LINKS BETWEEN DEPTHS OF MAGMA RESERVOIRS AND VOLCANIC ERUPTION RATES ON IO.

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Abstract: We (1) review the volume eruption rates deduced for various volcanic events on Io, (2) infer likely depths of magma reservoirs feeding the eruptions and (3) show that the observed eruption rates can be sustained from sufficiently voluminous crustal reservoirs.

Observations: The only eruption for which there has been a direct observation of the length of the surface fissure is that of Tvashtar [1]. By modeling the dynamics of the lava fountain produced, Wilson and Head [2] estimated that the ~25 km long fissure discharged magma at a rate of q = 0.7 to 7 m^3 s^{-1} per meter of fissure length along strike, implying a total flux Q of ~2 x 10^4 to 2 x 10^5 m^3 s^{-1}. The ~20 years of growth of the compound lava flow field from the Prometheus volcanic center [3] implies a much smaller average total supply rate of Q = 100-200 m^3 s^{-1} [3], consistent with the way this flow field is growing by many small pahoehoe breakouts at any one time [4]. However, the along-strike horizontal length of the feeder dike cannot be estimated because the flow source appears to be an overflow from a lava lake in the Prometheus caldera. This problem of identifying source geometry applies to most eruptions on Io due to the small amount of very high resolution coverage from Galileo. Volume fluxes for other volcanic centers have been estimated to be ~300 m^3 s^{-1} at the Pele lava lake between 1996 and 1998 [5] and ~2000-7000 m^3 s^{-1} for the Pillan lava flow in 1997 [5, 6]. Even larger eruption rates have been inferred from thermal fluxes: between ~2 x 10^5 and ~8 x 10^5 m^3 s^{-1} at Loki in 1990 [7, 8]. If these high fluxes are commonly being released from fissures at least several km long, consistent with the few observations that we have, then fluxes per meter of fissure length along strike, q, must be in the range ~1 to ~100 m^3 s^{-1} per meter.

Controls on eruption rates: Magmas can reach the surface of a planet either directly from mantle partial melt zones or from reservoirs somewhere in the lithosphere. Direct rise from the mantle requires the magmas to be positively buoyant all the way to the surface. However, the volcanic resurfacing rate on Io is so large that the crust is probably a very thick stack of volcanics, a mixture of lavas and pyroclastics. Such a crust will have a low mean density at the surface and will compact only slowly with depth due to the low acceleration due to gravity on Io [9]. We would therefore expect there to be density traps within the lithosphere at which possibly very large magma reservoirs evolve, and this inference seems to be supported by the presence of many large calderas on Io. It is hard to predict the depths at which these reservoirs should be located, however, because of the extreme uncertainties about the volatile contents of juvenile mantle magmas on Io [9].

Magma accumulating in a reservoir at a density trap will only erupt to the surface if either the reservoir is pressurized or the density of the magma is reduced. We have previously discussed ways in which otherwise volatile-poor magmas on Io might absorb SO_2 from sub-surface aquifers created by the progressive burial and eventual melting of SO_2 deposited on the surface in earlier eruptions [9]. Magma buoyancies of a few 10s of kg m^{-3} might be generated in this way. Alternatively, arrival of new batches of magma into reservoirs whose host rocks respond elastically will lead to pressure increases. In general, crustal reservoirs must contain fluids with excess pressures of several tens of MPa in order to ensure that the stresses acting across the roof of the reservoir are not so great that the roof fails in tension and a new dike is created or so small that the roof fails in compression and caldera subsidence occurs [10, 11]. Indeed, it is the creation of such imbalances that initiates eruptions and caldera formation events.

Depths of magma reservoirs: We first use the criterion that new eruptions are initiated when the pressure P_{roof} in the magma at the roof of a reservoir just exceeds the external lithostatic load P_{crust} = (ρ_c g D), where ρ_c is the eman density of the crust, by a critical amount P_{crit} equal toabout twice the tensile strength of the host rocks [12]. If the latter is ~5 MPa we have

$$(P_{roof} - P_{crust}) = P_{crit} = \sim 10 \text{ MPa.} \quad (1).$$

If an eruption is to occur, the pressure difference P_{erupt} driving the motion of the magma to the surface against wall friction is less than P_{roof} by the static weight of the column of magma extending to the surface, (ρ_m g D), where D is the depth of the roof of the reservoir below the surface. Thus

$$P_{erupt} = P_{roof} - (\rho_m g D) \quad (2)$$

and using (1)

$$P_{erupt} = P_{crit} - g D (\rho_m - \rho_c) \quad (3).$$

Hence the maximum depth D_{max} from which eruptions can occur is found by setting P_{erupt} = 0 and is

$$D_{max} = P_{crit} / [g (\rho_m - \rho_c)] \quad (4).$$

Table 1 shows the values of D_{max} implied for the Earth, Venus, Mars and Io (the bodies on which the presence of calderas implies the existance of shallow magma reservoirs at neutral buoyancy levels) using a magma density of 2600 kg m^{-3} and a crustal density of 2200 kg m^{-3}, values appropriate to a basaltic magma on Earth. Terrestrial reservoir depths, such as that at Kilauea, are somewhat shallower (~1.5-2 km) than the value shown, which is not surprising since some of the pressure in the reservoir must be used to drive the negatively buoyant magma to the surface.

Eruption rates on Io: We next calculate the expected magma rise rates through dikes from magma reservoirs having their roofs at depth D less than the maximum value D_{max}. The pressure gradient dP/dz driving the magma is equal to (P_{erupt} / D) which from (3) is

$$dP/dz = (P_{crit} / D) - g\,(\rho_m - \rho_c) \quad (5).$$

The mean width W of the dike is given by [13] as

$$W = [(1-\nu)/\mu]\, D\, [(\pi/2)\, P_{crit} + (1/3)\, D\, g\, (\rho_m - \rho_c)] \quad (6)$$

where ν = ~0.25 is the Poisson's ratio and μ = ~10 GPa is the shear modulus of the crustal rocks. The rise speed u of the magma is given in laminar flow by

$$u = (W^2\, dP/dz) / (12\, \eta) \quad (7)$$

where η is the magma viscosity and in turbulent flow by

$$u = [(2\, W\, dP/dz) / (f\, \rho_m)]^{1/2} \quad (8)$$

where f is a friction factor = ~10^{-2}. In practice the required value of u is the smaller of the values given by (7) and (8). Once u is known, q is found from the continuity requirement $q = (u\, w)$.

Table 2 shows the values of D, dP/dz, u and q for the eruption of a magma with η = 0.3 Pa s, a value appropriate to komatiitic magmas as suggested by [14]. The magma motion is turbulent in all cases.

Conclusions: The magma fluxes shown in Table 2, from ~12 to ~65 m^3 s^{-1} per meter length of surface fissure vent, span most of the ~1 to ~100 m^3 s^{-1} m^{-1} range deduced for observed eruptions on Io. Although we do not have room to give more examples here, we have found that geologically plausible adjustments to the assumed magma density ρ_m, average crustal density ρ_c, critical stress for dike initiation P_{crit} and fissure length along strike can accommodate the remaining differences. Thus, eruptions from sufficiently voluminous crustal reservoirs can, it appears, explain all of the activity so far observed on Io.

This does not, however, preclude the possibility that some eruptions are fed directly from mantle magma sources: all but the very smallest of the magma flow rates mentioned earlier could be maintained against thermal losses from depths of many hundreds of km.

Table 1: Values of the maximum depth D_{max} of the roof of a crustal magma reservoir located at a neutral buoyancy level from which eruptions to the surface can occur for various bodies if the magma density is 2600 kg m^{-3} and the crustal density is 2200 kg m^{-3}.

Planet	g/(m s^{-2})	D_{max} /km
Earth	9.8	2.6
Venus	8.8	2.8
Mars	3.7	6.7
Io	1.8	13.9

Table 2: Values of the pressure gradient dP/dz, the magma rise speed u and the volume flux per unit length along strike of the erupting fissure q for the eruption of a magma with η = 0.3 Pa s from a range of depths D less than the maximum depth D_{max} = 13.9 km on Io when the magma density is 2600 kg m^{-3} and the crustal density is 2200 kg m^{-3}.

D/km	W/m	dP/dz/(Pa/m)	u/(m/s)	q/(m^3 s^{-1} m^{-1})
1.0	0.6	9280.0	20.5	12.1
2.0	1.2	4280.0	19.7	23.2
3.0	1.8	2613.3	18.8	33.3
4.0	2.4	1780.0	18.0	42.3
5.0	2.9	1280.0	17.0	50.2
6.0	3.5	946.7	16.0	56.7
7.0	4.1	708.6	15.0	61.8
8.0	4.7	530.0	13.9	65.3
9.0	5.3	391.1	12.6	67.0
10.0	5.9	280.0	11.3	66.4
11.0	6.5	189.1	9.7	62.9
12.0	7.1	113.3	7.9	55.5
13.0	7.7	49.2	5.4	41.2

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