested by [4] (2200 and 2600 kg m^{-3}, respectively), result in a negatively buoyant magma that would be expected to pond within the crust at a depth given by $d[1-(\rho_c/\rho_m)]$, where d is crustal thickness. For a 30-km-thick crust, the magma chamber is predicted to form at ~5 km. While it is possible for overpressurization of a shallow magma chamber such as this to trigger an eruption, it is also worth investigating whether a more robust density structure model will transport magma to the surface.

Crustal Density Model: The density structure of Io's crust is, to first order, a function of its composition, porosity, pressure gradient and temperature profile. Composition is perhaps the best understood of these variables. Several lines of evidence suggest that the silicate portion of Io is undifferentiated (e.g., high eruptive temperatures [5-7], lava morphologies indicative of low viscosity flows [8], and the spectral signature of orthopyroxene [9]). Thus, a mafic or ultramafic composition similar to basalts or komatiites is a reasonable assumption. The relatively high density of these mafic/ultramafic lavas will be reduced to some extent by their vesicularity. Based on terrestrial and lunar mafic lava flows, a vesicularity of ~25% is likely correct to within a factor of 2 [10-13]. However, the fractional void space (u) decreases exponentially with increasing pressure according to the equation

$$u=u_0 \exp(-\lambda P) \qquad (2)$$

where u_0 is the fractional void space at the surface, λ is a constant empirically determined to be ~1.18 x 10^{-8} Pa^{-1}, and P is pressure [2,11]. For most planetary bodies, pressure can be taken as lithostatic, but for Io this assumption is inadequate. Both subsidence-induced and thermal horizontal compressive stresses exceed lithostatic pressure over certain depth ranges, with subsidence (driven by the superposition of successive lava flows) having the largest integrated effect on the crust [14]. In fact, the subsidence-induced stress builds so rapidly that at a depth of only 4 km, it exceeds the compressive strength of most mafic igneous rocks. Once the rock is fractured, the pressure profile is described by Amonton's law of frictional sliding.

As with pressure, the temperature profile on Io is unlike that of other planetary bodies. Studies frequently assume a linear geothermal gradient; however, [15] showed that rapid subsidence (≥1 cm/yr) leads to an essentially isothermal crust with only the base being strongly heated by conduction. The resulting temperature profile ($T(z)$) can be expressed as follows:

$$T(z) = T_0 + \left(T_a - T_0\right)\frac{e^{z/l}-1}{e^{d/l}-1} \qquad (3)$$

where z is depth, T_0 is surface temperature, T_a is the temperature at the base of the crust, l is the ratio of thermal diffusivity to subsidence rate, and d is crustal thickness (~30-50 km [14,16]) [15].

Hitherto we have described our approach to modeling the silicate portion of Io's crust, but it is also necessary to address the ubiquitous sulfur-rich volatiles. Observations of the gas plume at Pele suggest an

S:SO_2 ratio of 1:7 and an overall volatile content of 0.3 wt. % [17]. Because Pele's volatiles are in equilibrium with the melt, we will take 0.3 wt. % as an approximation of the amount of volatiles that can exsolve from bulk silicate Io. Thus, our preferred crustal composition contains 91 vol. % mafic silicates, 8 vol. % SO_2 and 1 vol. % S for a 30-km-thick crust. The SO_2 is modeled as filling void space in the lava flows to the extent that it is available. Because SO_2 exists as a fluid for all but the shallowest depths, it is thought to reside only in the upper part of Io's crust, decreasing in abundance with depth as pore space is lost. Elemental sulfur, which remains solid throughout most of Io's crust, is assumed to be homogeneously interbedded with the silicate lava flows.

Results: Typical model results are shown in Figure 1. For the preferred melt density (ρ_m) of 2800 kg m^{-3}, the estimated neutral density depth is ~1 km, which is comparable to the ~5 km found using the densities of [4]. However, our refined density structure allows for some intriguing new possibilities. Because ascending magma is likely to pond within (or, more likely, at the base of) the SO_2-rich layer at the top of the crust, intrusions will melt and vaporize the overlying volatiles, eventually unroofing themselves. This is consistent with the hypothesis that many of Io's paterae are lava pools formed by the exhumation of sills beneath a volatile-rich layer. For a wide range of reasonable input parameters, the model also suggests that intrusions will be largely confined to the uppermost crust and so should not significantly alter the geothermal gradient of [15]. Therefore, our model is self-consistent in the way it treats temperature.

Another important implication of our model is that variations in the thickness of the SO_2-rich surface layer may affect whether intrusive or extrusive activity occurs. In fact, once an intrusive sill has sufficiently thinned the overlying volatile blanket, renewed activity is likely to be extrusive. If the SO_2-rich layer is removed entirely, the bulk density of a 30-km-thick crust increases to ~2850 kg m^{-3}. Assuming this to be the case at the Prometheus volcanic center, Equation (1) predicts that a conduit 6 m in diameter is required to explain the peak observed effusion rate. In future work we hope to apply our model to other volcanic centers in order to better understand whether differences in magmatic systems can explain some of Io's volcanic diversity.

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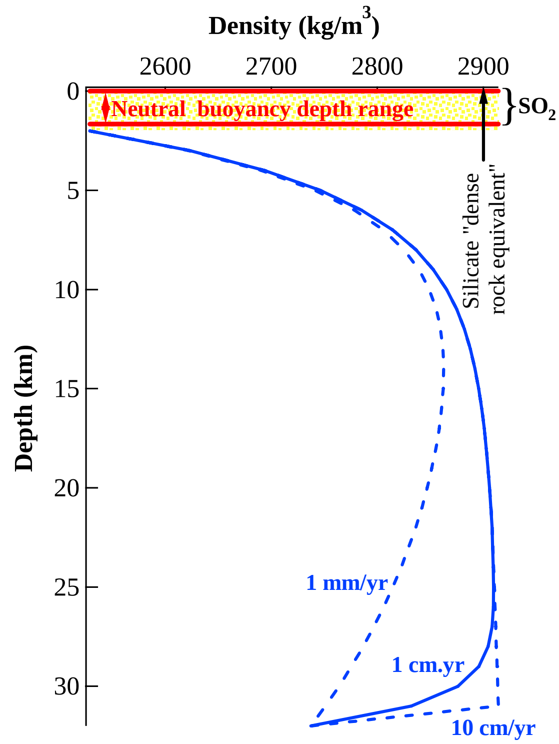

Figure 1. Model results for the density structure of Io's crust. The yellow zone is a low-density surface layer of residual SO_2 (i.e., that which could not be accommodated in pore space). The three blue trends are density profiles for the vertically recycling, silicate-dominated portion of the crust. Each corresponds to a different subsidence rate, with the (generally accepted) rate of 1 cm/yr drawn as a solid line. The density increase at shallow depths is due to the collapse of pore space, and the reduction in density at the base of the crust is due to thermal expansion. These trends were computed assuming d=30 km and T_a=1500 K, but altering d or T_a within reason does not change our results significantly. The red bars bracket the depths (~0-2 km) where rising magma is neutrally buoyant, assuming a melt/magma density contrast typical of mafic systems.

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