Conduit convection driving persistent degassing at basaltic volcanoes

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ABSTRACT

The persistent release of gas at basaltic volcanoes where there is a low magma eruption rate can be driven by an exchange flow of magma in the conduit, in which gas-rich magma ascends, degasses and crystallises and then sinks back down the conduit. The driving force of the flow is provided by the density difference between the buoyant bubble-rich magma at depth and the dense degassed crystallised magma at shallow levels. In this study we attempt to constrain the physical and chemical processes driving an exchange flow of magma at Stromboli, Aeolian Archipelago, Italy. The model uses a simple, cylindrical geometry. We define degassing and crystallisation paths of the ascending and descending magmas, constrained by gas flux and melt inclusion data given in the literature, to produce a three-phase model of ascending and descending magmas driving persistent gas fluxes. We calculate the viscosity of the three-phase magma using end-member rheological models for bubble and crystal suspensions. Combining our modelled magma properties with analogue exchange flow experiments we can relate the regime of magma flow driving persistent degassing to pressure. At pressures ≤ 90 MPa (≤ 3 km) the viscosity ratio is ≤ 100 and the regime is predicted to be side by side flow with both ascending and descending magmas adjacent to a portion of the conduit wall. At pressures ≥ 90 MPa (≥ 3 km) the viscosity ratio between the ascending and descending magma is ≥ 100 and the flow is predicted to be core annular flow, with the ascending vesiculating magma in the inner core and the more crystalline degassed magma flowing down along the conduit wall. By analogy, we hypothesise that degassed magma would flow down along the walls in dike-like plumbing geometries that have been proposed for the deeper Strombolian system. Analogue experiments suggest that exchange flows do not overturn under conditions of maximum volume flux; we use an empirical relationship to characterise the volume flux of the exchange flow and show that the radius of a cylindrical conduit required to account for the observed persistent gas flux at Stromboli at 0.1 MPa is ~ 1 m.

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1. Introduction

Persistent gas fluxes at many active basaltic volcanoes such as Etna (Sicily, Italy), Stromboli (Aeolian Archipelago, Italy), Izu Oshima (Japan) and Masaya (Nicaragua) cannot be reconciled with the exclusive degassing of magma erupted during effusive and explosive episodes. At Etna, Allard (1997) estimates that ~ 80 – 90% of gas emissions are generated by magma that is never erupted. At Stromboli $\sim 50,000$ t d^{−1} of magma must be degassed to account for the persistent gas flux of ~ 200 t d^{−1} sulphur dioxide (SO₂) (Allard et al., 2008). Normal Strombolian explosive activity and effusive eruptions emit on average only ~ 530 t d^{−1} of solid matter (Allard et al., 1994) and thus the extruded magma flux represents $\sim 1\%$ of the magma processed during persistent degassing. To account for the continuous supply of gas and heat recorded, where there is no significant extrusion of magma, an exchange flow of magma within the conduit is inferred, driven by the

density difference between dense degassed magma at shallow levels and buoyant gas-rich magma deeper within the system (Kazahaya et al., 1994; Harris and Stevenson, 1997; Stevenson and Blake, 1998). To interpret gas emissions and understand changes in activity at persistently active basaltic volcanoes therefore require some knowledge of the flow regime of the exchange flow, and how its volume flux relates to the degassing regimes observed and the physical properties of the magma: its viscosity and density and thus crystallinity, vesicularity and the composition of its melt.

Experimental studies of exchange flow in vertical cylindrical pipes, where the volume flux of the ascending fluid is matched by an equal flux of descending fluid, have shown that the flow regime of an exchange flow is a function of the viscosity ratio $\beta = \mu_2/\mu_1$, the viscosity of the more viscous fluid divided by the viscosity of the less viscous fluid (Stevenson and Blake, 1998; Huppert and Hallworth, 2007; Beckett et al., 2011). In this study we explore how to apply the experimental observations to a persistently degassing basaltic volcano. We use Stromboli as a case study as there are extensive data sets on the chemical composition of its magma and well constrained measurements of the flux and composition of the gas released. However,

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Fig. 1. Cartoon to show the flow types of buoyancy driven, laminar, exchange flow in a vertical pipe, a. Core annular flow (CAF), b. Side by side (SBS) flow and c. Transitional side by side (TSBS) flow. Arrows indicate the dominant flow direction, however at the fluid interface the less viscous fluid flows downwards, in the same direction as the more viscous fluid, as indicated by the velocity profiles for CAF and SBS flow.

application of the experimental observations requires that we know the density and viscosity of the magma as a function of pressure during both ascent and decent. Using melt inclusion (MI) data given by Métrich et al. (2010) and measured gas flux data (Burton et al., 2009), we use VolatileCalc (Newman and Lowenstern, 2002) and MELTS (Ghiorso and Sack, 1995; Asimow and Ghiorso, 1998) to constrain the crystallisation and degassing path of ascending and descending magma as a function of pressure. We then apply new methods to characterise the multiphase rheology of the ascending and descending magma, accounting for its crystallinity, vesicularity and melt composition. To calculate the rheology of the three-phase system we explore two approaches: in the first the viscosity of the bubble suspension is calculated taking the crystals plus melt as the suspending fluid; in the second the viscosity of a crystal suspension is calculated taking the bubbles plus melt as the suspending fluid. It is shown that there is no significant

difference in the calculated viscosities from the two methods. Using the calculated viscosity and density of the magma as a function of pressure, we determine the Reynolds number of an exchange flow at Stromboli and show that the flow is laminar, and thus we can apply experimental observations of low Reynolds number exchange flows. We calculate the viscosity ratio between the ascending and descending magma as a function of pressure and define the flow regime of an exchange flow at Stromboli. Although the degassing behaviour at Stromboli and the properties of its magma are well constrained, the geometry of its feeder system remains unclear and likely includes dyke-like structures (Chouet et al., 2003). All experimental studies of exchange flow have been conducted in cylindrical pipes and in order to apply the experimental results we assume a single cylindrical conduit, the most thermally efficient and mechanically robust geometry for a steady-state system. This conduit geometry is an over-simplification and we discuss the limitations of this approach and the implications for defining the flow regime of an exchange flow of magma at persistently degassing basaltic volcanoes.

2. Background

There have been a number of theoretical, numerical and experimental studies that have attempted to better understand how a density driven exchange flow can drive persistent gas fluxes at basaltic volcanoes (Kazahaya et al., 1994; Stevenson and Blake, 1998; Huppert and Hallworth, 2007; M.R. Burton et al., 2007; Beckett et al., 2011; Kerswell, 2011; Palma et al., 2011; Molina et al., 2012). Stevenson and Blake (1998) presented analogue experiments of simple transient exchange flows in which the flow regime in a vertical cylindrical pipe was found to be core annular flow (CAF), with the less viscous fluid ascending up the centre of the pipe and the more viscous fluid descending down the walls in an outer annulus. Imposing a CAF geometry they present a theoretical model to constrain the conduit radius required to drive gas fluxes by assuming that the overturn rate of magma is proportional to the gas flux. Dimensional analysis shows that in a cylindrical conduit magma exchange flux is proportional to the conduit radius to the fourth power, and thus the radius of the conduit is highly insensitive to the volume flux of the exchange flow. The properties of the exchanging fluids are estimated assuming that one represents magma in a chamber at a depth of 2 km and the other a shallow degassed magma. They take typical properties for basaltic magma and assume that properties are independent of pressure for each fluid and so do not account for degassing and crystallisation paths, such that ascending magma has a water content of 0.5 wt.% and a crystallinity of 30 vol.%. They

Fig. 2. Saturation pressures of MI found in olivine in the pumice and scoria ejected at Stromboli given by Métrich et al. (2010) calculated using VolatileCalc (Newman and Lowenstern, 2002) assuming closed-system degassing. Simple linear fits allow us to define the degassing path, such that a. $X_{H_2O} = 0.0105 \cdot P$ and b. $X_{CO_2} = 0.0005 \cdot P$.

conclude that to account for a water mass flux of $\sim 70 \text{ kg s}^{-1}$, calculated from the SO_2 flux and weighted ratios of H_2O , CO_2 , HCl and HF to SO_2 from analysis of both fumarolic and quiescent plume emissions given by Allard et al. (1994), the radius of the conduit at Stromboli has to be 4–10 m.

M.R. Burton et al. (2007) provide a more quantitative determination of the density difference driving overturn at Stromboli. They use constraints from the literature on the petrology and texture of erupted products as well as SO_2 flux data to define the degassing path of ascending magma as a function of pressure, but do not define its crystallisation path. They show that simple closed-system degassing of primitive basalt at Stromboli results in gas volume fractions of over 99.9%, which is incompatible with the vesicularity (ϕ_b) of scoria ejected during normal Strombolian eruptions, $\phi_b = 0.3\text{--}0.8$ (Polacci et al., 2009). They assume that at a vesicularity of 50 vol.%, which they find to occur at 50–100 MPa, there is a transition from closed- to open-system degassing where the magma becomes permeable to gas flow with the gas then rising separately from the magma. They present a conduit flow model to calculate the velocities of the ascending and descending magmas using density differences from their calculated density profiles. However, they use end-member values for the viscosities of the ascending and descending magmas ($\mu_{1,2}$) given by Métrich et al. (2001): $\mu_1 \sim 20\text{--}30 \text{ Pa s}$ for an original un-crystallised magma and $\mu_2 \sim 14,000 \text{ Pa s}$ for a degassed, 50 vol.% crystallised magma. They assume a CAF geometry and that the ascending magma occupies a radius such as to maximise the exchange flux, an assumption shown to be invalid by analogue exchange flow experiments conducted by Beckett et al. (2011). They conclude that the conduit radius required to account for the mass flow rate of ascending magma (constrained by gas flux data) is 1.28 m at 50 MPa, which increases to 1.45 m at 200 MPa in order to satisfy mass conservation.

Recent experiments by Huppert and Hallworth (2007), Beckett et al. (2011) and (Palma et al., 2011) have included fluid reservoirs at the ends of the pipes which allow longer exchange flow durations, representative of steady state conditions. Huppert and Hallworth (2007) performed experiments with $10^{-4} < Re_2 < 10^1$, where Re_2 is the Reynolds number of the more viscous fluid calculated given the volume flux of the flow, and β ranging from 10^0 to 10^5 . They state that the predominant behaviour is CAF, with the less viscous fluid occupying the core, but also note that 'the fluid may be highly irregular and break into discrete blobs or partially adhere to the side wall of the conduit'. Palma et al. (2011) conducted experiments in vertical and inclined cylindrical pipes, at viscosity ratios (β) from 3.38 to 1.83×10^3 , and Reynolds numbers $10^{-4}\text{--}10^1$. In pipes inclined at 6° and 20° to vertical, they observed a stratified flow pattern, where the dense fluid is in contact with the lower side of the pipe and the lighter fluid is in contact with the upper side of the pipe, noting that exchange flows at the lower inclination appeared unstable. In vertical pipes, they observed CAF and 'pseudo-stratified' flow (although noting that these flow patterns were unstable), turbulent flow and 'blobs of lighter fluid rising through the continuous dense fluid'. They defined 'pseudo-stratified' flow as a similar flow pattern to stratified flow in inclined pipes. Huppert and Hallworth (2007) and Palma et al. (2011) analysed the flux of their exchange flows in terms of a 'flux coefficient' (the ratio of Reynolds to Grashof numbers), and Palma et al. (2011) used experimental results to identify flux coefficients for vertical and inclined cylindrical pipe flows. They also proposed that the same analysis could be applied to flows in multiple cylindrical or slot-shaped conduits through the use of an 'equivalent radius', but did not present experimental results to identify flux coefficient values corresponding to these configurations.

Palma et al. (2011) applied their experimental results to degassing at Villarica volcano, using the magma properties and the melt viscosity model of Hui and Zhang (2007), and the Einstein–Roscoe equation to account for the effect of crystals on the magma viscosity. The volumetric exchange flow rate was obtained from the magma properties, measurements of sulphur outgassing and the initial sulphur content of melt inclusions, and the experimentally-determined flux coefficients. A range

Fig. 3. The crystallisation paths of ascending and descending magma at Stromboli, modelled using MELTS (Ghiorso and Sack, 1995; Asimow and Ghiorso, 1998) with the bulk composition of pumice ST82 (Métrich et al., 2010). The wt.% of H_2O is inputted into the model at each pressure using the $X_{\text{H}_2\text{O}} - P$ fit given in Fig. 2. Crystallisation paths of descending magma are shown assuming that magma ascends to 0.1 MPa, 40 MPa and 75 MPa in the system.

of different initial volatile contents and magma properties was used in the calculations, but the properties of the rising and degassed magma were taken to be constant during transport. Palma et al. (2011) identified volatile content and magma property ranges that produced outgassing fluxes comparable to observations at Villarica.

Fig. 4. Volume percent of plagioclase (PLAG), clinopyroxene (CPX) and olivine (OL) in the ascending magma at Stromboli determined from MELTS (Ghiorso and Sack, 1995; Asimow and Ghiorso, 1998) using the bulk composition of pumice ST82 (Métrich et al., 2010), calculated at $T = 1130^\circ \text{C}$ and f_{O_2} fixed at $\Delta \text{NNO} + 1$. The average crystal content of scoria ejected at Stromboli, after Métrich et al. (2001), Francalanci et al. (2004), Landi et al. (2004), and Métrich et al. (2010), are shown for comparison, indicated by the grey box.

Table 1
Computed melt composition (wt.%) and crystallisation path (vol.% crystals in the melt) of ascending Stromboli magma as a function of pressure, using MELTS with a bulk composition of pumice ST82 as given by Métrich et al. (2010) for T = 1130 °C.

Bulk composition												
Input H ₂ O	3.68	3.62	3.57	3.52	3.47	3.41	3.36	3.31	3.26	3.20	3.15	3.10
P (MPa)	350	345	340	335	330	325	320	315	310	305	300	295
SiO ₂	50.9	49.49	49.52	49.54	49.57	49.60	49.63	49.65	49.68	49.71	49.73	49.79
TiO ₂	0.9	0.89	0.89	0.89	0.89	0.89	0.89	0.89	0.89	0.89	0.89	0.89
Al ₂ O ₃	16.6	16.88	16.90	16.91	16.93	16.95	16.97	16.99	17.00	17.02	17.04	17.08
Fe ₂ O ₃		2.34	2.34	2.35	2.35	2.36	2.36	2.37	2.37	2.38	2.38	2.39
FeO	8.22	5.92	5.93	5.92	5.92	5.92	5.92	5.92	5.92	5.92	5.92	5.92
MnO	0.16	0.17	0.17	0.17	0.17	0.17	0.17	0.17	0.17	0.17	0.17	0.17
MgO	6.66	5.80	5.80	5.80	5.80	5.80	5.80	5.80	5.80	5.80	5.80	5.80
CaO	10.9	9.94	9.93	9.93	9.92	9.91	9.90	9.90	9.89	9.88	9.87	9.85
Na ₂ O	2.39	2.46	2.47	2.47	2.48	2.48	2.48	2.49	2.49	2.49	2.50	2.50
K ₂ O	1.82	1.88	1.89	1.89	1.89	1.90	1.90	1.90	1.90	1.91	1.91	1.92
P ₂ O ₅	0.4	0.41	0.42	0.42	0.42	0.42	0.42	0.42	0.42	0.42	0.42	0.42
H ₂ O		3.81	3.76	3.71	3.66	3.61	3.56	3.51	3.46	3.41	3.36	3.31
CaO/Al ₂ O ₃		0.59	0.59	0.59	0.59	0.59	0.58	0.58	0.58	0.58	0.58	0.58
Olivine vol.%		0.19	0.17	0.16	0.15	0.14	0.13	0.12	0.11	0.10	0.08	0.07
Cpx vol.%		4.75	4.87	4.92	5.00	5.08	5.18	5.24	5.33	5.42	5.51	5.60
Plag vol.%		0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Vol.% crystals		4.94	5.04	5.09	5.15	5.22	5.31	5.36	5.43	5.51	5.59	5.68
Fo%		85.90	85.90	85.90	85.90	85.90	85.90	85.90	85.90	85.90	85.90	85.90
An%												

Bulk composition												
Input H ₂ O	3.05	2.99	2.94	2.89	2.84	2.78	2.73	2.68	2.63	2.57	2.52	2.47
P (MPa)	290	285	280	275	270	265	260	255	250	245	240	235
SiO ₂	50.9	49.81	49.84	49.87	49.90	49.93	49.96	49.99	50.02	50.05	50.08	50.15
TiO ₂	0.9	0.89	0.90	0.90	0.90	0.90	0.90	0.90	0.90	0.90	0.90	0.90
Al ₂ O ₃	16.6	17.10	17.12	17.15	17.17	17.19	17.21	17.24	17.26	17.28	17.31	17.37
Fe ₂ O ₃		2.40	2.40	2.41	2.41	2.42	2.42	2.43	2.43	2.44	2.44	2.45
FeO	8.22	5.92	5.92	5.92	5.92	5.92	5.92	5.92	5.92	5.92	5.92	5.91
MnO	0.16	0.17	0.17	0.17	0.17	0.17	0.17	0.17	0.17	0.17	0.17	0.17
MgO	6.66	5.80	5.79	5.79	5.79	5.79	5.78	5.77	5.76	5.76	5.74	5.72
CaO	10.9	9.84	9.83	9.82	9.81	9.79	9.78	9.77	9.76	9.76	9.74	9.71
Na ₂ O	2.39	2.51	2.51	2.52	2.52	2.52	2.53	2.53	2.54	2.54	2.55	2.56
K ₂ O	1.82	1.92	1.92	1.93	1.93	1.93	1.94	1.94	1.94	1.95	1.95	1.96
P ₂ O ₅	0.4	0.42	0.42	0.42	0.42	0.42	0.43	0.43	0.43	0.43	0.43	0.43
H ₂ O		3.21	3.16	3.11	3.06	3.01	2.96	2.91	2.86	2.81	2.76	2.71
CaO/Al ₂ O ₃		0.58	0.57	0.57	0.57	0.57	0.57	0.57	0.57	0.57	0.56	0.56
Olivine vol.%		0.05	0.04	0.02	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Cpx vol.%		5.79	5.90	6.01	6.10	6.23	6.34	6.45	6.56	6.64	6.79	7.04
Plag vol.%		0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Vol.% crystals		5.84	5.94	6.03	6.12	6.23	6.34	6.45	6.56	6.64	6.79	7.04
Fo%		85.90	85.90	85.90	85.00	85.00						
An%												

Bulk composition												
Input H ₂ O	2.42	2.36	2.31	2.26	2.21	2.15	2.10	2.00	1.89	1.79	1.68	1.58
P (MPa)	230	225	220	215	210	205	200	190	180	170	160	150
SiO ₂	50.9	50.18	50.21	50.24	50.28	50.31	50.35	50.38	50.45	50.52	50.60	50.75
TiO ₂	0.9	0.90	0.91	0.91	0.91	0.91	0.91	0.91	0.91	0.92	0.92	0.92
Al ₂ O ₃	16.6	17.39	17.42	17.45	17.48	17.51	17.54	17.57	17.63	17.70	17.77	17.84
Fe ₂ O ₃		2.46	2.46	2.47	2.47	2.48	2.48	2.48	2.49	2.50	2.51	2.52
FeO	8.22	5.91	5.91	5.91	5.91	5.90	5.90	5.90	5.89	5.89	5.88	5.87
MnO	0.16	0.17	0.17	0.17	0.17	0.17	0.17	0.18	0.18	0.18	0.18	0.18
MgO	6.66	5.71	5.70	5.69	5.68	5.67	5.65	5.64	5.61	5.59	5.55	5.49
CaO	10.9	9.70	9.69	9.67	9.57	9.64	9.62	9.61	9.57	9.53	9.49	9.40
Na ₂ O	2.39	2.56	2.57	2.57	2.58	2.59	2.59	2.60	2.61	2.62	2.64	2.67
K ₂ O	1.82	1.96	1.97	1.97	1.98	1.98	1.99	1.99	2.00	2.02	2.03	2.05
P ₂ O ₅	0.4	0.43	0.43	0.43	0.43	0.44	0.44	0.44	0.44	0.44	0.45	0.45
H ₂ O		2.61	2.56	2.51	2.45	2.40	2.35	2.30	2.20	2.09	1.99	1.88
CaO/Al ₂ O ₃		0.56	0.56	0.55	0.55	0.55	0.55	0.55	0.54	0.54	0.53	0.53
Olivine vol.%		0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Cpx vol.%		7.16	7.30	7.44	7.58	7.72	7.87	8.03	8.35	8.69	9.05	9.43
Plag vol.%		0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Vol.% crystals		7.16	7.30	7.44	7.58	7.72	7.87	8.03	8.35	8.69	9.05	9.43
Fo%												
An%												

Bulk composition												
Input H ₂ O	1.47	1.37	1.26	1.16	1.05	0.95	0.84	0.79	0.74	0.63	0.53	0.42
P (MPa)	140	130	120	110	100	90	80	75	70	60	50	40
SiO ₂	50.9	50.84	50.93	51.02	51.11	51.21	51.32	51.45	51.52	51.62	51.84	52.29
TiO ₂	0.9	0.93	0.93	0.93	0.93	0.94	0.95	0.97	0.99	1.00	1.04	1.13
Al ₂ O ₃	16.6	18.00	18.08	18.17	18.26	18.35	18.33	18.13	18.03	17.89	17.61	17.02

Table 1 (continued)

Bulk composition													
Fe ₂ O ₃		2.53	2.53	2.54	2.54	2.55	2.57	2.63	2.66	2.69	2.76	2.84	2.92
FeO	8.22	5.86	5.85	5.84	5.83	5.81	5.85	5.95	6.01	6.07	6.19	6.31	6.46
MnO	0.16	0.18	0.18	0.18	0.19	0.19	0.19	0.20	0.20	0.20	0.20	0.21	0.22
MgO	6.66	5.45	5.41	5.37	5.32	5.27	5.26	5.30	5.32	5.28	5.21	5.13	5.04
CaO	10.9	9.34	9.28	9.22	9.16	9.09	8.98	8.82	8.75	8.69	8.56	8.43	8.28
Na ₂ O	2.39	2.69	2.70	2.72	2.74	2.76	2.79	2.81	2.83	2.84	2.87	2.90	2.92
K ₂ O	1.82	2.07	2.08	2.10	2.12	2.13	2.17	2.22	2.25	2.28	2.36	2.45	2.55
P ₂ O ₅	0.4	0.45	0.46	0.46	0.46	0.47	0.48	0.49	0.50	0.50	0.52	0.55	0.57
H ₂ O		1.67	1.56	1.45	1.34	1.23	1.12	1.03	0.98	0.93	0.82	0.72	0.60
CaO/Al ₂ O ₃		0.52	0.51	0.51	0.50	0.50	0.49	0.49	0.49	0.49	0.49	0.49	0.49
Olivine vol.%		0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.03	0.22	0.61	1.01	1.41
Cpx vol.%		10.30	10.70	11.20	11.70	12.20	12.70	13.10	13.20	13.20	13.30	13.50	13.70
Plag vol.%		0.00	0.00	0.00	0.00	0.00	0.69	2.34	3.23	4.35	6.64	9.03	11.60
Vol.% crystals		10.30	10.70	11.20	11.70	12.20	13.40	15.40	16.50	17.80	20.60	23.50	26.70
Fo%									84.00	84.00	83.80	83.00	82.80
An%							81.80	80.80	80.80	79.80	78.80	77.80	76.50

Bulk composition													
Input H ₂ O				0.21		0.11		0.05		0.02		0.00	
P (MPa)				20		10		5		2		0.1	
SiO ₂	50.9		52.76		52.98		53.09		53.15		53.19		
TiO ₂	0.9		1.25		1.34		1.40		1.44		1.46		
Al ₂ O ₃	16.6		16.36		15.99		15.79		15.67		15.59		
Fe ₂ O ₃			3.11		3.22		3.28		3.32		3.35		
FeO	8.22		6.79		7.00		7.12		7.19		7.24		
MnO	0.16		0.23		0.24		0.24		0.25		0.25		
MgO	6.66		4.82		4.68		4.60		4.55		4.51		
CaO	10.9		7.96		7.77		7.67		7.61		7.56		
Na ₂ O	2.39		2.93		2.91		2.88		2.86		2.84		
K ₂ O	1.82		2.82		3.00		3.11		3.19		3.24		
P ₂ O ₅	0.4		0.64		0.69		0.72		0.74		0.76		
H ₂ O			0.34		0.18		0.09		0.04		0.00		
CaO/Al ₂ O ₃			0.49		0.49		0.49		0.49		0.49		
Olivine vol.%			2.25		2.70		2.92		3.06		3.14		
Cpx vol.%			14.50		15.20		15.60		15.90		16.20		
Plag vol.%			17.30		20.60		22.50		23.70		24.50		
Vol.% crystals			34.00		38.50		41.00		42.70		43.80		
Fo%			81.8		80.8		80.8		79.8		79.80		
An%			72.4		70.4		69.4		68.4		68.00		

The experiments presented by Beckett et al. (2011) show that buoyancy driven, low Reynolds number ($Re < 1$) exchange flow in a vertical pipe can form two distinct flow regimes over long time durations, representative of steady state conditions (Fig. 1): i. core annular flow (CAF) – axisymmetric flow of the two fluids where the interface between them exists at some radial distance from the centre, so that one fluid occupies the central core and the other an annulus around it. The less viscous fluid always occupies the centre. ii. side by side (SBS) flow – in which the two fluids touch the pipe walls, and the interface between them is curved and contacts the wall at two vertical lines. The flow regime formed was found to depend on viscosity ratio (β), defined as the viscosity of the more viscous fluid divided by the viscosity of the less viscous fluid. CAF occurred at $\beta \geq 100$ and SBS flow at $\beta \leq 100$. A subset of the SBS flows, termed transitional side by side (TSBS) flow, were observed at $\beta = 5$ –59. In TSBS flow both CAF and SBS were observed simultaneously within different parts of the pipe; the majority of the vertical extent of the flow was SBS (and was found to increase in length with time during three experiments), but at some height in the upper portion of the pipe the ascending fluid moved off the pipe wall forming a core annular flow. Obtained velocity profiles of the CAF and SBS flows show that at the fluid interface, a portion of the less dense, less viscous fluid is dragged downwards by the more viscous fluid i.e. a portion of the less dense fluid has a negative velocity (Fig. 1). The volume flux (Q) of an exchange flow is defined as the net ascending flux of the less dense, less viscous fluid, which in the analogue experiments is constrained to be equal to the net downwards flux of the more dense, more viscous fluid. The experiments show that exchange flows do not adopt a flow regime that maximises volume flux. Instead

it is shown that the volume flux of the flow does not depend on the flow regime and is given by the following empirical relationship based on dimensional scaling for volume flux (Q):

$$Q = (0.059\beta^{-0.74}) \frac{g\Delta\rho r_2^4}{2\mu_1}, \quad (1)$$

where g is gravitational acceleration, $\Delta\rho$ is density difference, r_2 is the pipe radius and μ_1 is the viscosity of the less viscous fluid.

It is worth noting that Eq. (1) is the only dimensional form for expressing the exchange flow flux in terms of the properties of the fluids and conduit. Other studies have also identified this form (Kazahaya et al., 1994; Stevenson and Blake, 1998; Huppert and Hallworth, 2007; Kerswell, 2011; Palma et al., 2011), with different choices of the rising or descending magma viscosities and different coefficients to account for non-maximal exchange flow occurring. In particular, Palma et al. (2011), Kerswell (2011) and Beckett et al. (2011) presented detailed consideration of the system parameter space and its relationship to the condition of maximal volumetric flux of the exchange flow. The primary conclusion of these studies is that laboratory and natural vertical exchange flow of two fluids does not occur at maximal flux, and there is some relationship between the exchange flux and flow pattern adopted. Currently, there is incomplete understanding of the relationship between the exchange flux and the pressure gradients in a vertical region containing two fluids of different densities, and this restricts progress in volcanological applications to the use of empirical relationships such as Eq. (1) or flux coefficient values identified by Huppert and Hallworth (2007) and Palma et al. (2011).

Table 2
Computed melt composition (wt.%) and crystallisation path of descending (vol.% crystals in the melt) Stromboli magma as a function of pressure using MELTS with a bulk composition of pumice ST82 as given by Métrich et al. (2010) for T = 1130 °C.

Bulk composition		MELTS											
Input H ₂ O		0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011
P (MPa)		350	345	340	335	330	325	320	315	310	305	300	295
SiO ₂	50.9	53.49	53.49	53.48	53.47	53.47	53.46	53.45	53.44	53.43	53.42	53.41	53.40
TiO ₂	0.9	2.14	2.13	2.12	2.11	2.10	2.09	2.08	2.06	2.05	2.04	2.03	2.02
Al ₂ O ₃	16.6	17.99	17.96	17.94	17.91	17.88	17.85	17.83	17.80	17.77	17.74	17.71	17.68
Fe ₂ O ₃		2.13	2.15	2.18	2.20	2.22	2.24	2.26	2.28	2.30	2.32	2.34	2.36
FeO	8.22	5.04	5.08	5.13	5.17	5.21	5.25	5.30	5.34	5.38	5.43	5.47	5.52
MnO	0.16	0.64	0.63	0.63	0.62	0.61	0.61	0.60	0.60	0.59	0.58	0.58	0.57
MgO	6.66	2.33	2.35	2.37	2.40	2.42	2.44	2.47	2.49	2.52	2.54	2.57	2.59
CaO	10.9	6.34	6.34	6.34	6.34	6.34	6.35	6.35	6.35	6.36	6.36	6.36	6.37
Na ₂ O	2.39	2.57	2.59	2.60	2.62	2.63	2.65	2.66	2.68	2.69	2.71	2.72	2.73
K ₂ O	1.82	5.72	5.68	5.65	5.61	5.58	5.54	5.03	5.47	5.43	5.39	5.36	5.32
P ₂ O ₅	0.4	1.60	1.58	1.57	1.55	1.54	1.52	1.50	1.49	1.47	1.46	1.44	1.43
H ₂ O		0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO/Al ₂ O ₃		0.35	0.35	0.35	0.35	0.36	0.36	0.36	0.36	0.36	0.36	0.36	0.36
Olivine vol.%													
Cpx vol.%		36.30	36.20	36.00	35.90	35.70	35.60	35.40	35.30	35.10	35.00	34.80	34.70
Plag vol.%		35.80	35.70	35.50	35.40	35.30	35.10	35.00	34.80	34.70	34.50	34.40	34.20
Vol.% crystals		72.10	71.80	71.50	71.30	71.00	70.70	70.40	70.10	69.80	69.50	69.20	68.90
Fo%													
An%		55.30	55.90	55.90	55.90	55.90	56.40	56.40	56.40	56.40	56.40	56.40	57.40

Bulk composition		MELTS											
Input H ₂ O		0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011
P (MPa)		290	285	280	275	270	265	260	255	250	245	240	235
SiO ₂	50.9	53.39	53.38	53.36	53.35	53.34	53.32	53.31	53.29	53.28	53.26	53.24	53.23
TiO ₂	0.9	2.01	2.00	1.99	1.98	1.97	1.96	1.95	1.94	1.93	1.92	1.91	1.90
Al ₂ O ₃	16.6	17.66	17.63	17.60	17.57	17.54	17.51	17.48	17.45	17.42	17.39	17.36	17.33
Fe ₂ O ₃		2.38	2.40	2.42	2.44	2.46	2.49	2.51	2.53	2.55	2.57	2.59	2.61
FeO	8.22	5.56	5.60	5.65	5.69	5.73	5.78	5.82	5.86	5.91	5.95	5.99	6.03
MnO	0.16	0.57	0.56	0.55	0.55	0.54	0.54	0.53	0.53	0.52	0.51	0.51	0.50
MgO	6.66	2.62	2.65	2.67	2.70	2.73	2.76	2.79	2.82	2.85	2.88	2.91	2.94
CaO	10.9	6.37	6.38	6.39	6.39	6.40	6.41	6.41	6.42	6.43	6.44	6.45	6.46
Na ₂ O	2.39	2.75	2.76	2.77	2.78	2.79	2.81	2.82	2.83	2.84	2.85	2.86	2.87
K ₂ O	1.82	5.28	5.24	5.21	5.17	5.13	5.09	5.06	5.02	4.98	4.94	4.91	4.87
P ₂ O ₅	0.4	1.41	1.40	1.38	1.37	1.35	1.34	1.33	1.31	1.30	1.28	1.27	1.26
H ₂ O		0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO/Al ₂ O ₃		0.36	0.36	0.36	0.36	0.37	0.37	0.37	0.37	0.37	0.37	0.37	0.37
Olivine vol.%													
Cpx vol.%		34.50	34.30	34.20	34.00	33.80	33.60	33.50	33.30	33.10	32.90	32.70	32.50
Plag vol.%		34.10	33.90	33.80	33.60	33.50	33.30	33.10	33.00	32.80	32.70	32.50	32.30
Vol.% crystals		68.60	68.30	67.90	67.60	67.30	66.90	66.60	66.30	65.90	65.60	65.20	64.90
Fo%													
An%		57.40	57.40	57.40	57.40	57.90	57.90	57.90	58.50	58.50	58.90	58.90	58.90

Bulk composition		MELTS											
Input H ₂ O		0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011
P (MPa)		230	225	220	215	210	205	200	190	180	170	160	150
SiO ₂	50.9	53.21	53.19	53.18	53.16	53.14	53.12	53.10	53.06	53.02	52.98	52.94	52.90
TiO ₂	0.9	1.89	1.88	1.87	1.86	1.85	1.84	1.83	1.81	1.79	1.77	1.76	1.74
Al ₂ O ₃	16.6	17.30	17.27	17.24	17.21	17.18	17.15	17.12	17.06	17.00	16.94	16.88	16.82
Fe ₂ O ₃		2.63	2.65	2.67	2.69	2.71	2.73	2.75	2.79	2.83	2.87	2.90	2.94
FeO	8.22	6.08	6.12	6.16	6.20	6.24	6.28	6.32	6.40	6.47	6.55	6.62	6.69
MnO	0.16	0.48	0.49	0.49	0.48	0.48	0.47	0.47	0.46	0.45	0.44	0.43	0.42
MgO	6.66	2.97	3.00	3.03	3.07	3.10	3.13	3.17	3.24	3.31	3.38	3.46	3.54

CaO	10.9	6.47	6.48	6.49	6.50	6.51	6.53	6.58	6.57	6.59	6.62	6.66	6.69
Na ₂ O	2.39	2.88	2.89	2.90	2.90	2.91	2.92	2.93	2.94	2.95	2.96	2.97	2.98
K ₂ O	1.82	4.83	4.79	4.76	4.72	4.68	4.64	4.61	4.53	4.46	4.39	4.32	4.25
P ₂ O ₅	0.4	1.24	1.23	1.22	1.21	1.19	1.18	1.17	1.14	1.12	1.09	1.07	1.05
H ₂ O		0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO/Al ₂ O ₃		0.37	0.38	0.38	0.38	0.38	0.38	0.39	0.39	0.39	0.39	0.39	0.40
Olivine vol.%													
Cpx vol.%		32.30	32.10	31.90	31.70	31.50	31.30	31.10	30.70	30.30	29.80	29.40	28.90
Plag vol.%		32.20	32.00	31.80	31.70	31.50	31.30	31.20	30.80	30.50	30.10	29.80	29.40
Vol.% crystals		64.50	64.10	63.80	63.40	63.00	62.70	62.30	61.50	60.70	59.90	59.10	58.30
Fo%													
An%		58.90	58.90	60.00	60.00	60.00	60.00	60.00	60.40	61.10	61.50	61.50	62.50
Bulk composition		MELTS											
Input H ₂ O		0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011	0.0011
P (MPa)		140	130	120	110	100	90	80	75	70	60	50	40
SiO ₂	50.9	52.85	52.81	52.77	52.82	52.88	52.94	53.01	53.04	53.08	53.15	53.22	53.22
TiO ₂	0.9	1.72	1.70	1.68	1.66	1.65	1.63	1.61	1.60	1.59	1.56	1.54	1.53
Al ₂ O ₃	16.6	16.76	16.70	16.64	16.54	16.44	16.33	16.23	16.18	16.12	16.02	15.92	15.85
Fe ₂ O ₃		2.98	3.01	3.04	3.08	3.11	3.14	3.17	3.18	3.19	3.22	3.25	3.27
FeO	8.22	6.75	6.81	6.87	6.93	6.99	7.04	7.10	7.12	7.14	7.19	7.22	7.23
MnO	0.16	0.41	0.40	0.39	0.37	0.35	0.33	0.32	0.31	0.30	0.28	0.27	0.26
MgO	6.66	3.62	3.70	3.78	3.84	3.90	3.97	4.03	4.07	4.10	4.16	4.23	4.29
CaO	10.9	6.72	6.76	6.80	6.85	6.90	6.95	7.01	7.04	7.07	7.13	7.19	7.26
Na ₂ O	2.39	2.99	2.99	2.99	2.98	2.96	2.94	2.92	2.91	2.90	2.88	2.86	2.85
K ₂ O	1.82	4.18	4.11	4.04	3.96	3.88	3.80	3.72	3.68	3.64	3.55	3.47	3.42
P ₂ O ₅	0.4	1.03	1.01	0.98	0.96	0.94	0.92	0.89	0.88	0.87	0.85	0.82	0.81
H ₂ O		0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO/Al ₂ O ₃		0.40	0.41	0.41	0.41	0.42	0.43	0.43	0.44	0.44	0.45	0.45	0.46
Olivine vol.%		0.31	0.67	1.05	1.46	1.68	1.90	2.36	2.78	2.89			
Cpx vol.%		28.40	27.90	27.40	26.50	25.40	24.20	23.00	22.40	21.80	20.40	19.20	18.50
Plag vol.%		29.00	28.70	28.30	28.00	27.70	27.40	27.00	26.90	26.70	26.30	25.90	25.60
Vol.% crystals		57.50	56.60	55.70	54.80	53.70	52.70	51.50	51.00	50.40	49.10	47.90	47.10
Fo%									78.60	78.80	78.80	78.80	78.80
An%		62.50	62.50	63.50	63.50	63.90	64.60	64.90	64.90	64.90	65.60	66.00	66.00
Bulk composition		MELTS											
Input H ₂ O				0.0011		0.0011		0.0011		0.0011		0.0011	
P (MPa)				20		10		5		2		0.1	
SiO ₂	50.9			53.21		53.20		53.20		53.19		53.19	
TiO ₂	0.9			1.49		1.48		1.47		1.47		1.46	
Al ₂ O ₃	16.6			15.71		15.65		15.62		15.60		15.59	
Fe ₂ O ₃				3.31		3.33		3.34		3.34		3.35	
FeO	8.22			7.24		7.24		7.24		7.24		7.24	
MnO	0.16			0.26		0.25		0.25		0.25		0.25	
MgO	6.66			4.40		4.46		4.49		4.50		4.51	
CaO	10.9			7.41		7.49		7.53		7.55		7.56	
Na ₂ O	2.39			2.85		2.85		2.84		2.84		2.84	
K ₂ O	1.82			3.33		3.28		3.22		3.25		3.24	
P ₂ O ₅	0.4			0.78		0.77		0.76		0.76		0.76	
H ₂ O				0.00		0.00		0.00		0.00		0.00	
CaO/Al ₂ O ₃				0.47		0.48		0.48		0.48		0.49	
Olivine vol.%				3.04		3.09		3.12		3.13		3.14	
Cpx vol.%				17.30		16.70		16.50		16.30		16.20	
Plag vol.%				25.10		24.80		24.60		24.50		24.50	
Vol.% crystals				45.40		44.60		44.20		43.90		43.80	
Fo%				79.80		79.80		79.80		79.80		79.80	
An%				67.00		68.00		68.00		68.00		68.00	

Table 3
Computed melt composition (wt.%) and crystallisation path (vol.% crystals in the melt) of ascending Stromboli magma as a function of pressure using MELTS with a bulk composition of pumice ST82 as given by Métrich et al. (2010) for T = 1115 °C.

Bulk composition												
Input H ₂ O	3.64	3.59	3.54	3.48	3.43	3.38	3.33	3.28	3.22	3.17	3.12	3.07
P (MPa)	350	345	340	335	330	325	320	315	310	305	300	295
SiO ₂	50.9	49.57	49.60	49.63	49.66	49.69	49.72	49.75	49.82	49.85	49.88	49.91
TiO ₂	0.9	0.90	0.88	0.88	0.88	0.88	0.88	0.88	0.89	0.89	0.89	0.89
Al ₂ O ₃	16.6	17.59	17.60	17.62	17.64	17.66	17.68	17.70	17.72	17.73	17.75	17.80
Fe ₂ O ₃	2.31	2.32	2.32	2.33	2.33	2.33	2.34	2.34	2.35	2.35	2.36	2.37
FeO	8.22	5.93	5.93	5.93	5.93	5.93	5.92	5.92	5.92	5.92	5.92	5.92
MnO	0.16	0.18	0.18	0.18	0.18	0.18	0.18	0.18	0.18	0.18	0.18	0.18
MgO	6.66	5.22	5.22	5.22	5.22	5.22	5.22	5.22	5.22	5.22	5.21	5.21
CaO	10.9	9.25	9.25	9.24	9.23	9.22	9.22	9.21	9.20	9.19	9.18	9.16
Na ₂ O	2.39	2.61	2.61	2.62	2.62	2.62	2.63	2.63	2.64	2.64	2.65	2.65
K ₂ O	1.82	2.00	2.01	2.01	2.01	2.02	2.02	2.02	2.03	2.03	2.03	2.04
P ₂ O ₅	0.4	0.44	0.44	0.44	0.44	0.44	0.44	0.44	0.45	0.45	0.45	0.45
H ₂ O	4.01	3.96	3.91	3.85	3.80	3.75	3.70	3.65	3.59	3.54	3.49	3.44
CaO/Al ₂ O ₃	0.53	0.53	0.52	0.52	0.52	0.52	0.52	0.52	0.52	0.52	0.52	0.51
Olivine vol.%	0.23	0.22	0.21	0.20	0.20	0.19	0.18	0.17	0.16	0.15	0.14	0.13
Cpx vol.%	9.32	9.38	9.47	9.54	9.61	9.69	9.78	9.87	9.95	10.00	10.10	10.20
Plag vol.%	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Vol.% crystals	9.55	9.60	9.68	9.74	9.81	9.88	9.96	10.00	10.10	10.20	10.30	10.40
Fo%	84.80	84.80	84.80	84.80	84.80	84.80	84.80	84.80	84.80	84.80	84.80	84.80
An%												

Bulk composition												
Input H ₂ O	3.02	2.96	2.91	2.86	2.81	2.76	2.70	2.65	2.60	2.55	2.50	2.44
P (MPa)	290	285	280	275	270	265	260	255	250	245	240	235
SiO ₂	50.9	49.94	49.97	50.01	50.04	50.07	50.10	50.14	50.17	50.20	50.24	50.31
TiO ₂	0.9	0.89	0.89	0.89	0.89	0.89	0.89	0.89	0.90	0.90	0.90	0.90
Al ₂ O ₃	16.6	17.82	17.84	17.86	17.88	17.91	17.93	17.95	17.98	18.00	18.03	18.08
Fe ₂ O ₃	2.37	2.37	2.38	2.38	2.39	2.39	2.40	2.40	2.41	2.41	2.41	2.42
FeO	8.22	5.91	5.91	5.91	5.91	5.91	5.90	5.90	5.90	5.89	5.89	5.89
MnO	0.16	0.18	0.18	0.18	0.18	0.18	0.18	0.18	0.18	0.18	0.18	0.18
MgO	6.66	5.21	5.21	5.20	5.20	5.19	5.19	5.19	5.18	5.18	5.17	5.17
CaO	10.9	9.15	9.14	9.13	9.12	9.10	9.09	9.08	9.06	9.05	9.03	9.00
Na ₂ O	2.39	2.65	2.66	2.66	2.67	2.67	2.68	2.68	2.69	2.70	2.70	2.71
K ₂ O	1.82	2.04	2.05	2.05	2.05	2.06	2.06	2.07	2.07	2.08	2.08	2.09
P ₂ O ₅	0.4	0.45	0.45	0.45	0.45	0.45	0.45	0.45	0.46	0.46	0.46	0.46
H ₂ O	3.38	3.33	3.28	3.23	3.17	3.12	3.07	3.02	2.96	2.91	2.86	2.80
CaO/Al ₂ O ₃	0.51	0.51	0.51	0.51	0.51	0.51	0.51	0.50	0.50	0.50	0.50	0.50
Olivine vol.%	0.12	0.12	0.11	0.10	0.08	0.07	0.07	0.06	0.05	0.03	0.03	0.02
Cpx vol.%	10.30	10.40	10.50	10.60	10.80	10.90	11.00	11.10	11.20	11.40	11.50	11.60
Plag vol.%	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Vol.% crystals	10.50	10.60	10.60	10.70	10.80	10.90	11.10	11.20	11.30	11.40	11.50	11.60
Fo%	84.80	84.80	84.80	84.00	84.00	84.00	84.00	84.00	84.00	84.00	84.00	84.00
An%												

Bulk composition												
Input H ₂ O	2.39	2.34	2.29	2.24	2.18	2.13	2.08	1.98	1.87	1.77	1.66	1.56
P (MPa)	230	225	220	215	210	205	200	190	180	170	160	150
SiO ₂	50.9	50.34	50.38	50.42	50.46	50.50	50.54	50.58	50.66	50.74	50.83	51.01
TiO ₂	0.9	0.90	0.90	0.90	0.90	0.90	0.90	0.91	0.91	0.91	0.91	0.91
Al ₂ O ₃	16.6	18.10	18.13	18.16	18.19	18.22	18.25	18.28	18.34	18.41	18.48	18.63
Fe ₂ O ₃	2.42	2.43	2.43	2.43	2.44	2.44	2.44	2.44	2.45	2.45	2.46	2.47
FeO	8.22	5.89	5.88	5.88	5.87	5.87	5.87	5.86	5.85	5.84	5.83	5.80
MnO	0.16	0.18	0.18	0.18	0.19	0.19	0.19	0.19	0.19	0.19	0.19	0.19
MgO	6.66	5.16	5.15	5.14	5.13	5.12	5.11	5.10	5.07	5.04	5.01	4.98
CaO	10.9	8.99	8.97	8.95	8.94	8.92	8.90	8.89	8.85	8.81	8.76	8.66
Na ₂ O	2.39	2.71	2.72	2.73	2.73	2.74	2.74	2.75	2.76	2.78	2.79	2.81
K ₂ O	1.82	2.09	2.10	2.10	2.11	2.11	2.12	2.12	2.14	2.15	2.16	2.19
P ₂ O ₅	0.4	0.46	0.46	0.46	0.46	0.46	0.47	0.47	0.47	0.47	0.48	0.48
H ₂ O	2.75	2.70	2.64	2.59	2.54	2.48	2.43	2.32	2.21	2.10	1.99	1.88
CaO/Al ₂ O ₃	0.50	0.49	0.49	0.49	0.49	0.49	0.49	0.49	0.48	0.48	0.47	0.47
Olivine vol.%	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Cpx vol.%	11.80	11.90	12.00	12.20	12.30	12.50	12.60	13.00	13.30	13.60	14.00	14.40
Plag vol.%	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Vol.% crystals	11.80	11.90	12.00	12.20	12.30	12.50	12.60	13.00	13.30	13.60	14.00	14.40
Fo%	84.00											
An%												

Bulk composition												
Input H ₂ O	1.46	1.35	1.25	1.14	1.04	0.94	0.83	0.78	0.73	0.62	0.52	0.42
P (MPa)	140	130	120	110	100	90	80	75	70	60	50	40
SiO ₂	50.9	51.11	51.21	51.32	51.43	51.57	51.78	51.99	52.10	52.20	52.43	52.88
TiO ₂	0.9	0.92	0.92	0.92	0.93	0.96	0.99	1.02	1.04	1.06	1.10	1.20
Al ₂ O ₃	16.6	18.71	18.79	18.87	18.87	18.66	18.40	18.13	18.00	17.86	17.57	16.96

Table 3 (continued)

Bulk composition													
Fe ₂ O ₃		2.47	2.47	2.48	2.49	2.55	2.61	2.67	2.70	2.74	2.81	2.88	2.96
FeO	8.22	5.78	5.76	5.74	5.76	5.85	5.95	6.05	6.10	6.15	6.27	6.39	6.52
MnO	0.16	0.19	0.20	0.20	0.20	0.20	0.21	0.21	0.21	0.22	0.22	0.22	0.23
MgO	6.66	4.92	4.88	4.84	4.82	4.84	4.78	4.73	4.69	4.66	4.59	4.51	4.42
CaO	10.9	8.61	8.55	8.49	8.40	8.28	8.19	8.09	8.04	7.98	7.87	87.75	7.62
Na ₂ O	2.39	2.84	2.86	2.88	2.91	2.93	2.96	2.99	3.00	3.01	3.03	3.04	3.04
K ₂ O	1.82	2.20	2.22	2.24	2.27	2.32	2.38	2.46	2.50	2.54	2.63	2.73	2.85
P ₂ O ₅	0.4	0.48	0.49	0.49	0.50	0.51	0.53	0.54	0.55	0.56	0.59	0.61	0.64
H ₂ O		1.76	1.65	1.53	1.42	1.33	1.23	1.13	1.08	1.03	0.91	0.80	0.67
CaO/Al ₂ O ₃		0.46	0.46	0.45	0.45	0.44	0.44	0.45	0.45	0.45	0.45	5.08	0.45
Olivine vol.%		0.00	0.00	0.00	0.00	0.10	0.46	0.82	1.01	1.19	1.56	1.95	2.33
Cpx vol.%		14.80	15.30	15.70	16.10	16.30	16.10	16.00	16.00	16.00	16.00	16.10	16.30
Plag vol.%		0.00	0.00	0.00	0.51	2.18	4.19	6.28	7.34	8.44	10.70	13.10	15.60
Vol.% crystals		14.80	15.30	15.70	16.70	18.50	20.80	23.10	24.30	25.60	28.30	31.10	34.20
Fo%						83.00	83.00	82.80	82.80	82.00	81.80	81.80	80.80
An%					81.80	80.80	79.80	78.80	78.80	77.80	76.80	75.50	73.70
Bulk composition													
Input H ₂ O				0.21		0.10		0.05		0.02			0.0010
P (MPa)				20		10		5		2			0.1
SiO ₂	50.9			53.35		53.58		53.62		53.76			53.79
TiO ₂	0.9			1.35		1.46		1.61		1.62			1.63
Al ₂ O ₃	16.6			16.30		15.93		16.11		15.75			15.55
Fe ₂ O ₃				3.14		3.24		3.16		3.27			3.34
FeO	8.22			6.81		6.98		6.93		7.08			7.15
MnO	0.16			0.24		0.25		0.29		0.27			0.26
MgO	6.66			4.20		4.06		3.73		3.81			3.86
CaO	10.9			7.33		7.17		6.81		6.92			7.00
Na ₂ O	2.39			3.00		2.93		2.91		2.83			2.80
K ₂ O	1.82			3.17		3.40		3.81		3.75			3.73
P ₂ O ₅	0.4			0.73		0.76		0.92		0.90			0.90
H ₂ O				0.38		0.21		0.12		0.05			0.00
CaO/Al ₂ O ₃				0.45		0.45		0.42		0.44			0.45
Olivine vol.%				3.15		3.59		3.00		3.81			4.03
Cpx vol.%				17.60		20.60		23.60		23.80			24.20
Plag vol.%				20.70		21.70		25.20		24.20			23.40
Vol.% crystals				41.50		45.90		51.80		51.80			51.50
Fo%				79.80		78.80		76.80		77.60			77.80
An%				70.40		68.00		64.90		64.90			64.90

3. Constraining the Stromboli system

Two distinct products are ejected at Stromboli; normal Strombolian explosions eject a crystal-rich, highly porphyritic (HP) black scoria, while the major eruptions and the most violent, paroxysmal, activity eject a low porphyritic (LP), volatile-rich, crystal-poor “golden” pumice in addition to the crystal-rich HP scoria (Francalanci et al., 1999). The bulk rock compositions of both the pumice and scoria are very similar, with 49–53 wt.% SiO₂ and 1.6–2.2 wt.% K₂O (Métrich et al., 2001; Landi et al., 2004), but have different crystal contents and glass compositions. The pumice contains <10 vol.% (vesicle-free) microphenocrysts of olivine and clinopyroxene in a shoshonitic glass (Métrich et al., 2001, 2010), whereas the scoria have ~45–55 vol.% (vesicle-free) crystals, made up of ~10–17 vol.% clinopyroxene, ~2–6 vol.% olivine and are dominated by ~27–38 vol.% plagioclase (Métrich et al., 2001; Francalanci et al., 2004; Landi et al., 2004; Métrich et al., 2010). As a result of crystallisation, the glasses of the scoria are more evolved than those of the pumice, with SiO₂ contents ranging from 51 to 53 wt.% and K₂O contents from 3.5 to 5.2 wt.% (Francalanci et al., 2005).

Melt inclusions (MIs) trapped in olivine provide information on the chemical and volatile composition of the melt at the depth at which the MI formed. Olivine-hosted MIs found in the pumice at Stromboli have high volatile contents, 2.3–3.8 wt.% H₂O and 894–1704 ppm CO₂ (Métrich et al., 2001, 2010), whereas the olivine-hosted MIs in the scoria are more extensively degassed, 0.4–1.1 wt.% H₂O and 273–473 ppm CO₂ (Métrich et al., 2010). This leads Métrich et al. (2001, 2010) to conclude that the MIs in the scoria are a result of decompression-driven degassing and crystallisation of the volatile-rich magma represented by the MI in the golden pumice.

3.1. Modelling the degassing path

Assuming equilibrium and volatile saturation, entrapment pressures for the MI compositions given by Métrich et al. (2010) are calculated using VolatileCalc (Newman and Lowenstern, 2002), which has been shown to describe the solubility of CO₂ and H₂O bearing fluids in Strombolian basalts well (Lesne et al., 2011; Witham et al., 2012). Entrapment pressures for MI in the scoria are found to range from 75 to 110 MPa and in the pumice from 210 to 340 MPa at 1130 °C (Fig. 2). Although simple open- or closed-system equilibrium degassing models (e.g. Dixon, 1997; Newman and Lowenstern, 2002; Papale et al., 2006) agree well with experimentally observed degassing paths, the models fail to reproduce the degassing trends represented by the MI; they underestimate the amount of dissolved CO₂ in the melt. This could be explained by fluxing of CO₂ through the magma from depth which would drive dehydration of the melt (Blundy et al., 2010; Métrich et al., 2010), or by mixing of ascending and descending magmas as discussed by Witham (2011a). Here we are interested only in the volatile content of the melt as a function of pressure and not the mechanism of the degassing path. Witham (2011b) highlights that if there is mixing between the two magmas, some melt may be undersaturated in volatiles. Here we follow Métrich et al. (2010) in assuming that the melt is volatile saturated and perform a simple empirical linear fit to the X_{H₂O} – P and X_{CO₂} – P data where X is the concentration (wt.%) and P is the pressure (MPa), such that X_{H₂O} = 0.0105 · P and X_{CO₂} = 5.17 · P (Fig. 2). This approach allows us to reproduce well the composition and crystallinity of the pumice and scoria ejected at Stromboli, as we discuss in Section 3.2.

We calculate the saturation pressure of the most primitive MI given by Métrich et al. (2010) (ST82 139a 3.5 wt.% H₂O, 1704 ppm CO₂ and

47.5 wt.% SiO₂) using VolatileCalc (Newman and Lowenstern, 2002) to be 336 MPa for which there is a CO₂/H₂O weight ratio in the equilibrium gas phase of 3.71. From Fourier Transform Infrared Spectroscopy (FTIR) measurements of gas CO₂/SO₂ molar ratios (M. Burton et al., 2007) during persistent degassing, M.R. Burton et al. (2007) estimate the total primitive CO₂ content to be 2.44 wt.%, assuming bulk magma degassing. Assuming closed-system degassing this suggests that at 336 MPa, 2.27 wt.% CO₂ and 0.716 wt.% H₂O are already in the gas phase. This would require a total primitive H₂O content of 4.22 wt.%. Given these total primitive contents of H₂O and CO₂ and the wt.% of dissolved volatiles in the melt (Fig. 2) we can calculate the wt.% of exsolved volatiles as a function of pressure. To determine the volume occupied by the exsolved gas we use the equation of state for H₂O and CO₂ given by Pitzer and Sterner (1994). We assume the magma will become permeable to gas flow at 50 vol.% bubbles in the magma (melt + bubbles + crystals) following M.R. Burton et al. (2007), which is found to occur at 75 MPa, from which point the vesicularity of the magma is assumed to remain constant (equal to 50 vol.%) and any extra gas is lost from the system.

3.2. Modelling the crystallisation path

The crystallisation paths and chemical compositions of the melts for both the ascending and descending magmas are modelled as a function of pressure using MELTS (Ghiorso and Sack, 1995; Asimow and Ghiorso, 1998). We use the whole rock composition of pumice ST82 (Métrich et al., 2010) and assume that the system is in equilibrium and isothermal, and that there is no fractional crystallisation. We fix f_{O_2} at $\Delta NNO + 1$. MELTS does not account for CO₂, which acts to lower the effective solubility of H₂O. We assume that CO₂ does not affect crystallisation other than by reducing the activity of water in the melt, and thus is inert in regard to the crystal phase assemblage (e.g. Pichavant et al., 2009). Further we follow Dixon and Stolper (1995) in assuming ideal mixing in the fluid phase. To model the ascending magma we run MELTS in a step-wise manner, with the activity of H₂O in the melt calculated using the linear fit of $X_{H_2O} - P$ (Fig. 2) from 350 to 0.1 MPa (over 53 pressure steps). As the fluid phase in MELTS is pure water only, the programme considers our modelled melt to be undersaturated in H₂O. This, however, allows us to represent well a melt saturated in a mixed CO₂ and H₂O fluid phase. We assume that the descending magma has no bubbles, a reasonable approximation for magma density and viscosity calculations given that any remaining bubbles would resorb back into the melt as the pressure increases. The H₂O content of the descending magma is assumed to be constant and taken to be that of the ascending magma at the lowest pressure achieved before the magma overturns. Optical thermometry experiments on the MIs suggest homogenisation temperatures ranging from 1160 to 1110 °C (Métrich et al., 2001); we find that a temperature of 1130 °C results in crystallinities and melt compositions that represent well the properties of the pumice and scoria as a function of pressure (Figs. 3 and 4). Data from all of the MELTS runs and the bulk composition of pumice ST82 (Métrich et al., 2010) are given in Tables 1, 2, and 3.

Our modelled melt at 350 MPa has a SiO₂ content of 49 wt.% and K₂O content of 2 wt.%, and at 0.1 MPa a SiO₂ content of 53 wt.% and K₂O content of 3 wt.%. Actual olivines in the Stromboli pumice span a range of compositions from Fo_{91–84} and those in the scoria range from Fo_{69–70} (Métrich et al., 2010). The olivine output from our petrologic modelling has a composition ranging from Fo_{86–85} over pressures of 350–270 MPa, which agrees well with the natural olivines. At pressures ≤ 90 MPa, the modelled olivine chemistry is slightly more forsteritic (Fo_{68–81}) than natural olivine in scoria (Fo_{69–70}). This is not significantly improved by using $T = 1115$ °C which results in compositions ranging from Fo_{65–80} over 100–0.1 MPa (for which olivine is stable at this temperature). Taking the density of olivine to be 3300 kg m⁻³, clinopyroxene to be 3400 kg m⁻³ and plagioclase to be 2800 kg m⁻³ (Ahrens, 1995) the crystal content in the melt, excluding bubbles, at 350 MPa is 5 vol.%

Fig. 5. Volume percent of crystals, melt and bubbles in the ascending magma at Stromboli. The dashed line indicates the onset of open-system degassing at 75 MPa, where the vesicularity of the magma has reached 50 vol.%.

made up of olivine and clinopyroxene. In our model, plagioclase begins to form at 90 MPa and by 40 MPa the crystallinity is 27 vol.% (1 vol.% olivine, 14 vol.% clinopyroxene, 12 vol.% plagioclase). A final crystal content of 44 vol.% (3 vol.% olivine, 16 vol.% clinopyroxene and 25 vol.% plagioclase) was achieved at 0.1 MPa. As the H₂O-undersaturated magma returns back down the conduit it further crystallises upon descent, as the solidus temperature increases with an increase in pressure; at 350 MPa magma which has descended from 0.1 MPa has a crystallinity of 74 vol.%. Comparing the volume percent of each crystal phase from the petrologic modelling to the crystal contents of scoria presented by Métrich et al. (2010) it is evident that there is a slight mismatch between the petrologic modelling and that observed in the scoria ejected at Stromboli: the vol.% of plagioclase is slightly underestimated by the petrologic modelling with 24 vol.% at 0.1 MPa compared to 27–38 vol.% found in the scoria (Fig. 4). This is not improved significantly by using a bulk composition of scoria (ST82s) (Métrich et al., 2001) which results in 27 vol.% plagioclase at 0.1 MPa. Neither is it substantially improved by lowering the temperature from 1130 °C to 1115 °C, which results in a total crystallinity of 55 vol.%, with only 23 vol.% represented by plagioclase at 0.1 MPa.

In this study we calculate magma saturation pressures for the MI in the pumice to range from 210 to 340 MPa at a temperature of 1130 °C using VolatileCalc (Newman and Lowenstern, 2002), which has recently been shown to best represent the solubility of H₂O and CO₂ bearing fluids in Stromboli basalt (Lesne et al., 2011; Witham et al., 2012). Using the VolatileCalc model (Newman and Lowenstern, 2002) results in greater saturation pressures for the MI in the pumice than those obtained using the (Papale et al., 2006) model: 130–250 MPa (Métrich et al., 2010). Using MELTS we find that $T = 1130$ °C best represents crystal volume fractions, melt composition and mineral chemistry of the Stromboli magma over 350–0.1 MPa. This is a lower temperature than has been determined by crystallisation experiments conducted by Di Carlo et al. (2006) using a sample of pumice from Stromboli (PST 9, SiO₂ 49.4 wt.% K₂ 1.85 wt.%) who constrained pre-eruptive conditions to be 100–270 MPa and 1140–1160 °C with 2.5–2.7 wt.% H₂O, and also later Pichavant et al. (2009) who further constrained the pre-eruptive conditions of PST-9 to be 1150 °C and 150–250 MPa. As a result we suggest deeper pre-eruptive storage conditions for LP magma of 8–12 km (using an average crustal density of 2700 kg m⁻³ following Métrich et al. (2010)) compared to 7–10 km concluded by Métrich

et al. (2010), and ~8 km by Di Carlo et al. (2006) and Pichavant et al. (2009).

Métrich et al. (2010) also used MELTS to compute the crystallisation conditions of olivine; they found that olivine Fo_{89.8–88.7} joins the liquidus phase with clinopyroxene in equilibrium with a melt having ~2.9 wt.% H₂O and ~8.1 wt.% MgO between 300 and 250 MPa, and that the two phases start crystallising between 1190 and 1185 °C at ΔNNO + 1, but they note that the temperature of crystallisation may be unreliable due to MELTS not treating the effect of CO₂ on the phase equilibria. Here we represent a melt saturated in a mixed CO₂ and H₂O fluid phase by forcing MELTS to be undersaturated in H₂O. We find that olivine is only stable over 275–350 MPa and 0–75 MPa at T = 1130 °C. Given that MIs are exclusively from olivine this explains the discontinuous nature of the MI saturation pressures.

We have calculated the volume percent of gas exsolved in the ascending magma (Section 3.1) and the volume percent of crystals in the melt. It is then important to define the volume percent of each phase as a function of the total volume of the melt, crystals and bubbles. In Fig. 5 the volume percent of crystals, bubbles and melt is re-calculated relative to the total volume of melt + bubbles + crystals. It is shown that at 0.1 MPa the crystallinity of the ascending magma is now defined as 22 vol.%, as melt occupies 28 vol.% and gas occupies 50 vol.% of the total volume.

4. Physical properties of Strombolian magma

4.1. Magma rheology

Calculating the viscosity of a three phase fluid (crystals, melt, bubbles) is challenging, and to date there is no complete model for this. We therefore exploit knowledge on the 2-phase end members, of a bubble suspension and of a crystal suspension. We explore two approaches: in the first we calculate the viscosity of the magma by first calculating the viscosity of the bubble–melt suspension, and use this as the effective suspending fluid in the rheological model for the crystal suspension; in the second approach we first calculate the viscosity of the crystal–melt suspension and use this as the effective suspending fluid in the rheological model for the bubble suspension. The limitation of this method is that we do not account for bubble–crystal interactions on the final viscosity of the system.

We use the model of Giordano et al. (2008) to calculate the viscosity of the silicate melt as a function of its chemical composition (determined from the MELTS calculations) at a temperature of 1130 °C and fluorine content of 0.1 wt.% (Métrich et al., 2001) (Fig. 6).

For steady flow the rheology of the bubble–melt suspension depends on the melt viscosity, abundance of bubbles and the shape of the bubbles. Bubble shape depends on the Capillary number (Ca), which defines the balance of viscous stresses which act to deform the bubbles and interfacial stresses which act to restore them to sphericity:

$$Ca = \lambda \dot{\gamma}, \quad (2)$$

where $\lambda = \mu R_b / \Gamma$ is the bubble relaxation time, μ is the viscosity of the suspending fluid (the melt), R_b is the radius of the bubble found to range from 10^{-2} to 10^{-5} m in the pumice and scoria ejected at Stromboli (Lautze and Houghton, 2007), Γ is the liquid–bubble interfacial tension which is ~0.25 N m⁻¹ for basaltic magma (Murase and McBirney, 1973), and $\dot{\gamma}$ is strain rate.

We can estimate a characteristic strain rate for an exchange flow at Stromboli, given the volume flux (Q) of the flow:

$$\dot{\gamma} = Q / r_L^3, \quad (3)$$

where r_L is the characteristic length scale and depends on the flow geometry. For SBS flow we take $r_L = r_2$, where r_2 is the conduit radius. For CAF $r_L = r_1$ for the inner (core) fluid, where r_1 is the radius of the

Fig. 6. Calculated viscosities of ascending and descending melt at Stromboli.

core, and for the descending outer fluid $r_L = r_2$. We assume that the flow is steady and that Q , and thus $\dot{\gamma}$, do not change substantially with depth in the conduit; in Section 5.1 we show that the decrease in magma density during ascent reduces Q by less than an order of magnitude.

Eq. (3) does not account for variation in strain rate with radial position in the conduit. To validate this approach we apply Eq. (3) to laboratory exchange flows presented by Beckett et al. (2011) and compare to the actual strain rates calculated from flow velocities, $\dot{\gamma}(r) = \frac{du}{dr}$ (Figs. 7 and 8). For CAF $\dot{\gamma}(r)$ is calculated using the best fit velocity profiles to the theoretical solution of Huppert and Hallworth (2007) (Fig. 9; Beckett et al., 2011). For SBS flow $\dot{\gamma}(r)$ is calculated from high order polynomial fits to the measured data points because the available theoretical velocity profiles of Kerswell (2011) do not well represent the SBS experimental data (Fig. 10; Beckett et al., 2011).

Figs. 7 and 8 show that using the volume flux of the flow (Eq. (3)) gives useful order of magnitude estimates for strain rate. Given that the minimum observed radius of the ascending fluid in the CAF regime from the exchange flow experiments is $0.2r_2$ (Table 2; Beckett et al., 2011) and taking the volume flux of the magma at Stromboli required to account for measured persistent gas fluxes to be $\sim 10^{-1}$ m³ s⁻¹ we can estimate the strain rate acting on an exchange flow of magma in a conduit with radius 1–10 m to range from 10^1 to 10^{-4} s⁻¹.

Ca is now calculated for an exchange flow of magma at Stromboli as a function of pressure for a conduit with radius 1–10 m (i.e. $\dot{\gamma} = 10^1$ – 10^{-4} s⁻¹) (Fig. 9). For most of the conditions considered $Ca \ll 1$, except for the extreme case where $r_2 = 1$ m and $R_b = 10^{-2}$ m for which $Ca \approx O(100)$. However, taking $Ca < 1$ would predict relative viscosities for $Ca \leq 100$ within an order of magnitude for $\phi_b \leq 0.5$, where ϕ_b is the bubble volume fraction (Pal, 2004). We therefore assume that $Ca \leq O(1)$ best represents the flow: the bubbles remain spherical and with an increase in bubble volume fraction (ϕ_b) viscosity increases (Rust and Manga, 2002). The rheology of bubble suspensions for low Ca number is given by (Pal, 2002; Llewellyn and Manga, 2005; Mader et al., 2013):

$$\mu_r = (1 - \phi_b)^{-1} \quad (4)$$

where μ_r is relative viscosity, normalised to that of the interstitial fluid, and ϕ_b is the volume fraction of bubbles.

Fig. 7. Strain rate ($\dot{\gamma}$) acting on CAFs presented in Beckett et al. (2011). The dotted vertical lines represent the fluid interface (r_1). The solid black lines represent calculated strain rates from best fit velocity profiles of the flow using the theoretical solution of Huppert and Hallworth (2007) (Fig. 9; Beckett et al., 2011). The solid horizontal grey line represents calculated strain rates given the volume flux of the flow, Q , see text for details. It should be noted that volume fluxes are known for only four out of a total of six flows, and so comparisons are limited to these experiments.

There are a number of published models to describe the rheological behaviour of fluids with a crystal volume fraction (ϕ) (e.g. Costa, 2005; Caricchi et al., 2007; Costa et al., 2009; Mueller et al., 2010). At low crystal contents the parameterisations of Mueller et al. (2010) and Costa et al. (2009) are very similar, however as the Costa et al. (2009) model extends to higher crystal fractions we shall use their semi-

empirical parameterisation to calculate the relative viscosity of the crystal suspension:

$$\eta_r = \frac{1 + \phi^\delta}{[1 - F(\phi, \varepsilon, \zeta)]^{5/2\phi}} \quad (5)$$

Fig. 8. Strain rate ($\dot{\gamma}$) acting on SBS flows presented in Beckett et al. (2011). The solid black lines represent calculated strain rates from high order polynomial fits to measured velocity profiles (Fig. 10; Beckett et al., 2011). The solid horizontal grey line represents calculated strain rates from the volume flux of the flow, Q , see text for details.

Fig. 9. Capillary number (Ca) of the bubble suspension in ascending melt at Stromboli as a function of pressure, for conduit radii of $r_2 = 1$ m and $r_2 = 10$ m and bubble radii ranging from $R_b = 10^{-5}$ to 10^{-2} m.

$$F = (1 - \xi) \cdot \text{erf} \left[\frac{\sqrt{\pi}}{2 \cdot (1 - \xi)} \varphi \cdot (1 + \varphi^\zeta) \right]. \quad (6)$$

For a critical solid fraction (ϕ^*) the viscosity of the suspension tends to a constant value, where $\varphi = \phi/\phi^*$ and ξ, ζ, δ and ϕ^* are empirical parameters given by Caricchi et al. (2007) as a function of strain rate $\dot{\gamma}$ for suspensions of isotropic particles with ϕ ranging from 0.1 to 0.8.

The viscosities of the ascending and descending magma at Stromboli are shown in Fig. 10. There is no significant difference in the calculated viscosities of the ascending magma when using the crystals plus melt as the suspending fluid for bubbles compared to using the bubbles plus melt as the suspending fluid for crystals. At 350 MPa the viscosity of the ascending magma is $\mu_1 \sim 10^1$ Pa s and at 0.1 MPa $\mu_1 \sim 10^3$ Pa s, for a conduit with radius 1–10 m. The viscosity of the descending magma at a given pressure depends on the pressure at which the magma is assumed to change from ascending to descending (Fig. 3) as this affects the water content of the descending magma which in turn affects its crystallinity. For example, if the magma has descended from 0.1 MPa, then at 350 MPa $\mu_2 \sim 10^9$ – 10^{10} Pa s, but if it has descended from 75 MPa (the point at which the vesicularity remains at 50 vol.%) it has a viscosity of $\mu_2 \sim 10^2$ Pa s.

4.2. Magma density

The process of magma convection in the conduit is driven by the density difference between degassed crystallised magma at shallow levels and buoyant gas-rich magma deeper within the system. The densities of the ascending and descending magmas as a function of pressure can be calculated given the proportion of crystals, melt and bubbles (Fig. 11a). The density of the melt is taken to be 2700 kg m^{-3} (Richet et al., 2000), and the density of olivine is taken to be 3300 kg m^{-3} , clinopyroxene 3400 kg m^{-3} and plagioclase 2800 kg m^{-3} (Ahrens, 1995). The vesicularity of the magma was calculated in Section 3.1. As magma ascends through the conduit volatile exsolution drives an increase in vesicularity which decreases the density of the magma. At 75 MPa the vesicularity of the magma reaches 50 vol.%, and the magma is assumed to become permeable to gas flow (M.R. Burton et al., 2007). The magma continues to ascend (as it is less dense than the degassed magma) but its vesicularity now remains constant and any extra gas is lost from the system; its density now decreases only slightly due to changes in the density of the gas phase as a function of pressure. As magma descends back down the conduit after losing all of its gas, it continues to crystallise and its density increases with pressure. The calculated density difference between the ascending and descending magmas as a function of pressure is shown in Fig. 11b. Note that this local density contrast is different to the total vertical density difference across the length of the conduit, between the degassed magma at 0.1 MPa and the buoyant ascending magma at 350 MPa, which is $\sim 1500 \text{ kg m}^{-3}$.

5. Exchange flow

5.1. Volume flux and conduit radius

Following the method of M.R. Burton et al. (2007) the mass flux of magma required to account for the observed persistent gas flux at Stromboli of $200 \text{ t d}^{-1} \text{ SO}_2$ is 565 kg s^{-1} , taking the density of the melt to be 2700 kg m^{-3} . Using our calculated densities for the magma as it ascends through the conduit we can calculate the volume flux of magma required to achieve a mass flow rate of 565 kg s^{-1} (Fig. 12). As gas exsolves the density of the magma decreases and the volume flux of ascending magma increases with a decrease in pressure. Using Eq. (1) we can calculate the minimum radius of the conduit required to account for this volume flux, and hence the observed gas flux, as a function of depth in the conduit. Using our modelled fluid properties at 0.1 MPa suggests a conduit radius of 1.6 m (Fig. 13). This is similar

Fig. 10. a. Calculated viscosities of a. ascending magma at Stromboli comparing methods where the viscosity is calculated using the rheological model for the crystal suspension where the effective suspending fluid is the bubble suspension (Cx), and where viscosity is calculated using the rheological model for bubble suspensions where the effective suspending fluid is the crystal suspension (Bb). b. Calculated viscosities of descending magma (crystals–melt) at Stromboli from 0.1 MPa and 75 MPa at Stromboli.

Fig. 11. a. The calculated densities of ascending and descending magmas as functions of pressure at Stromboli, where density is a function of melt composition, vol.% crystals and vol.% bubbles. b. The density difference between ascending and descending magma as a function of pressure.

to that obtained by M.R. Burton et al. (2007) who employ the assumption of maximum flux, $r_2 = 1.25\text{--}1.45$ m, but for a larger mass flux of 575 kg s^{-1} , and to measured vent diameters at Stromboli of 4–8 m (Harris and Stevenson, 1997; Ripepe et al., 2007). The minimum required conduit radius increases with increasing pressure due to systematic changes in magma viscosities and densities; with the magma properties at 350 MPa, the calculated conduit radius is 12–17 m, although we do not expect the plumbing system to be a simple cylinder so deep in the system.

5.2. Flow regime

Assuming that the volume fluxes of the ascending and descending magma are matched, the Reynolds number of the more viscous fluid can be defined as:

$$Re_2 = \frac{Q\rho_2}{\mu_2 r_2}, \quad (7)$$

Fig. 12. The volume flux of ascending magma at Stromboli given that the mass flux of magma required to drive the persistent gas flux of $\sim 200 \text{ t d}^{-1}$ of SO_2 (Allard et al., 2008) is 565 kg s^{-1} . The increase in volume flux, as magma ascends, is due to the decrease in density driven by gas exsolution and decompression.

where ρ_2 and μ_2 are the density and viscosity of the more viscous fluid. In Fig. 14 calculated Re_2 of the experimental exchange flows presented by Beckett et al. (2011) and those in the studies of Huppert and Hallworth (2007) and Palma et al. (2011) are compared to Re_2 calculated for Stromboli based on the descending magma properties calculated in this study and the required volume flux of magma needed to account for the measured persistent gas fluxes. For an exchange flow at Stromboli both μ_2 and ρ_2 vary with pressure; consequently as pressure increases, Re_2 decreases and β increases. Over 0.1–350 MPa, for a conduit with radius 1–10 m, $Re_2 < 0.3$, suggesting that the flow will be laminar and can be well described by the low Reynolds number exchange flows presented by Beckett et al. (2011).

In the conduit of a volcano the growth of bubbles and crystals causes the viscosities and buoyancies of the magmas to change with pressure. In the analogue exchange flow experiments viscosities and buoyancy forces are independent of depth in the pipe. To apply the experimental results to model the flow regime of magma driving persistent degassing at Stromboli we calculate the viscosity ratio between the ascending and

Fig. 13. The minimum conduit radius required to account for the volume flux of ascending magma at Stromboli needed to drive the persistent gas flux of $\sim 200 \text{ t d}^{-1}$ of SO_2 (Allard et al., 2008). The radius as a function of depth is determined from Eq. (1) using calculated magma properties, where the viscosity of ascending magma has been determined for strain rates $\dot{\gamma} = 10^{-1}\text{--}10^{-4} \text{ s}^{-1}$.

Fig. 14. Flow regimes as a function of β and Re_2 of exchange flows observed by Beckett et al. (2011) ($r_2 = 19.2$ mm) compared to the experimental studies by Huppert and Hallworth (2007) (HH) ($r_2 = 4, 5.5, 8, 10, 12$ and 15 mm) and Palma et al. (2011) (PBC) ($r_2 = 10$ and 25 mm), where Re_2 is calculated using the known volume fluxes of the flow, Eq. (7). The solid lines represent exchange flow at Stromboli over 0.1 – 350 MPa, where $r_2 = 1$ – 10 m. As pressure increases in the conduit Re_2 decreases and β increases. All $Re_2 \leq 0.3$, exchange flow at Stromboli is laminar.

descending magma as a function of pressure in the conduit, using the results from the rheology calculations in Section 4. Given that calculated viscosities using the crystal suspension as the effective suspending fluid in the rheological model for the bubble suspension did not differ significantly from calculated viscosities using the bubble suspension as the effective suspending fluid in the rheological model from the crystal suspension, we just use the latter method to calculate viscosity ratios. Assuming that the magma descends from 0.1 MPa, at $P \geq 90$ MPa, $\beta > 100$ and the flow regime will be CAF, at $P \leq 90$ MPa, $\beta < 100$ and the flow regime will be SBS (Fig. 15a). In the experimental exchange flows Beckett et al. (2011) observed both regimes co-existing in the pipe separated by an abrupt transition (TSBS flow), with no step change in volume flux (compared to SBS flow at the same β), for $\beta = 5$ – 59 , suggesting that it is reasonable to expect that a flow regime transition could occur from CAF to SBS flow within the conduit. If the magma instead descends from 75 MPa (the point at which the vesicularity

of the magma remains at 50 vol.%), $\beta < 30$, due to the higher volatile content in the descending melt, and SBS will dominate below the level of overturn (Fig. 15b).

6. Discussion and conclusions

By constraining degassing and crystallisation paths at Stromboli using petrological and gas flux data we model the three-phase composition (crystals, melt, bubbles) of the ascending and descending magmas at Stromboli as a function of pressure. We calculate the rheology of this three-phase magma, accounting for its vesicularity, crystallinity and the composition of its melt, by employing recent studies on the rheology of crystal and bubble suspensions. We find that the viscosity of the ascending magma increases from $\sim 10^1$ to $\sim 10^3$ Pa s from 350 MPa to 0.1 MPa. The three-phase model of the viscosity of ascending and descending magma in conjunction with analogue exchange flow experiments results in a change in perspective of the flow regime of magma driving persistent degassing at Stromboli. We find that the viscosity ratio between the ascending and descending magma at Stromboli at $P \geq 90$ MPa (~ 3 km) is ≥ 100 and so we predict the flow regime to be CAF, but at $P \leq 90$ MPa (~ 3 km) the viscosity ratio is ≤ 100 and so the predicted flow regime is SBS. Although analogue exchange flow experiments show that the volume flux is independent of flow regime, the difference in melt velocity and viscosity distribution in the conduit between SBS and CAF could have important implications for the ascent of gas slugs driving explosive activity.

Regardless of the geometry of the exchange flow and conduit, the viscosity of the magma is highly dependent on its crystal volume fraction. Using a temperature of 1130 °C allows us to model the system well, resulting in crystallinities in accordance with those observed in the golden pumice and black scoria ejected at Stromboli, as discussed in Section 3.2. If we were to model the system with a lower temperature of 1115 °C, which is within the temperature range suggested from optical thermometry (Métrich et al., 2001), the crystallinity of the model system would increase. At 40 MPa we would predict that ~ 34 vol.% crystals and the transition from CAF to SBS flow would occur at a pressure of ~ 150 MPa.

We have assumed that the magma ascends to 0.1 MPa, before overturning and descending. If magma were to descend from 75 MPa (the point at which the vesicularity of the magma remains at 50 vol.%) it would have a lower crystallinity due to a higher water content in the melt, reducing the viscosity ratio between the ascending and descending magma such that SBS flow would dominate below the level of overturn. Further only $\sim 1/3$ of S would degas (Lesne et al., 2011), thus to supply the observed SO_2 flux the

Fig. 15. The viscosity ratio (β) between ascending and descending magmas at Stromboli considering overturn at a. 0.1 MPa and b. 75 MPa. Corresponding flow regimes (CAF and SBS flow) are shown, separated by the solid vertical line. Curves are shown for $\dot{\gamma} = 10^1$ and $\dot{\gamma} = 10^{-4}$, representing conduit radii of 1 and 10 m respectively. The grey box in a. indicates the range of pressures where a change in regime is predicted for the suite of curves shown.

magma flux would have to increase by a factor of 3, and the radius of the conduit would increase to 2–23 m (0.1–350 MPa).

We assume that the gas phase is entirely lost before the magma descends, and thus no bubbles are transported in the descending magma and the magma becomes increasingly undersaturated as it descends. Diffusion of volatiles from volatile-rich ascending magma into the degassed descending magma would reduce the crystallinity of the descending magma, as the presence of volatiles reduces the liquidus temperature. To consider whether this process is significant we consider a characteristic length scale for diffusion given by $\sqrt{D t}$, where D is diffusivity ($\text{m}^2 \text{s}^{-1}$) and t is time (s). Typical diffusivities for H_2O in a basaltic melt are on the order of 10^{-9} – $10^{-11} \text{ m}^2 \text{s}^{-1}$ for 1–3 wt.% H_2O at 0–100 MPa and 1300 °C (Persikov et al., 2010). We estimate the time scale for diffusion by considering how long it takes the magma to travel 1 km in the conduit. The velocity of the magma is $\sim 10^{-1} \text{ m s}^{-1}$, based upon our calculated $Q \sim 10^{-1} \text{ m}^3 \text{s}^{-1}$ for Stromboli (Fig. 12), and that the radius of the ascending fluid is $\geq 0.2r_2$ (Beckett et al., 2011). Thus for 1 km of flow in the conduit we have $t = 10^3 \text{ s}$ ($1000 \text{ m}/0.1 \text{ m s}^{-1}$) for diffusion, and so the length scale over which diffusion would occur is of the order 10^{-3} – 10^{-4} m . To allow homogenisation of volatiles despite such a short diffusion length scale would therefore require physical mixing between the volatile-rich ascending magma and the volatile poor descending magma. Using our calculated magma properties we showed that all $Re_2 < 0.3$, indicating that exchange flow at Stromboli is laminar (Fig. 14). We conclude that there will be very little physical mixing, and the length scales over which diffusion will occur in an exchange flow will be insignificant. However, a further observation from the analogue experiments presented by Beckett et al. (2011) is that a proportion of the less viscous fluid is dragged downwards with the more viscous fluid at the fluid interface. This could account for recycling of magma within the conduit as evidenced by mineral dissolution and reverse and/or oscillatory zoning of crystals in the pumice and scoria emitted at Stromboli (e.g. Landi et al., 2004; Francalanci et al., 2005; Landi et al., 2008).

The empirical relationship given by Beckett et al. (2011) is used in this study to constrain the radius of the conduit at Stromboli (Eq. (1)) from the magma volume flux based on SO_2 emissions and calculations of the three-phase magma viscosities and densities. However, it is unlikely that the volume flux and magma properties, which will vary temporally, solely control the conduit radius. Instead this represents only the minimum radius required to account for the gas fluxes observed. M. Burton et al. (2007) note that the ‘quiescent gas release between the explosions has a well-defined, quite steady mean composition’. However, if the bubble and/or crystal fraction in the ascending magma and hence the viscosity ratio between the ascending and descending magmas varies in time, then the depth of transition from SBS to CAF will also vary.

In the analogue experiments the volume flux of the ascending fluid is constrained to be equal to the net downwards flux of the more dense, more viscous fluid. Thus Eq. (1) also implicitly assumes that the volume fluxes (Q) of the ascending and descending magmas are matched. Given that the vesicular ascending magma is less dense than the degassed descending magma, matching the volume flux means that the mass flux of the ascending magma is less than that of the descending magma. This is partly, but not entirely, offset by the several weight percent of volatiles that are released at shallow levels and do not descend with the degassed magma. Therefore matching of volume flux would drive the free surface level of the magma in the conduit to fall over time. Although a drop in the magma level is reasonable over short time scales (e.g. Ripepe et al., 2002; Lautze and Houghton, 2007) clearly this is not representative of a system such as Stromboli which has been continuously active for 2000–2500 years (Rosi et al., 2000). This mismatch in mass fluxes could be resolved if the lateral interface position between the ascending and descending magmas and their velocities were to vary with pressure. The experimental studies of exchange flow have shown that the flow regime of an exchange flow

depends only on the viscosity ratio between the ascending and descending fluids (Stevenson and Blake, 1998; Huppert and Hallworth, 2007; Beckett et al., 2011), and so it may be reasonable to hypothesise that SBS flow with a transition to CAF would still be observed in this system.

In this study we have assumed a cylindrical conduit, which is clearly an oversimplification. Analysis of seismic signals indicates a transition from a shallow conduit to an inclined dike structure at a depth of ~ 2000 – 3000 m below the craters (Chouet et al., 2003). We note that the solution to the viscous flow equations in a slot is analogous to those in a pipe (Palma et al., 2011). It seems reasonable therefore to hypothesise, by analogy, that for the viscosity ratios ≥ 100 that we find for Stromboli magmas deeper than about 3 km, exchange flow in such dyke-like structures would, as in cylindrical pipe-like structures, be characterised by less viscous, vesiculating magma rising up the centre of the slot surrounded by more viscous, degassed and crystallised magma sinking along the dike walls. It would be a useful future experimental study to observe laminar exchange flow in a slot in order to test this hypothesis. A further complication that has yet to be explored experimentally is that several vents are persistently active at Stromboli which may be connected at some depth (e.g. Chouet et al., 2003). It would be interesting to establish whether individual conduits act independently or are affected by their neighbours, and whether the presence of multiple shallow conduits affects the geometry of the deeper exchange flow.

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