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Magma budgets and steady-state activity of Vulcano and Stromboli

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Abstract. We present three models for magma budget during steady-state volcanic activity, by which non-erupted magma is emplaced as dykes or cumulates or crystallises in place. Using gas and thermal data we apply our models at Vulcano to calculate degassing, cooling and crystallisation of magma at a rate of 40-375 kg s^{-1} within a magma body with an upper surface at a ~2 km depth. At Stromboli we calculate a steady magma supply of 300-1300 kg s^{-1} to shallow (<1 km) depths.

Introduction

At many volcanoes active magma systems drive persistent, steady state, activity at open conduits, fumarole fields or crater lakes between major eruptions. Activity is characterised by continuous gas and heat emission. Understanding and monitoring such systems and their fluxes are crucial for hazard assessment because divergence from steady-state may indicate higher eruption potential. Enhanced gas and heat fluxes occurred prior to eruptions at Etna [Malinconico, 1987], Vulcano [Chiodini et al., 1995], Pelée [Chrétien & Brousse, 1989] and Ruapehu [Ruapehu Surveillance Group, 1996]. Here we define steady-state magmatic systems in terms of three magma budget models, which we apply to Vulcano and Stromboli (Italy). Persistent steady state activity currently occur at Vulcano and Stromboli. Since Vulcano's last eruption (1888-90) activity has been characterised by continuous gas discharge from fumaroles with 80-700 °C vent temperatures. Stromboli's activity is characterised by continuous gas discharge and frequent explosions, this has persisted for at least 2000 years.

Estimation of thermal and mass flux

For volcanoes characterised by persistent non-eruptive degassing, total thermal flux (Q_{ne}) will be the sum of radiative and convective heat losses from the active vents (Q_{rad} and Q_{conv}), conductive heat loss through conduit or chamber walls (Q_{cond}) and heat carried by the gas phase (Q_{negas}). At intermittently erupting systems, such as Stromboli, heat will also be carried by gas and ejecta expelled during eruptions (Q_{egas} and Q_{ejecta}). Thermal flux during such eruptions (Q_e) will equal $Q_{egas} + Q_{ejecta}$. These parameters can be estimated using the formulae given in Table 1. Mass flux of magma through the system (M) is then calculated from $M = Q_{tot} / (c_L \Delta f + c_p \Delta T)$, where $Q_{tot} = Q_{ne} + Q_e$ and c_L, Δf, c_p and ΔT are latent heat of crystallisation, crystallised mass fraction, specific heat capacity of the magma, and the temperature through which magma cools before removal from the system.

At Vulcano we measured vent temperatures during September 1995 at 523 fumaroles within the Vulcano Fossa crater using infrared thermometers and thermocouples. These gave a vent temperature maximum of 531 °C, with a mean and standard deviation of 196 and 88 °C. Using a total station sited on the crater rim we located each fumarole measurement and mapped the fumarole field (Fig. 1). This defined a 1.63×10^4 m^2 field, across which fumaroles typically occupied 2.5 % (410 m^2). Total gas fluxes of 5-15 and 15-45 kg s^{-1} were derived respectively from Italiano et al. [1994] and COSPEC measurements [Bruno et al., 1995] in tandem with gas geochemistry [Chiodini et al., 1994]. Using these we estimate magma depth by applying Stevenson's [1993] porous pipe model for fumarole field gas flow. To estimate depth, we divide the field into 6 (N) porous conduits of radius (r_o) 5 m (from the total exhalative area, A_{exhal}: $r_o = \sqrt{A_{exhal}/[N\pi]}$), which cool to ambient temperature at a radius (r_∞) of 25 m (based on the total area >100 °C, A_{field}: $r_\infty = \sqrt{A_{field}/[N\pi]}$). This yields magma depths of 0.2-4 km, with higher gas fluxes indicating deeper sources. We consider the higher (COSPEC) flux more reliable, suggesting that the ~4 km depth is more accurate. However, heat lost by interaction with the hydrothermal system [e.g. Bolognesi & D'Amore, 1993; Chiodini et al., 1995; Tedesco et al., 1995] is not included in our calculations. Thus, we regard 4 km as a maximum depth for the magma body upper surface. Borehole data suggest magma is a few hundred metres deeper than 2 km [Faraone et al., 1986]. Therefore the magma body upper surface cannot be shallower than 2 km. Our evidence suggests that the upper surface is 2-4 km deep.

We used these data to calculate Q_{tot} (Tab. 1) and constrain the magmatic system which drives persistent activity at the fumarole field. Three models were used to estimate magma flux (M) through a closed magma system: dyke, cumulate, and stagnant models (Fig. 2). In the dyke and cumulate models buoyant, volatile rich magma rises from a deep source (Stevenson & Blake, submitted). Above the saturation depth gas is exsolved and lost to the surface. Denser degassed magma sinks back down the conduit. Convection cools the system leading to the formation of dykes or cumulates. In the dyke model, magma is assumed to release its gas and crystallise 45 % before emplacement. In the cumulate model, convecting magma releases gas and heat through the crater, degassed magma sinks to form cumulates; all of remaining latent heat is

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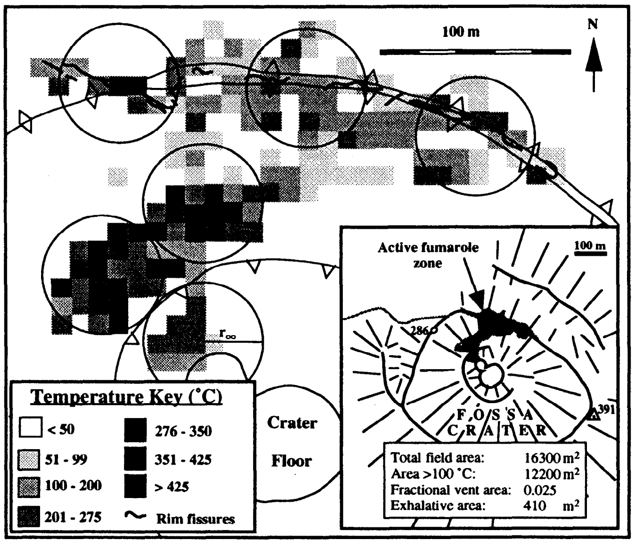

Figure 1. Fumarole temperature distribution at the Vulcano Fossa fumarole field, mapped during September 1995. Teeth mark breaks of slope: teeth point down-slope. Temperatures refer to the maximum obtained in each 100 m^2 cell. Division of the field into the 6 porous conduits (radius r_∞) used in the fumarolic flow model to calculate magma depth are indicated.

released to the reservoir. In the stagnant model, magma is intruded into the country rock. The intrusion remains stagnant and fully crystallises, losing heat through the active crater.

The COSPEC derived gas flux (30±15 kg s^{-1}) gives magma fluxes of 250 and 180 kg s^{-1} (dyke and cumulate models), or magma crystallisation at a rate of 85 kg s^{-1} (stagnant model). This is equivalent 0.3 km^3 of dykes, 0.2 km^3 of cumulates, or crystallisation of 0.1 km^3 since 1890. Our estimates have errors of ±50 % due to gas flux uncertainty.

At Stromboli during 1995 we define two eruption types: pyroclast and gas ejections to 50-300 m, and incandescent gas flares to 50 m. Udine University summit seismic station data (1992-96) give a mean rate for both types of 8.8 per hour [R. Carniel, pers com]. We obtained a maximum vent temperature of 940 °C using infrared thermometers. Triangulation gave a vent diameter of 4 m. Using these in Stevenson's [1993] rough pipe model with a 150±90 kg s^{-1} average gas flux [Allard et al., 1994], we calculate a magma surface depth of 600±350 m

These parameters constrain the magmatic system which drives the persistent activity. If a system displays persistent magmatic activity at the surface, the stagnant model can be rejected. At Stromboli, magma at a shallow (<1 km) level is connected to the surface by a linked open conduit system [Fig. 3; Harris et al., 1996]. We assume that the 1975 and 1985-86 flank eruption vents located ~110 m below the active craters [Capaldi et al., 1978; Rosi & Sbrana, 1988; Nappi & Renzulli, 1989] were fed from this shallow source. The shallow source is linked to a 10-15 km deep reservoir, inferred from seismic and geochemical data [Capaldi et al., 1978; Francalanci et al., 1989]. The shallow source is constantly replenished by buoyant volatile rich magma rising from the deep reservoir, displacing degassed magma downwards. Rising gases form foam at the shallow source roof. Foam collapse feeds gas slugs which drive explosive activity [Jaupart & Vergniolle, 1988]. Non-erupted degassed magma sinks to be emplaced as dykes or cumulates. This requires convection in the feeder conduit [Fig. 3; Kazahaya et al., 1994; Stevenson & Blake, submitted].

Stromboli's Q_{ne} is ~10 times Q_e (Tab. 1). To maintain Q_{nc}, 860 kg s^{-1} of magma must be emplaced as dykes in or beneath the edifice, equivalent to injection of 22 km^3 in 2000 years. Alternatively 640 kg s^{-1} of cumulates must be formed,

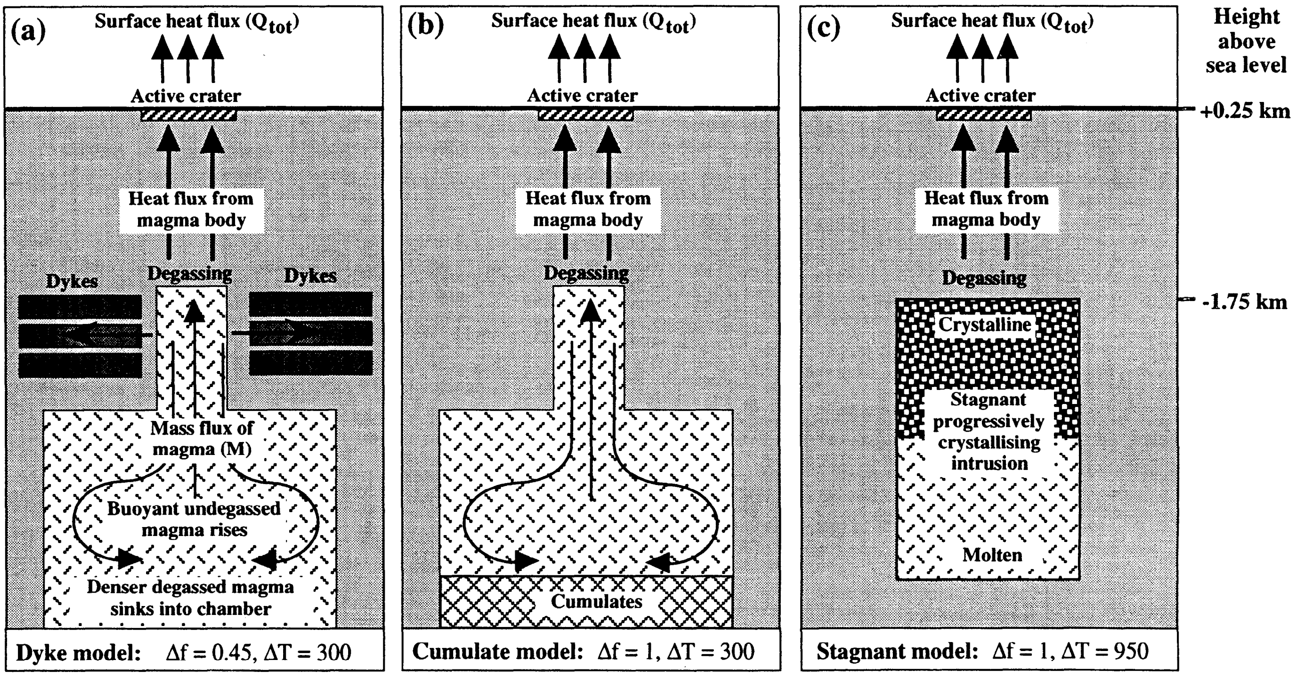

Figure 2. Schematic models for steady-state degassing magma systems, giving the crystallised mass fraction (Δf) and temperature range through which magma cools (ΔT) used to calculate magma flux (M). Non-erupted degassed magma is either emplaced as (a) dykes, (b) cumulates, or (c) cools and crystallises in place.

Table 1. Heat flux at Vulcano and Stromboli. At Vulcano T_{exit}, r_o and r_∞ are estimated using our 1995 fumarole field measurements. T_{exit} was set to the mean temperature for all fumaroles within each 100 m^2 cell defined in Fig. 1 and πr_o^2 was substituted with total vent area for all fumaroles within the cell. Fluxes derived for each cell were summed. For Stromboli T_{exit} and r_o were taken from Harris and Stevenson [in press] and r_∞ was set at the maximum distance to fumaroles >100 °C (~120 m). D was calculated using the Stevenson [1993] model. For Stromboli M_e of 6 kg s^{-1} was calculated from data for explosive and effusive eruptions over the last 100 years given by Chouet et al. [1974], Blackburn et al. [1976], Capaldi et al. [1978], Rosi and Sbrana [1988], Nappi and Renzulli [1989], Barberi et al. [1993], Napoleone et al. [1993] and Ripepe et al. [1993]. Gas and water fluxes for Vulcano were taken from Bruno et al. [1994], Chiodini et al. [1994] and Italiano et al. [1994] and for Stromboli were calculated using Allard et al. [1994]. k, α, p, μ and κ were taken from Kays & Crawford [1980], ε, k_w, c_g, L_v, c_p and c_L were set to 0.9887, 3 W m^{-1} K^{-1}, 1600 J kg^{-1} K^{-1}, 2.26 x 10^6 J kg^{-1}, 1150 J kg^{-1} K^{-1} and 3 × 10^5 J kg^{-1} respectively. Magmatic and eruption temperatures of 1000 °C, the near solidus temperature for magma at Vulcano and Stromboli, were used with T_{air} and T_{amb} of 0 °C.

Mode of heat loss	Vulcano (MW)	Stromboli (MW)	Derivation
Q_{rad}	1	2	$Q_{rad} = \pi r_o^2 \sigma \varepsilon T_{exit}^4$ (r_o = conduit radius, σ = Stefan-Boltzmann constant, ε = Emissivity, T_{exit}= conduit exit or vent temperature)
Q_{conv}	1	1	$Q_{conv} = 0.14\, \pi r_o^2\, k\, (g\alpha p/\mu\kappa)^{1/3}\, (T_{exit} - T_{air})$ (T_{air} = ambient air temperature, g = acceleration due to gravity, k, α, p, μ and κ are thermal conductivity, density, cubic expansivity, dynamic viscosity and thermal diffusivity for gas at a temperature of $[T_{air}+T_{exit}]/2$)
Q_{cond}	24±10	11±6	$Q_{cond} = [2\pi k_w D(T_{gas} - T_{amb})] / \ln(r_\infty/r_o)$ (k_w = wall rock thermal conductivity, D = magma depth, T_{gas} = gas conduit temperature [approximated from $\{T_m+T_{exit}\}/2$, T_m = magma surface temperature], T_{amb} = ambient surface temperature, r_∞ = distance to T_{amb})
Q_{negas}	92±46	400±200	$Q_{negas} = Fc_g\Delta T_g + F_wL_v$ (F = non-eruptive gas flux, c_g = gas specific heat capacity, ΔT_g = gas cooling from T_m to T_{air}, F_w = water flux, L_v = latent heat of condensation)
Q_{egas}	0	1	$Q_{egas} = F_ec_g\Delta T_g + F_{we}L_v$ (F_e = erupted gas flux, F_{we} = erupted water flux)
Q_{ejecta}	0	8	$Q_{ejecta} = (c_p\Delta T_e + c_L)\, M_e$ (c_p = specific heat capacity of ejecta, ΔT_e = ejecta cooling from eruption to T_{amb}, c_L = latent heat of crystallisation, M_e = mass flux of erupted material)
Q_{tot}	118±56	423±226	

equivalent to 16 km^3 in 2000 years. Gas flux uncertainty cause ±55 % error. Such endogenous growth has been postulated by Francis et al. [1993] and our estimate of 0.004-0.016 km^3 yr^{-1} agrees with that based on sulfur budgets [Allard et al, 1994].

Implications and conclusions

Using Vulcano and Stromboli we have shown how thermal and gas fluxes can constrain steady-state magma systems. At Vulcano a 2-4 km deep closed system fuels fumarolic activity. At Stromboli a convecting open-system, with a 600±350 m deep upper surface, drives explosive activity. Both have characteristic gas (30±15 and 150±90 kg s^{-1}), heat (120±60 and 420±230 MW) and mass (40-375 and 300-1300 kg s^{-1}) fluxes. Simultaneous gas chemistry and flux measurements may have improved our ability to distinguish which model is applicable and more tightly constrain magma flux and depth.

Determination of steady-state is essential in understanding an active volcano and eruption hazard. While typical steady-state fluxes are maintained, eruption potential will be low. Divergence from steady-state implies changes in the system, and perhaps increased eruption potential. At Vulcano, increased fluxes would result from magma ascent into shallow, wet environments or water descent towards the magma body [Chiodini et al., 1995]. During such events risk of phreato-magmatic eruption will be high. Although gas and thermal flux increases have occurred without eruption since 1890 [Chiodini et al., 1995], it is apparent from reports quoted in Keller [1980] and Chiodini et al. [1995] that the 1888-90 eruption was preceded by increased gas and thermal flux. At Stromboli, increased gas and mass flux will increase the potential for violent explosions and effusive activity, both of which interrupt steady-state activity at Stromboli [Barberi et al., 1993]. We infer increased gas and magma flux prior to the 1975 effusive eruption and 1985-86 major explosive and effusive events from reports of increased fumarolic and explosive activity before both events [Capaldi et al., 1978; Rosi & Sbrana, 1988; Nappi & Renzulli, 1989]. Equally divergence from steady-state will result from decreased fluxes. This does not necessarily imply decreased eruption potential: decreased fluxes may indicate blockage, which by causing gas build-up, may increase eruption potential. At Stromboli some violent explosions have resulted from gas pressure increase following vent blockage [Barberi et al., 1993].

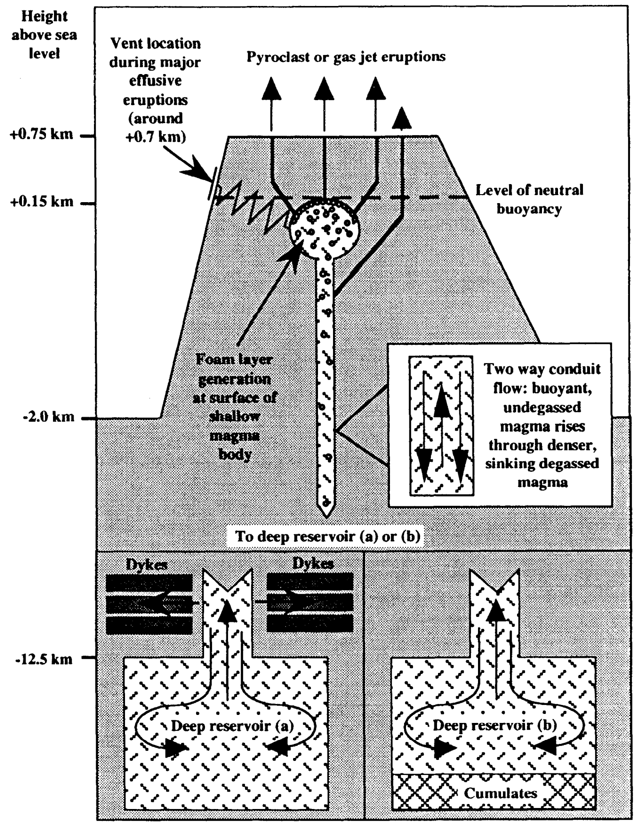

Figure 3. Schematic model of Stromboli's magma system.

Our steady-state models have been kept simple to allow application to similar systems elsewhere (e.g. Kudriavy, Nisyros and Etna); to crater lakes (e.g. Poás and Ruapehu) or degassing conduits (e.g. Masaya). Determination of the steady-state system and its typical gas, thermal and magma fluxes will allow the hazard posed by divergence to be judged.

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