Mechanisms and dynamics of strombolian activity

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SUMMARY

Strombolian explosions at Heimaey and Stromboli are described. Two main components of activity within a typical strombolian explosion are distinguished: an initial, high velocity, gas thrust part due to gas decompression and a subsequent convective part. Initial gas velocities at Heimaey averaged 157 m/s (standard deviation 28 m/s from 15 observations) and at Stromboli 31 m/s (standard deviation 12 m/s from 8 observations) for one vent and 56 m/s for a second vent. Velocities decreased approximately exponentially with height, and decelerations of up to 50 gravities were observed during the gas thrust events. A model of the gas thrust process is developed and values are deduced for the gas/solid mass ratio in the ejected material. Evidence is presented for the several-fold concentration of gas into that part of the magma expelled explosively, and a model in which large bursting gas bubbles are responsible for the explosions is shown to be compatible with the observations. Excess pressure within such bubbles is found to be of order $2{\cdot}5 \times 10^4$ N/m (0·25 atmospheres) at Heimaey and 600 N/m^2 (0·006 atmospheres) at Stromboli. Pressures inside bubbles of a few metres diameter are found to be of comparable magnitudes. Average gas release rates of 3 to 6×10^3 kgm/s at Heimaey and at least 0·13 kgm/s at Stromboli are indicated.

STROMBOLIAN ACTIVITY consists of a series of discrete explosions separated by periods of less than 0·1 seconds to several hours. Although strombolian activity has been observed and described many times, there have been few attempts hitherto to quantify the dynamics of the explosions or to explain the principal features by a physical model.

Cine film (16 mm) of strombolian explosions was obtained at Heimaey on 1 February and 22 February 1973, and at Stromboli on 25 April 1975. A range of framing rates, from 8 to 64 frames per second, was employed to follow various details of the events.

It is convenient to define, specifically, the two main components of individual strombolian explosions: (a) the gas thrust part, in which a large deceleration occurred, and which occupied a time and height interval determined largely by the initial, maximum velocity, and (b) the convective thrust part, in which the rate of rise, and height reached, were determined largely by the bulk density and heat content of the erupted cloud at the end of the gas thrust phase.

1. Dynamics of explosions

At Heimaey, explosive activity took place simultaneously from three vents, with the steady effusion of lava. Explosions occurred at intervals of ½ to 2 seconds, though gaps of several tens of seconds occurred occasionally. Each explosion commenced with the rapid, near vertical rise of a cloud, consisting of a mixture of incandescent pyroclasts and gas. The mean initial gas velocity of 15 analysed bursts was 157 m/s, with standard deviation 28 m/s; the maximum recorded velocity was 230 m/s (Table 1). The rising mixture expanded by a factor of one to

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two in horizontal linear dimension during its subsequent, rapid deceleration to near-zero velocity. While the upper edge of the rising cloud could usually be traced easily on each cine frame, the sides of the cloud were harder to discern—possibly as a result of the partial incorporation of atmospheric air. The base of each cloud was even less well defined, due not only to the mixing of the cloud and the surrounding air but also to the emergence from the vent of some gas and pyroclasts at relatively low velocities late in the event. At the end of the period of rapid deceleration, gas and small particles of each cloud became mixed with air and commonly became incorporated into an eruption column driven, largely by convection, to heights of 6–10 km (Thorarinsson *et al.* 1973). Large blocks (diameters greater than 0·2 m) became decoupled from the erupted gas during its deceleration and pursued near-ballistic trajectories through the atmosphere, their motion being modified by air drag (Wilson 1972, Self *et al.* 1974).

Intermediate sized pyroclasts (diameters 0·05–0·2 m) were released near the end of the gas deceleration phase, and smaller particles (diameters 50–1 mm) were carried, by convection, to heights of 200 to 1000 m in the eruption column (Self *et al.* 1974).

Figure 1 shows the change of gas velocity with height in five typical explosions on Heimaey and clearly demonstrates the initial exponential decrease in velocity. The heights shown in Fig. 1 are measured above an arbitrary level (the lip of the vent) and the lava surface is estimated to be about 15 m beneath this level. It is considered that each explosion initially accelerates gas and particles to a maximum velocity due to decompression of the gas; the velocity then decreases exponentially. The maximum velocity is probably not very much more than that shown on the curves of Fig. 1, which only shows the deceleration periods. The exit velocities of bombs of 0·5–1 m diameter were 80–110 m/s (Self *et al.* 1974) and as the terminal velocities of such bombs (Walker *et al.* 1971) are of the order 60–100 m/s it is unlikely that velocities greatly in excess of 230 m/s were ever attained.

In Fig. 1 the five curves all appear to become asymptotic, with height, towards a constant value of 30–40 m/s. If the gas thrust phase was the only factor the curves should asymptote to zero. This does not occur because the gas/solid dispersion is less dense than the atmosphere and so convective uprise takes place. Thus Fig. 1 gives an indication of the velocities in the Heimaey cloud due to convection. The initial convective velocities of 30–40 m/s are equivalent to the terminal fall velocities of the largest clasts (diameters 1–10 cm) found in the scoria deposit 1–2 km from the vent (Self *et al.* 1974, Fig. 3a). These particles were observed to fall (Self *et al.* 1974) from the lower part of the convective part of the column. Self *et al.* (1974) also showed that most of the clasts had diameters between 10 and 1 mm and fell out at heights of 300 to 1000 m. The terminal velocities of these small particles (Walker *et al.* 1971) suggest convective velocities of 8–12 m/s at these heights, which is consistent with the expansion and deceleration of a convective cloud with initial velocity 30–40 m/s.

A further point emerges from Fig. 1, in that there is a considerable variation of vigour between individual explosions. This undoubtedly contributes to the sorting characteristics of scoria fall deposits. Particles with a given fall velocity will fall out over a range of heights and will hence vary in their final range.

At Stromboli, the character of the activity was noticeably different from that at Heimaey. Explosions occurred at intervals of 12 minutes, on average, from one or more of six vents. Two vents, referred to as vent 1 and vent 3, were studied. The duration of each individual explosion was from 3 to 10 seconds, and activity was similar to that described previously for Stromboli (Chouet *et al.* 1974). The two vents studied were very different in behaviour: vent 1 had a similar location and behaviour to the vent numbered 1 by Chouet *et al.* (1974) and vent 3 resembled their vent 5a in activity. Initial gas velocities at vent 1 averaged 31 m/s (maximum value 49·3 m/s) and at vent 3 averaged 56 m/s (maximum 65 m/s). The explosions

Table I: Water contents and pressures for explosions at Heimaey

	Gas Velocity (m/s)	Pressure (units of 10^5 N/m^2)	Water content (wt %)	Total mass ejected (units of 10^4 kgm)
(22nd Feb. 1973)	168.0	1.08	38.0	2.65
		1.28	12.0	3.0
	168.0	1.07	35.0	4.17
		1.22	12.0	4.42
	156.0	1.08	30.0	6.0
		1.22	12.0	6.74
	192.0	1.09	34.0	6.17
		1.27	13.0	7.5
(1st Feb. 1973)	96.0	1.04	25.0	3.95
		1.08	12.0	4.31
	153.0	1.09	24.0	6.9
		1.78	12.0	7.27
	102.0	1.04	20.0	3.72
		1.08	14.2	3.93
	140.3	1.05	32.2	3.32
		1.17	11.2	3.57

Two estimates of water content and pressure are shown for each explosion. The first is obtained from the cylinder model, and the second by assuming the cloud density is the same as the air density.

Note: "1st Feb. 1973" should be read opposite "153.0" column.

of longer duration were able to form a stable turbulent gas jet structure in the atmosphere. No permanent eruption column was formed, however, for explosions occurred so infrequently that the convective cloud of gas and small pyroclasts remaining after the gas thrust phase of each event was completely dissipated before the next event took place. Convective velocities of 4–8 m/s were estimated, taking gas and dust to heights of 150–250 m above the vent. Although most 'strombolian' eruptions are much closer in style to the near-continuous type of activity observed at Heimaey than to that of Stromboli itself, there is no reason to suspect any differences in the fundamental mechanisms of the individual explosions.

Figure 2 shows the detailed variations of velocity with height for explosion clouds at Heimaey and Stromboli. Superimposed on the generally rapid, approximately exponential decrease in velocity (Fig. 1) are high frequency fluctuations. Chouet *et al.* (1974) documented these fluctuations on Stromboli and it is interesting that similar phenomena occurred during the Heimaey explosions. Decelerations of 25 to 60 gravities occur. The fact that a stable jet structure was not formed in the case of most explosions, and that the maximum velocity of the gas/dust cloud was much less than the speed of sound in either the gas (830 m/s for steam at about 1200°K) or the surrounding air (about 330 m/s), suggests that a theoretical model based on the deceleration of the cloud treated as an incompressible body may be applicable.

2. Theoretical models

In view of the difficulty of identifying the lower edge of the rising cloud, two versions of the model are developed: one treats the motion of a spherical cloud of

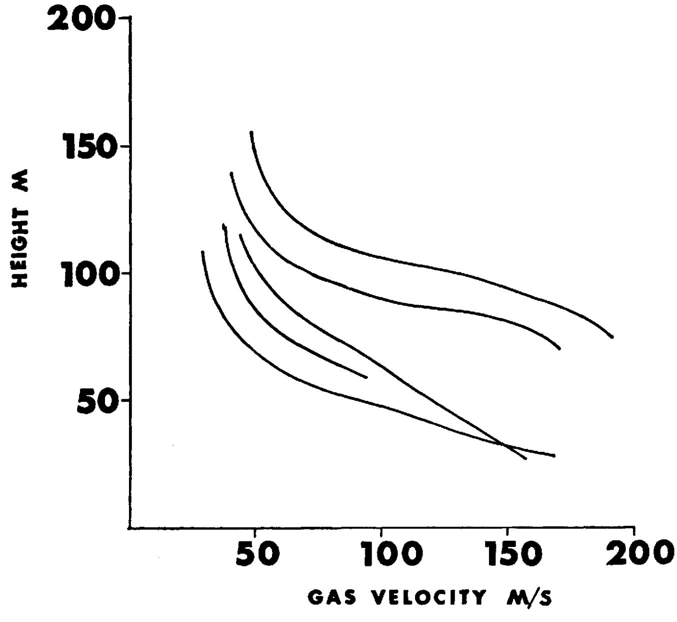

FIG. 1. Smoothed variation of gas velocity with height for five typical explosions at Heimaey. Heights are measured above the lip of the vent.

radius R, and the other that of a circular cylinder of radius A and length L. Let the speed of the cloud be U at height H above the point of maximum velocity (approximately the vent) and let its bulk density be σ. The initial, maximum velocity of the cloud is U_0 at $H = 0$ and the density of the surrounding air is ρ. After a little manipulation, the equation of motion can be written:

$$U\frac{dU}{dH} = -\left(\frac{\sigma-\rho}{\sigma}\right)g - \frac{CU^2}{\sigma X} \qquad (1)$$

where g is the acceleration due to gravity, C is the drag coefficient (approximately equal to unity) and X is given by:

$$X = \begin{cases} 2L, & \text{cylinder model} \\ \frac{8}{3}R, & \text{sphere model} \end{cases} \qquad (2)$$

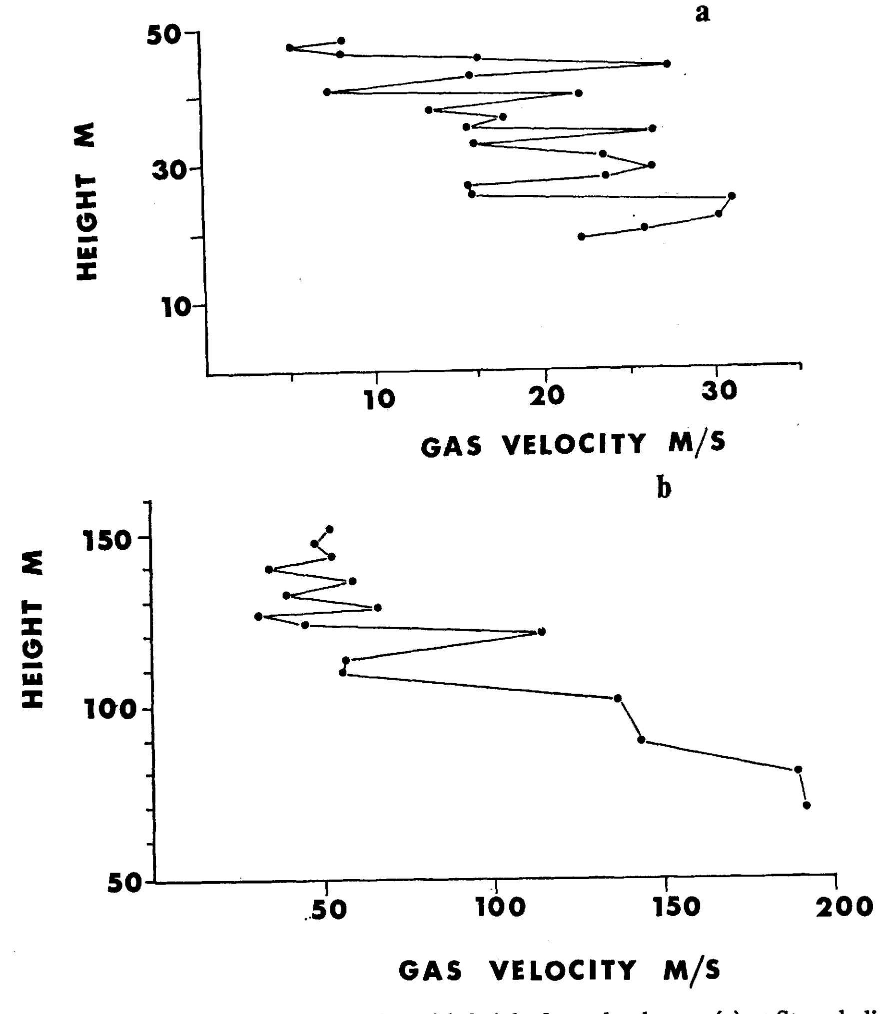

FIG. 2. Detailed variation of velocity with height for 2 clouds, one (a) at Stromboli and the other (b) at Heimaey.

Integrating once gives:

$$U^2 + Xg\frac{(\sigma - \rho)}{\rho} = \left[U_0^2 + Xg\frac{(\sigma - \rho)}{\rho}\right] \exp\left(\frac{-2H\rho}{X\sigma}\right) \tag{3}$$

for the variation of U with H. This equation was fitted to the velocity/height data obtained from Heimaey and Stromboli explosions to yield the bulk cloud densities listed in Tables 1 and 2. For each explosion, the spherical cloud model was

Table 2: Water contents and pressures for explosions at Stromboli on 25th April 1975

	Gas Velocity (m/s)	Pressure (units of 10^5 N/m^2)	Water content (wt. %)	Total mass ejected (kgm)
Vent 1	28.12	1.002	36.0	6.2×10^2
		1.005	12.0	12.2×10^2
	31.1	1.003	24.0	3.8×10^3
		1.006	11.0	5.99×10^3
	45.9	1.006	28.2	1.32×10^3
		1.015	12.0	1.67×10^3
	49.3	1.007	29.2	1.89×10^3
		1.017	12.0	2.2×10^3
	32.6	1.008	18.2	1.4×10^3
		1.012	12.0	1.55×10^3
Vent 3	65.35	1.017	21.2	33.6
		1.03	12.2	35.0
	60.9	1.016	20.2	61.9
		1.026	12.0	65.6
	63.9	1.01	34.2	16.3
		1.028	12.2	17.5

Two estimates of water content and pressure are shown for each explosion as in Table 1.

assumed to represent the extreme case; and a range of values L, constrained by careful examination of the film, was inserted into the cylinder model. Also given in Tables 1 and 2 is the solution corresponding to the assumption that the cloud density is equal to the air density. This certainly represents an upper limit to the cloud density, since all the clouds analysed were observed to convect upwards at the end of the gas thrust phase, implying a density less than that of the surrounding air.

The bulk cloud density can be related to the gas/solid ratio in the explosively erupted material. It may be assumed that the gas and pyroclasts forming the cloud are accelerated to their observed maximum velocities by the adiabatic decompression of magmatic gas. If P is the total gas pressure prior to expansion, N the weight percentage of gas (assumed to be steam with gas constant $Q = 461$ joule $kgm^{-1} K^{-1}$ and specific heat ratio $\gamma = 1{\cdot}3$) in the erupted cloud, T and magma temperature (taken as 1300°K) and P_A the atmospheric pressure, then energy conservation dictates that

$$U_0^2 = \frac{2NQT\gamma}{(\gamma - 1)}\left[1 - \left(\frac{P_A}{P}\right)^{[(\gamma-1)/\gamma]}\right] \tag{4}$$

The initial pressure, P, can be related to the final bulk density of the cloud, σ, by

$$P = \left[\frac{(QTN\sigma)^{\gamma}}{P_A}\right]^{[1/(\gamma-1)]} \tag{5}$$

where a very small term due to the finite volume of the solids in the cloud has been neglected. Equations (4) and (5) can be solved analytically for P and N as a function of σ if γ is assumed to be constant; however, during the adiabatic expansion the gas temperature falls, and γ changes appreciably with temperature. An iterative procedure was, therefore, used to solve the two equations, and values of P and N are given for each explosion in Tables 1 and 2.

It has been assumed, in this treatment, that the magmatic gas is water; if all the erupted gas were carbon dioxide, the values of N in Table 1 would be doubled, approximately, but the implied initial pressures would not change appreciably. Also, it has been assumed that the atmospheric air around the clouds has not been heated or contaminated by the erupted material. Heating of the surrounding air would decrease its density; the effect of adding magmatic gas would depend on the composition of that gas. If the density assumed for the atmosphere were half the normal value, then the values of N in Table 1 would, again, be doubled, and the values of excess pressure (above that of the atmosphere) would also be doubled, approximately.

3. Mechanisms of strombolian activity

The dynamic data presented above must place considerable constraints on any model of strombolian activity. Of particular interest is the mass ratio of solids to gas deduced from individual explosions. In general, the weight percentage of gas, irrespective of the species, could not possibly be dissolved within the quantity of magma expelled with the gas. The Heimaey magma would only dissolve 15–20% water, for example, at a depth of several tens of kilometres. It has been found that bubbles nucleate in many liquids, including silicates, at solute supersaturation pressures of only 10^5–10^7 N/m² (Gale 1966, Budd *et al.* 1962). Therefore high percentages of gas could not remain dissolved in the magma during its ascent. The high ratios of volatiles to ejected pyroclasts are therefore considered to be related to the mechanism of gas release driving the explosions. On Stromboli, mass ratios of gas to solids can reach very high values (Chouet *et al.* 1974), and rare explosions involve almost exclusively gas.

The most plausible explanation of the explosions is that they represent bursting of large bubbles at the surface of the magma. Thus, the accompanying pyroclasts are accidental in the sense that they are simply parts of the magma, neighbouring the large bursting bubbles, which are torn apart in the resulting expansion. On Heimaey, bubbles of several metres diameter were observed updoming the lava surface when the lava level was high. On Stromboli, normal degassing was accompanied by 'roar and rush' sounds similar to those accompanying outlet of gas under pressure. Each sound was followed by a puff of gas and, occasionally, ejection of pyroclasts. This behaviour is interpreted as the bursting of sizeable bubbles on the lava surface, as has previously been interpreted for lava lakes (MacDonald 1972).

Magma at various stages of degassing may be ejected by the bursting bubbles and hence form pyroclasts with a variety of densities and vesicularities. Relatively degassed epimagma may be torn off accidentally to form the large, dense, incandescent bombs (or may effuse as lava). The presence of highly vesicular tephra having vesicles down to a few millimetres suggests that the large bubbles envisaged are also rising alongside vesiculating pyromagma. Gas rich pyromagma vesiculates and is torn apart by the explosions to form the highly vesicular scoria. In situations such as Stromboli or Nyiragonga, where there is little or no lava effusion, each large explosion is envisaged as the bursting of a large bubble or group of bubbles passing through the static column of magma. The pyroclasts are accidental portions torn off by the bursting bubbles.

The presence of large bubbles, implied by the above evidence, indicates some complexities in the degassing of magma, which is hardly surprising considering the great range in solubilities of common volcanic gases. It is suggested that the poorly soluble gases, notably CO_2 and sulphur-compounds, may be important in this respect. Even small quantities of CO_2 for example would become supersaturated at several kilometres (Kadik & Lukanin 1973) and bubbles could nucleate. These bubbles would have time to increase their size by growth, by incorporation of more soluble gases at higher levels and by coalescence and could form bubbles of appropriate size.

Moore (1965) presented evidence to show that vesicles were present in Hawaiian submarine basalts at water depths of 4·5 km and that they increased in size with decreasing water depth. However the weight percentage of H_2O, the main volatile component, remained constant until depths of less than 800 m, suggesting that H_2O was undersaturated beneath this depth. Killingley & Muenow (1975) demonstrated that Moore's data could be interpreted as representing deep level nucleation of bubbles of CO_2 and sulphurous volatiles and that water started to exsolve on saturation at 800 m. Preliminary studies of bubble growth rates (Olmstead, Sparks, pers. comm.) in basalt suggest that metre sized bubbles must initially form at depths of at least a few hundred metres. For example, Moore (1965) records vesicles in submarine basalts of 1·0 mm diameter at 1 km water depth. Bubbles of such a size, if transported by magma or their own buoyancy to surface pressures, could grow up to metre size by diffusion and decompression in periods of 10^4–10^5 seconds (Gale 1966, Sparks, unpubl.). Coalescence of bubbles would make growth even more rapid.

The initial gas pressures of Table 1 are compatible with the behaviour of large gas bubbles near the surface. Internal pressures would, inevitably, be somewhat larger than hydrostatic pressures, due to the combined effects of surface tension, yield strength and viscous resistance to growth during the rapid inflation caused by gas decompression over the last few tens of metres rise.

The internal pressure of an expanding bubble in a viscous liquid can be determined by the following equation:

$$P_B = \frac{2s}{R} + P_H + 4\mu \frac{\mathrm{d}R}{\mathrm{d}t} \frac{1}{R} \tag{6}$$

where P_B is the bubble pressure, P_H is the hydrostatic pressure, μ and s are the kinematic viscosity and surface tension, R is the bubble radius and $\mathrm{d}R/\mathrm{d}t$ is the bubble growth rate.

Using $s = 30$ N/m (Murase & McBirney 1973) bubble pressures due to surface tension are negligible (1·2 N/m^2 for 0·5 m bubble). However, yield strengths of basic magmas are of the order of a few hundred N/m^2 (Shaw *et al.* 1968, Pinkerton & Sparks, pers. comm.). In order for growth to occur the bubble pressure must be in excess of the ambient pressure ($P_H + 2s/R$) by a value greater than the yield strength of the magma. This excess pressure would be comparable in magnitude to the initial pressures inferred for Stromboli (Table 2), but cannot account for those found at Heimaey. Once a bubble is growing an excess pressure due to viscous resistance $(4\mu/R)(\mathrm{d}R/\mathrm{d}t)$ occurs. It appears that this is the most important contribution to the formation of an excess bubble pressure over hydrostatic pressure. Bubble growth near the surface depends on the bubble rise velocity and hence on its diameter, as the rise velocity determines the rate of expansion of the bubble by decompression of the gas.

In the last few metres of a bubble's rise to the surface, expansion can be rapid. Table 3 shows the pressures due to viscous resistance predicted from equation (6) (which has been solved iteratively) in large bubbles on Heimaey and Stromboli where the dynamic viscosities of the anhydrous magmas are taken as 4×10^3 Nsm^{-2} (4×10^4 poise) (H. Pinkerton, pers. comm.) and 100 Nsm^{-2} (10^3 poise) respectively. Rise velocities were estimated using the empirical formulae of Grace (1974) at the appropriate Reynolds numbers. Table 3 also shows the total pressure within spherical and ellipsoidal bubbles breaking the magma surface. The hydrostatic pressure (P_H in equation 6) is taken to be that at the depth below the magma surface of the centre of the bubble. It is noted that the residual pressures of 0·5 to 4 m diameter bubbles are comparable in magnitude to the starting pressures indicated in the Heimaey and Stromboli explosions (Tables 1, 2). Bubbles with diameters less than *c.* 0·5 m would show negligible pressures due to viscous resistance.

For a given sized bubble a higher viscosity produces slightly higher residual pressures. However, the overriding control of the residual pressure appears to be the bubble size. This suggests that the bubble diameters in the case of Heimaey were considerably larger than on Stromboli. ten metre diameter bubbles were observed on Heimaey (G. P. L. Walker, pers. comm.) and these bubbles (Table 3) would give residual pressures comparable with those indicated from the gas velocities. The gas mass associated with each explosion on Heimaey is about 50 times greater than on Stromboli, perhaps indirectly suggesting much larger dimensions of the bubbles and explaining the higher starting pressures and consequent gas velocities.

On Heimaey the vent diameter was a maximum of 20 m during the period observed. If the mass of gas associated with each explosion is incorporated into a single bubble then it is found that many of these bubbles would occupy the whole diameter of the vent as they approach the surface. However, these bubbles would only expand to the vent diameter at 12–15 metres depth.

The energy needed to disrupt the magma into clasts has been neglected in the above discussion. Also, the mass, of solids indicated in Tables 1 and 2 does not include fragments with diameters greater than 0·1 m which quickly become dissociated from the motion of gas. Several tens to a few hundred bombs in the range 0·1 to 3 m were observed in individual explosions, but the majority did not clear the crater rim and were recycled in subsequent explosions. Both these factors

Table 3

Pressures in bubbles just reaching the magma surface. Values are given for spherical bubbles, and for ellipsoidal bubbles with the same volume, circular horizontal section and depth equal to one third the horizontal diameter (in an attempt to allow for bubble flattening near the surface).

Spherical bubble diameter (m)	pressure due to viscosity		total pressure in spherical bubble		total pressure in ellipsoidal bubble	
	Heimaey	Stromboli	Heimaey	Stromboli	Heimaey	Stromboli
	(All pressures in units of 10^5 N/m^2)					
0.101	0.00006	0.00005	1.0222	1.0221	1.0065	1.0064
0.496	0.0020	0.0004	1.0770	1.0754	1.0332	1.0320
0.970	0.0059	0.0009	1.1472	1.1422	1.0676	1.0626
1.88	0.0145	0.0016	1.2749	1.2620	1.1344	1.1212
4.34	0.0360	0.0024	1.6278	1.5942	1.3127	1.2783
7.97	0.0545	0.0027	2.1496	2.0978	1.5579	1.5104
11.17	0.0670	0.0027	2.5568	2.4925	1.7784	1.7140

would indicate rather higher residual pressures than those listed, because a proportion of the energy obtained from decompression would be used in tearing magma apart and in accelerating the large bombs.

4. Discussion

The model proposed suggests that gas segregation must start at considerable depth, probably with nucleation of poorly soluble gases. These nuclei may act as preferential seeds for the diffusion of more soluble and perhaps more abundant gas species. Large bubbles grow and migrate through the rising magma leaching out gas. The strombolian explosions are largely the result of bursting of individual bubbles or clusters of bubbles. The magma through which the bubbles rise is probably composed of pyromagma, epimagma and magma, at intermediate stages of degassing. Pyromagma is probably more fluid due to dissolved gases and the large bubbles may be preferentially associated with pyromagma. Vesiculation of the highly soluble gases from pyromagma near the surface may account for highly vesicular tephra. The large bombs are fragments of magma in various stages of degassing, ranging from highly inflated bombs to dense, poorly vesicular bombs.

The bubbles, on reaching the surface, are thought to have a residual pressure largely resulting from the viscous resistance to rapid expansion. The decompression of the gas from the bursting bubbles results initially in high gas velocities, rapidly decelerating as the solid/gas dispersion interacts with the atmosphere. The finer grained particles are transported to much greater heights by convection. The physical models discussed allow estimates of the total quantities of gas released and of the rates of release. On Heimaey, the gas release rate is estimated at $3–6 \times 10^3$ kgm/s which compares with $7{\cdot}5 \times 10^4$ kgm/s for the magma (Self *et al.* 1974) and indicates a 'gas content' of 4–8%. These quantities do not imply the amount of gas dissolved in the magma, because these estimates apply to the first few weeks of eruption. If large bubbles are forming at depth and rising through the magma during the eruption then a 'concentration' of a gas would be expected in the upper part of the magma, in addition to any previous equilibrium distribution attained due to gas diffusion under a gravitational field, as predicted by Kennedy (1955).

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