Geophys. J. Int. (1995) **121**, 226–232

Explosive volcanic eruptions—IX. The transition between Hawaiian-style lava fountaining and Strombolian explosive activity

E. A. Parfitt[1] and L. Wilson[1,2]

[1] *Geological Sciences Department, Brown University, Providence,* RI 02912, *USA*
[2] *Environmental Science Division, Institute of Environmental and Biological Sciences, Lancaster University, Lancaster* LA1 4YQ, *UK*

Accepted 1994 September 12. Received 1994 September 12; in original form 1994 January 14

SUMMARY

The commonest eruption styles of basaltic volcanoes involve Hawaiian lava fountaining or intermittent Strombolian explosions. We investigate the ways in which magma rise speed at depth, magma volatile content and magma viscosity control which of these eruption styles takes place. We develop a model of the degree of coalescence between gas bubbles in the magma which allows us to simulate the transition between the two extreme styles of activity. We find that magma rise speed is the most important factor causing the transition, with gas content and viscosity also influencing the rise speed at which the transition occurs. Counter to intuitive expectations, a decrease in gas content does not cause a transition from Hawaiian to Strombolian activity, but instead causes a transition to passive effusion of vesicular lava. Rather, a change from Hawaiian to Strombolian style requires a significant reduction in magma rise speed.

Key words: explosive eruptions, Hawaiian, Strombolian.

INTRODUCTION

Basaltic volcanic activity can vary widely in character from passive effusion through Strombolian explosions and Hawaiian-style lava fountaining to sub-Plinian and Plinian eruptions. The exact style of activity that occurs is expected to depend strongly on the volatile content of the magma and the magma rise speed (Wilson & Head 1981, 1983; Wilson 1984). The two extremes of basaltic activity, passive effusion and Plinian explosion, result from the extremes of gas content—very low gas contents leading to effusion of somewhat vesiculated lavas and very high gas contents (either due to unusually high primary gas contents or, more commonly (e.g., Walker, Self & Wilson 1984) due to the interaction of the magma with restricted amounts of groundwater—Wohletz 1986; Kokelaar 1986) causing Plinian activity. While these extremes of behaviour are certainly of interest, the volatile contents of most terrestrial basaltic magmas are such that basalts are commonly erupted explosively in Strombolian or Hawaiian-style activity (Macdonald 1972; Basaltic Volcanism Study Project 1981). In this paper we discuss the explosive eruption of basalts with typical terrestrial gas contents, consider the influence on eruption character of magma rise speed at depths below the level where significant volatile exsolution begins, and develop a model of the transition from Strombolian to Hawaiian eruption styles.

BACKGROUND

The basic processes that give rise to Strombolian and Hawaiian activity have been investigated by a number of authors and are relatively well understood (Macdonald 1972; Chouet, Hamisevicz & McGetchin 1974; Blackburn, Wilson & Sparks 1976; Wilson & Head 1981; Head & Wilson 1987). Magma rising towards the surface on Earth usually has sufficient volatiles dissolved within it to become supersaturated in one or more volatile species before it reaches the surface. When rising magma exsolves volatiles to form gas bubbles, the bubbles are always buoyant relative to the surrounding liquid, and they rise through it at a rate that depends on their size and the magma viscosity. The sizes of existing bubbles increase as they approach the surface, due to a combination of decompression and continued volatile migration from the liquid (Sparks 1978), both processes being driven by the decreasing pressure. At the same time, new, small ($\sim 10\ \mu$m diameter) bubbles continue to nucleate and the magma becomes ever more supersaturated at lower pressures.

At some depth below the surface, the volume fraction of the magma that consists of gas bubbles becomes sufficiently high that the largest bubbles become closely packed and begin to deform. As thin liquid films between these bubbles collapse, the magma disrupts into a spray of gas and vesicular magma clots, this process taking place over a finite range of pressure conditions and corresponding depths below the surface. The gas volume fraction at which the process begins is commonly taken as 75 per cent (and we use this value in our treatment below), although vesicularities of pumice and scoria clasts imply that gas volume fractions in disrupting magmas may range from less than 60 to more than 90 per cent (Sparks 1978). In the region below the disruption level, larger bubbles rise faster than smaller bubbles and may overtake them. If two bubbles come into physical contact, which is a function of the horizontal component of their initial separation and their sizes, they are likely to coalesce (Wilson & Head 1981; Toramaru 1988; Manga, Stone & O'Connell 1993). The exact extent to which bubble coalescence can take place in an erupting magma is a complex function of its rise speed, viscosity and total volatile content (Wilson & Head 1981). It is the degree to which coalescence occurs that essentially differentiates between the Strombolian and Hawaiian styles of basaltic explosive eruptions.

At high magma rise speeds, small, freshly nucleated gas bubbles are unable to move appreciably relative to the magmatic liquid in the time interval available between their nucleation and the magma's arriving at the surface vent, and thus they stay strongly coupled to the motion of the liquid. Little bubble coalescence occurs and the residual gas–magma mixture will reach the 75 per cent exsolved gas condition and disrupt at a relatively deep level and high pressure. Subsequent rise of the disrupted gas–pyroclast mixture will lead to considerable gas expansion to lower pressure levels, resulting in a great acceleration of the mixture and a high eruption velocity. This produces a steady discharge of liquid clots and released gas, forming a lava fountain in its truest sense (the situation treated by Head & Wilson 1987). At low or zero magma rise speed, gas bubbles are able to move by significant distances through the magma, and as they rise they coalesce and form larger bubbles. The residual mixture is thus strongly depleted of gas and, at low enough rise speeds, will not reach the 75 per cent gas volume fraction disruption condition. A pond of vesicular lava forms in the vent; the bursting of the large bubbles at the surface of the pond gives rise to classic Strombolian activity (Blackburn *et al.* 1976). For intermediate magma rise speeds, the behaviour of the gas–magma mixture is more complex, with a range of large bubbles formed by coalescence of smaller bubbles rising through the residual mixture of magmatic liquid and smaller bubbles that have not been involved in coalescence events. This residual mixture is thus relatively depleted of gas and will only reach the 75 per cent gas volume fraction disruption condition at a relatively shallow level. Not only is the mass fraction of gas causing disruption smaller in this case, but also the lower pressure at which disruption takes place means that less gas expansion can occur to drive the acceleration of the gas–pyroclast mixture, and eruption velocities will be lower. Although some energy is potentially available from the expansion of the gas released from the large bubbles, this will not be very strongly coupled to the motion of the pyroclasts formed by disruption of the residual mixture, and so will not add significantly to the maximum height reached by the pyroclasts.

Figure 1 shows the theoretical relationship between gas bubble diameter and magma rise speed for a magma having

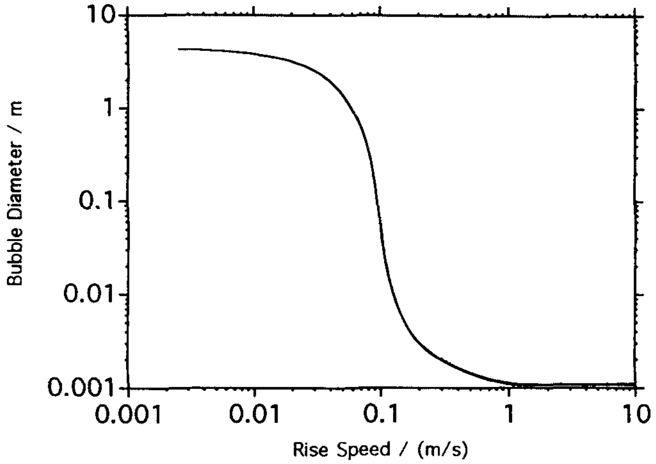

Figure 1. The variation with magma rise speed of the diameter of the largest bubble that can be formed by coalescence of smaller bubbles, in a magma with viscosity 30 Pa s and total initial volatile content 0.5 wt per cent; the volatile is assumed to consist entirely of H_2O.

a total gas (H_2O) content of 0.5 wt per cent and a (Newtonian) viscosity of 30 Pa s. The values plotted in this figure were computed using a computer program described by Wilson & Head (1981), which was based on an earlier numerical model developed by Sparks (1978), and illustrate the range of gas bubble behaviour which reflects the change from lava fountaining to Strombolian activity. The transition from negligible bubble coalescence and lava fountaining (small bubbles, high magma rise speed) to dominant coalescence and Strombolian activity (large bubbles, low magma rise speed) is seen to be fairly abrupt and to occur at a rise speed of $\sim$0.1 m s^{-1}. The rise speed at which this transition occurs and the abruptness of the transition are a function of both gas content and magma viscosity, as shown in Fig. 2 (which is based on the same computations as Fig. 1). The range of gas contents (0.25–1 wt per cent) and viscosities (30–300 Pa s) used in Fig. 2 covers typical ranges for terrestrial basalts (Basaltic Volcanism Study Project 1981); thus on the basis of these results we can see that the transition from Strombolian to Hawaiian activity typically occurs at rise speeds of 0.01 to 0.3 m s^{-1}. Fig. 2 shows that higher gas contents cause the transition to Strombolian activity to occur at high rise speeds. This is because the high number density of gas bubbles causes significant coalescence, and hence Strombolian behaviour, even when the magma is rising relatively rapidly. It is apparent from Fig. 2 that lower magma viscosities have a similar effect, because the mobility of the bubbles through low-viscosity magma is high and thus bubble coalescence can occur even when the magma itself is rising at a relatively high rate. Low gas contents and, to a lesser extent, higher viscosities cause the transition from Strombolian to Hawaiian activity to be more gradual (Fig. 2). Although these plots allow us to define where the transition between the two styles of activity will occur, current models of these two styles (Blackburn *et al.* 1976; Wilson & Head 1981; Head & Wilson 1987) only permit us to simulate activity which is strictly of one style or the other (i.e. either well to the left or well to the right in Figs 1 and 2). In the next section, we present the results of a model of the bubble coalescence process which allows the

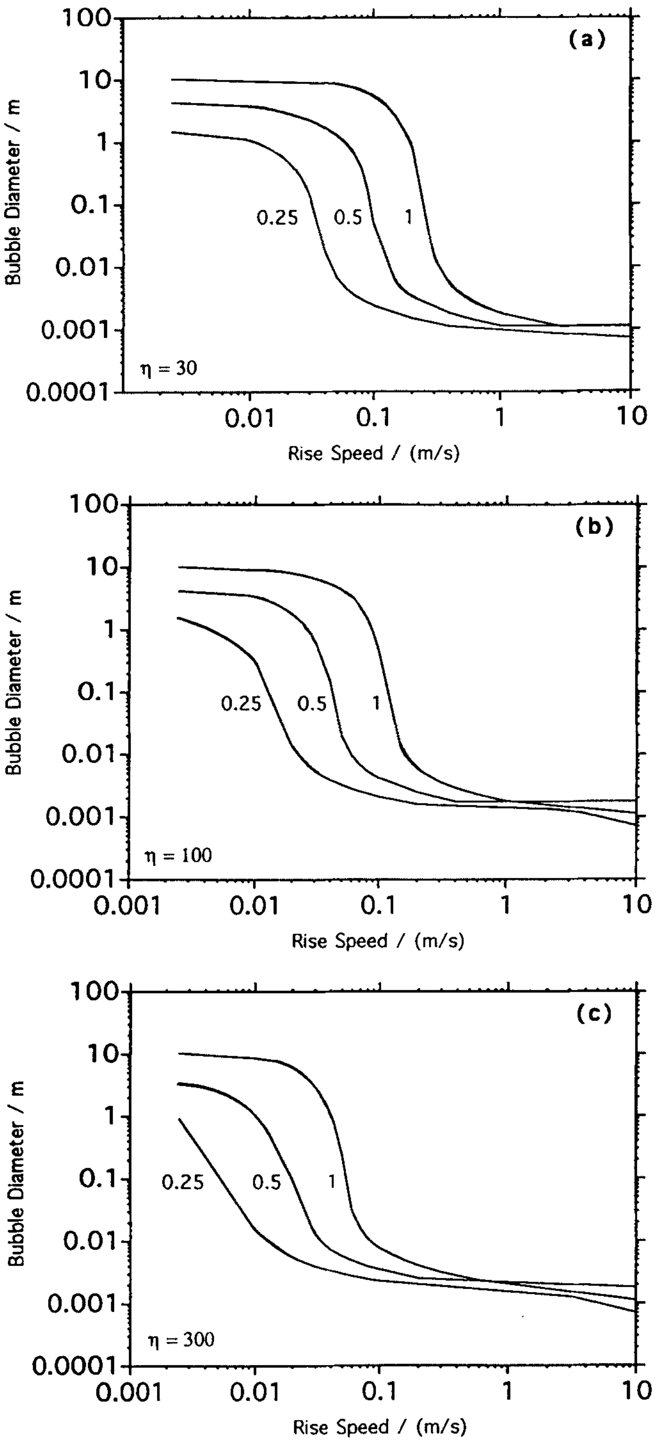

Figure 2. The variation with magma rise speed of the diameter of the largest bubble that can be formed by coalescence of smaller bubbles in a rising magma. Curves are given for three values of the initial total magma water content (0.25, 0.5 and 1 wt per cent) and viscosities of (a) 30, (b) 100 and (c) 300 Pa s.

transitional eruptive behaviour to be simulated in more detail.

BUBBLE COALESCENCE MODELLING

We summarize here the earlier work of Sparks (1978) and Wilson & Head (1981) and then describe the additions that we have made to their computer algorithms. Sparks (1978) began his analysis with the evidence that bubbles will nucleate in a rising magma when any volatile species is supersaturated by a small amount (~0.1 MPa) and that initial bubble sizes are a few microns. He used experimental data on the solubility of common volatiles such as H_2O and CO_2 as a function of pressure and temperature to define the depth at which bubble nucleation begins. Each bubble is buoyant in the surrounding liquid and rises through it at a rate that depends on the size and shape of the bubble and the difference between the density of the gas in the bubble and the density of the surrounding liquid. Information on the dependence of the fluid-dynamic drag coefficient on the Reynolds number was used which takes account of the complex change in bubble shape (from near-spherical through flattened to spherical cap) which occurs as a bubble grows and its rise speed increases. Bubble growth is a result of decompression (which was assumed to be isothermal, since the gas in the bubble is in excellent thermal contact with the surrounding liquid) and also of the continued, concentration-gradient-dependent diffusion of the volatile species from the liquid into the bubble. Addition of the rise speed of the bubble through the liquid to the rise speed of the liquid through the surrounding crust allowed the variation of bubble properties with both time and depth below the surface vent to be found as a function of initial magma volatile content and magma rise speed. The equations representing the above sequence of calculations were solved numerically with a FORTRAN program by Sparks (1978).

Wilson & Head (1981) expanded the above analysis by noting that, as an existing bubble, growing by decompression and diffusion, rose through the magmatic liquid, it was always moving relative to a spectrum of other bubbles: new 10 micron-sized bubbles that were being created as a result of the increasing supersaturation of the rising liquid, and older bubbles that had nucleated over a range of depths shallower than that at which the first nucleation events had occurred. By tracking numerically the volume of liquid swept by the largest existing bubble during each integration time-step, and allowing for the distortion of the streamlines in the liquid caused by its passage, it was possible to calculate the number of smaller bubbles with which it would come into physical contact. Assuming a high efficiency of coalescence whenever such an encounter occurred, the increase in volume of the oldest (and hence largest) bubble was found. It was thus possible to compute the size and rise speed of the largest bubble present at any depth below the surface. Wilson & Head (1981) added the necessary code to Sparks' (1978) computer program to accomplish the above calculations.

We have had to make only minor changes and additions to the above code, designed to allow us to keep track, as a function of depth below the surface, of the masses of the gas contained in the largest bubble and the gas distributed as smaller bubbles through the residual gas–liquid mixture. At whatever depth, d, this gas–liquid mixture attains an average gas volume fraction of 75 per cent (and thus disrupts into released gas and pyroclasts), values are recorded for the pressure, P_i, the exsolved gas mass fraction in the mixture, n, and the rise velocity of the magma, u_i. The final velocity u_f with which the gas–pyroclast mixture emerges at the surface vent is then found by integration of the momentum equation between the disruption level and the surface in the form discussed by Wilson (1980):

$$u_f^2/2 = u_i^2/2 - gd + [(nQT)/m] \ln (P_i/P_f) + [(1-n)/\rho_1](P_i - P_f). \quad (1)$$

Here, P_f is the pressure immediately above the vent (taken as being that of the atmosphere, ~0.1 MPa), ρ_1 is the density of the bubble-free magma (~2600 kg m^{-3}), m is the molecular weight of the volatile species (18 if H_2O predominates), Q is the universal gas constant (8.3145 J mol^{-1} K^{-1}) and T is the magma temperature (taken as 1400 K for a mafic magma).

There are several approximations involved in the above treatments. For example, energy losses due to wall friction have been neglected in deriving eq. (1). This is justified on the grounds that frictional energy losses are relatively small above the level at which the magma fragments, as shown by detailed numerical calculations (Wilson, Sparks & Walker 1980; Wilson & Head 1981; Dobran 1992a, b). Analysis of friction losses below the magma fragmentation level are avoided by expressing the results of the calculations in terms of the magma rise speed at depth and the magma viscosity, these quantities being treated as independent variables. In fact they are coupled, in that the rise speed is a function of the magma viscosity, the conduit width, and the driving pressure gradient. However, the pressure gradient is unlikely to be dramatically different from the lithostatic gradient in the country rocks (Giberti & Wilson 1990), and so choosing any pair of values of magma rise speed and viscosity effectively amounts to also choosing a conduit width and hence a particular mass eruption rate (Wilson & Head 1981). Finally, no attempt has been made to allow for the fact that exsolution of a volatile phase (especially water) from a magmatic liquid will increase its viscosity; we have neglected this because, although the resulting decrease in the rise speed of the magma through the conduit system will increase the time available for bubble motion and coalescence, this will, to a good approximation, be offset by the lower rise speed of the gas bubbles through the liquid.

We have run the computer program implementing the above model for magma rise speeds at depth in the range 0.001 to 1 m s^{-1}, magma viscosities of 30, 100 and 300 Pa s (McBirney & Murase 1984; Heslop *et al.* 1989) and total gas contents of 0.25, 0.5 and 1 wt per cent (representative values for terrestrial basaltic eruptions—Basaltic Volcanism Study Project 1981). In each case, we used the resulting exit velocity to calculate an *energy efficiency factor* (EEF), which we define as the square of the ratio of the actual eruption speed to the speed that would have been attained if no gas bubble coalescence had occurred (which is found by re-running the program with the same initial parameters, but with the coalescence process turned off). Because the height of a lava fountain is found by equating the kinetic energy per unit mass ($0.5\,u^2$) of the eruption products leaving the vent at speed u to the potential energy per unit mass (gh) they have at the top of the fountain of height h, the energy efficiency factor is also the ratio of the actual fountain height expected when bubble coalescence does take place to the fountain height that would have been reached by a magma with the same total volatile context but a sufficiently great rise speed at depth in its conduit system that coalescence did not take place.

We summarize the results in Fig. 3 by plotting the energy efficiency factor as a function of magma rise speed for three gas contents (0.25, 0.5 and 1 wt per cent) and three

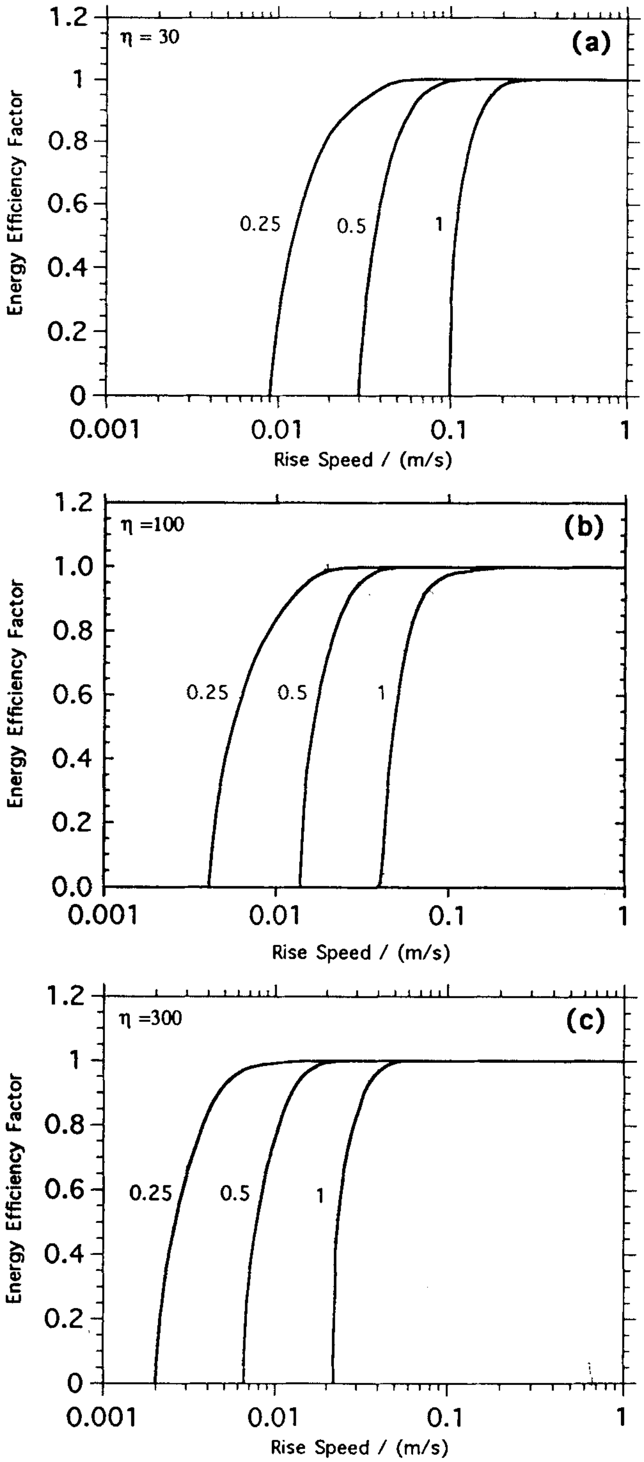

Figure 3. The variation with magma rise speed of the energy efficiency factor (see text for definition). Curves are given for three values of the initial total magma water content (0.25, 0.5 and 1 wt per cent) and viscosities of (a) 30, (b) 100 and (c) 300 Pa s. An energy efficiency factor of 1 denotes Hawaiian lava fountaining and a value of 0 denotes purely Strombolian activity.

viscosities (30, 100 and 300 Pa s). In Fig. 3, an EEF value of 0 indicates that the eruptive activity is purely Strombolian in character and an EEF of 1 indicates purely Hawaiian lava fountaining. For $0 < \mathrm{EFF} < 1$, the eruptive behaviour is transitional and the height of the resulting fountain can be calculated from the EEF value if the height of the fountain during pure fountaining is known (Wilson, Parfitt & Head 1995). For example, consider a magma with a total gas content of 0.5 wt per cent and a viscosity of 300 Pa s. At rise speeds greater than 0.016 m s^{-1}, the magma will erupt in a purely Hawaiian fashion and will produce a fountain of height 640 m (EFF = 1 in Fig. 3c). Below a rise speed of 0.0065 m s^{-1}, the eruption will be Strombolian (EEF = 0 in Fig. 3c). For intermediate rise speeds, for example 0.007 and 0.009 m s^{-1}, the EEF values will be 0.363 and 0.702, respectively, leading to fountain heights of 232 and 449 m.

APPLICATIONS OF THE MODEL

As we have discussed above, the transition from Strombolian to Hawaiian activity is most strongly influenced by variation in magma rise speed at depth, rather than changing gas content. As an illustration of the effects of increasing rise speed on eruptive behaviour, we consider a magma with a fixed viscosity of 100 Pa s and a fixed gas content of 0.5 wt per cent erupted at rise speeds in the range 0.0003 to 0.1 m s^{-1}. Fig. 3(b) shows that for rise speeds less than 0.014 m s^{-1} the eruptive behaviour is purely Strombolian (EEF = 0) and at rise speeds greater than 0.07 m s^{-1} the behaviour is purely Hawaiian (EEF = 1). At intermediate rise speeds, the behaviour is transitional between the two end members.

The height to which clasts will be thrown during Strombolian activity was calculated using eqs (3) and (4) in Blackburn *et al.* (1976) for magma rise speeds of 0.0003, 0.001 and 0.01 m s^{-1}. The Blackburn *et al.* (1976) model takes account of the fact that only part of the magma overlying a large bubble approaching the surface of the lava pond is expelled in a Strombolian explosion and so utilizes an apparent gas mass fraction in the explosion products. We used the average value, 20 wt per cent, found by Blackburn *et al.* (1976) to calculate the pyroclast launch speeds and hence rise heights. In all three cases, the bubble sizes (3.6 to 4.4 m) and exit velocities (242 to 263 m s^{-1}) are approximately the same, and thus the clasts all reach approximately the same height (~25 m). Although the heights reached by the clasts are all much the same, the frequency of explosions will increase as the magma rise speed increases. By calculating the mass of gas in a single large bursting bubble and knowing the total weight fraction of gas in the magma, we have estimated the approximate time interval between explosions on the basis of how long it would take to erupt the volume of magma originally containing the bubble gas volume. In this way we estimate that at a rise speed of 0.0003 m s^{-1} the time interval between explosions would be ~95 s; for a rise speed of 0.001 m s^{-1} it would be ~42 s; and at a rise speed of 0.01 m s^{-1} (close to the transition zone) the time interval between explosions would have decreased to ~4 s. Thus, as the rise speed of the magma increases within the Strombolian range, the height to which the clasts are thrown remains fairly constant and low, but the frequency of explosions progressively increases.

Once the rise speed exceeds 0.014 m s^{-1}, the eruptive behaviour becomes transitional between Strombolian and Hawaiian and we can calculate the height now attained by the clasts from the EEF plot (Fig. 3b). As the EEF values represent the ratio of the actual fountain height to that which the fountain would have had if no bubble coalescence had occurred, in order to obtain actual fountain heights we first need to calculate the maximum height of the fountain. From our computer model (see above) we found an exit velocity for a purely Hawaiian fountain of 110 m s^{-1}, which, using the ballistic relationship $u^2 = 2gh$, gives a maximum fountain height of 619 m. Using this value and the EEF values from Fig. 3(b) we calculated the variation in 'fountain' height within the transitional zone for this example. These results are summarized in Fig. 4, which show that once the magma rise speed increases beyond the critical value of 0.014 m s^{-1}, the height to which the clasts are propelled increases very rapidly. Although our model probably overemphasizes the rapidity with which this increase occurs (because at the lower end of the EEF range we should allow for air drag on the ejected clasts as we do in the true Strombolian case) the transition will still be relatively abrupt. At low EEF values, the height of the 'fountain' increases dramatically as the EEF increases because for the first time the gas-magma mixture reaches the 75 per cent gas disruption criterion. This means that, although there is still considerable bubble coalescence and bursting, this is now superimposed on an initially low but rapidly growing fountain. At low EEF values, the fountain is likely to fluctuate considerably in height due to the effects of the largest bubbles bursting; but as the magma rise speed increases, the amount of bubble coalescence declines and the height of the fountain increases and is also more stable at any given rise speed. Once the rise speed exceeds 0.07 m s^{-1}, the fountain attains its maximum height, fluctuations almost completely vanish and the activity is truly Hawaiian in nature.

Figures 3 and 4 show very clearly that magma rise speed

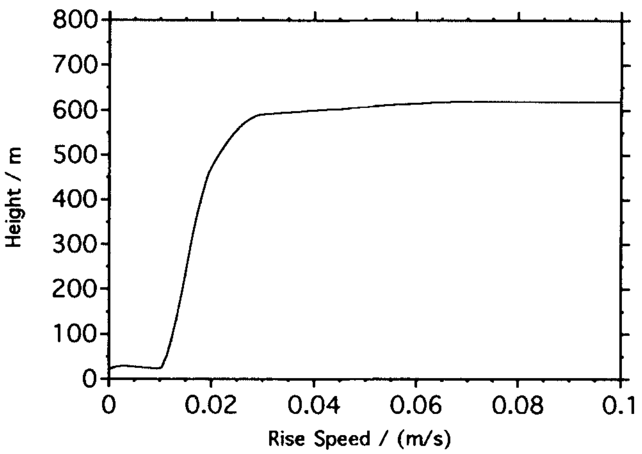

Figure 4. The variation with magma rise speed of the maximum height reached by pyroclasts formed in explosive eruptions of a magma with a total water content of 0.5 wt per cent and a viscosity of 100 Pa s. At magma rise speeds greater than 0.07 m s^{-1} the activity is purely Hawaiian; at speeds less than 0.01 m s^{-1} the activity is purely Strombolian and at intermediate rise speeds the activity is transitional between the two.

has a strong influence on whether an eruption is Strombolian or Hawaiian in nature. Fig. 3 can also be used to demonstrate that variation in gas content does not have a major impact on eruption style unless it is also accompanied by a change in rise speed, a somewhat counterintuitive result. Consider Fig. 5, which shows the transition from Hawaiian to Strombolian eruption styles as a function of magma rise speed and gas content. For any given magma rise speed, a reduction in gas content causes the eruption style to change from Strombolian to transitional to truly Hawaiian (i.e. the EEF increases). Some authors (e.g. Vergniolle & Jaupart 1990) have suggested that a decrease in the gas content should cause a change from Hawaiian to Strombolian eruption style. Although this latter pattern would seem to be intuitively correct, Fig. 5 shows that the opposite is actually true. This is because account must be taken of the greater degree of bubble coalescence that occurs in the cases with high gas content. For any given rise speed, a high gas content leads to more coalescence and therefore places the activity closer to the Strombolian end of the spectrum of eruption styles. As gas content decreases, the amount of coalescence is reduced and thus the eruptive activity trends towards the Hawaiian style. If an eruption were Hawaiian in nature initially and the gas content then reduced without any change in the magma rise speed, a point would eventually be reached at which the magma would no longer fragment and thus fountaining would stop. However, eruptive activity would then occur as passive effusion, *not* Strombolian activity, because magma rise speed would still be high enough to prevent gas bubble coalescence.

To demonstrate the importance of understanding the influence of magma rise speed rather than gas content on eruptive behaviour, consider the change in character from Hawaiian fountaining to Strombolian activity exhibited during the recent Pu'u 'O'o-Kupaianaha eruption of Kilauea Volcano, Hawaii (Wolfe *et al.* 1988; Heliker & Wright 1991; Parfitt & Wilson 1994). During the Pu'u 'O'o stage of the eruption, activity was intermittent and was associated with fountains as high as 450 m. In July 1986 the style of the eruption changed such that activity occurred continuously but was Strombolian in nature. Vergniolle & Jaupart (1990)

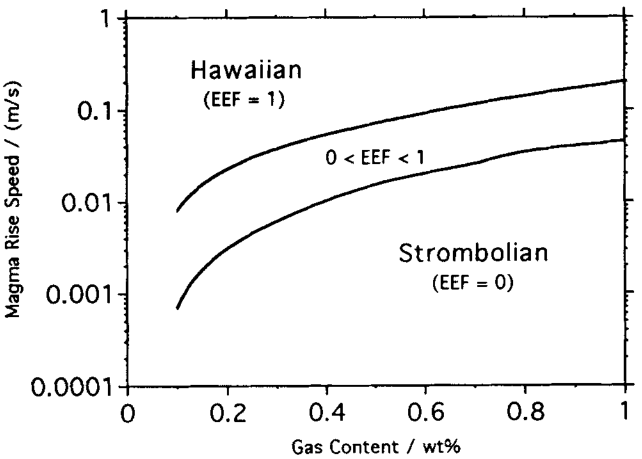

Figure 5. The variation in style of explosive basaltic eruptions shown as a function of magma rise speed and total magma water content.

explained the change in eruption character in terms of decreasing magma gas content which, as we have shown above, cannot in fact be the cause. Instead, this change in behaviour can be shown to be a direct result of a decline in eruption rate and hence magma rise rate at depth beneath the eruptive vent. Parfitt & Wilson (1994) calculated magma rise speeds of 0.3 and 0.01 $m\,s^{-1}$ for the Pu'u 'O'o and Kupaianaha stages of the eruption, respectively, using observed eruption rate values. The total gas content of the erupting magma has been estimated to be 0.30 to 0.45 wt per cent (Gerlach & Graeber 1985; Greenland *et al.* 1985; Greenland 1988) and typical viscosities for Hawaiian lavas in the vent range from 30 to 100 Pa s (Heslop *et al.* 1989). From Figs 3(a) and (b) it is clear that the rise speed of 0.3 $m\,s^{-1}$ for the Pu'u 'O'o eruption is consistent with lava fountaining activity (EEF = 1) and that the rise speed of 0.01 $m\,s^{-1}$ for the Kupaianaha is consistent with Strombolian activity (EEF = 0). Thus the change in eruptive character is consistent with the measured change in eruption rate, and therefore magma rise speed, and does not require any change in magma gas content.

CONCLUSIONS

Our results show that magma rise speed is the primary control on eruption style for typical basaltic gas contents and viscosities—Strombolian activity occurring at low rise speeds and Hawaiian activity at higher rise speeds. Between these two extremes the eruptive behaviour is transitional between the two styles. As in Hawaiian eruptions, the transitional behaviour involves a magma–gas mixture which has fragmented before reaching the surface and which therefore generates a lava fountain. However, because of the relatively low magma rise speed of these eruptions, significant gas bubble coalescence still occurs (as in Strombolian eruptions), and the effect of the coalescence is essentially to deplete the gas–magma mixture of gas so that fountain heights are lower than in a true Hawaiian case. In addition, the presence of large bubbles moving through the gas–magma mixture is such that it makes the fountain unstable, particularly at low rise speeds, with the fountain height fluctuating around a mean value. We have developed a model of the transition from Strombolian to Hawaiian style which allows us to express the height of the transitional fountain in terms of an energy efficiency factor, which is equivalent to the height of the fountain divided by the height the fountain would have had under true Hawaiian conditions (i.e. if the magma rise speed at depth were high enough to ensure that no gas bubble coalescence occurred).

The results show that gas bubble coalescence within the transition zone can cause fountain heights to be reduced very significantly compared with their true Hawaiian counterparts. This observation suggests that caution should be exercised if fountain height is used as an index of magma gas content in the way proposed by Head & Wilson (1987)—if the magma rise speed is not sufficiently high to prevent coalescence, the gas content found using this method may be greatly underestimated. Our results show that for normal basaltic conditions this is unlikely to be a problem provided that the magma rise speed at depth exceeds ~0.3 $m\,s^{-1}$. Significant rhythmic fluctuation in fountain height is likely to be caused by bubble coalescence effects, so the observation of such a phenomenon provides a way of assessing whether a given fountain is behaving in true Hawaiian fashion; also, as the characteristic time period of fluctuation depends on the amount of coalescence, a qualitative assessment of the effects on the estimated gas content can be made.

Our results show that the magma rise speed at which the transition occurs and the sharpness of the transition depend on both the gas content and the viscosity of the magma. The transition from Hawaiian to Strombolian activity occurs at lower rise speeds for higher viscosities and lower gas contents; this is because higher viscosity and lower gas content reduce the amount of bubble coalescence and therefore suppress the onset of Strombolian activity even at low rise speeds. Similarly, high viscosity and low gas content cause the transition from one eruptive style to the other to be relatively gradual.

A surprising result of this study is that reduction in gas content does not cause activity to change from Hawaiian to Strombolian, as has been suggested in the past (Vergniolle & Jaupart 1990). Instead, a reduction in gas content at a constant rise speed reduces the amount of bubble coalescence that will occur and so pushes the eruption style towards the Hawaiian end of the spectrum (Fig. 5). If an eruption were initially Hawaiian, a reduction in the gas content would eventually lead to passive effusion of vesicular lava rather than to Strombolian activity. Thus we emphasize that, without an accompanying decline in rise speed, a decline in gas content cannot be invoked as a cause of a change from Hawaiian to Strombolian activity.

ACKNOWLEDGMENTS

EAP was supported by NASA grants NAGW-2185 and the William F. Marlar Memorial Foundation at Brown University. LW thanks the Royal Society for support through a Leverhulme Senior Research Fellowship. We are grateful to M. T. Mangan and K. V. Cashman for discussions on bubble coalescence processes in magmas and to two anonymous reviewers for their valuable comments.

REFERENCES

Basaltic Volcanism Study Project, 1981. *Basaltic volcanism on the terrestrial planets,* Pergamon, New York.

Blackburn, E.A., Wilson, L. & Sparks, R.S.J., 1976. Mechanisms and dynamics of strombolian activity, *J. geol. Soc. Lond.*, **132,** 429–440.

Chouet, B., Hamisevicz, N. & McGetchin, T.R., 1974. Photoballistics of volcanic jet activity at Stromboli, Italy, *J. geophys. Res.*, **79,** 4961–4976.

Dobran, F., 1992a. Nonequilibrium flow in volcanic conduits and applications to the eruptions of Mt. St. Helens on May 18, 1980, and Vesuvius in AD79, *J. Volc. Geotherm Res.*, **49,** 285–311.

Dobran, F., 1992b. Modelling of structured multiphase mixtures, *Int. J. Eng. Sci.*, **30,** 1497–1505.

Gerlach, T.M. & Graeber, E.J., 1985. Volatile budget of Kilauea volcano, *Nature,* **313,** 273–277.

Giberti, G. & Wilson, L., 1990. The influence of geometry on the ascent of magma in open fissures, *Bull Volc.*, **52,** 513–521.

Greenland, L.P., 1988. Gases from the 1983–84 east-rift eruption, in *The Pu'u 'O'o eruption of Kilauea Volcano, Hawaii: Episodes 1–20, 1983–84, U.S.G.S. Prof. Pap.*, **1463.**

Greenland, L.P., Rose, W.I. & Stokes, J.B., 1985. An estimate of gas emissions and magmatic gas content from Kilauea volcano, *Geochim. cosmochim. Acta,* **49,** 125–129.

Head, J.W. & Wilson, L., 1987. Lava fountain heights at Pu'u 'O'o Kilauea, Hawai'i: indicators of amount and variations of exsolved magma volatiles, *J. geophys. Res.*, **92,** 13 715–13 719.

Heliker, C. & Wright, T.L., 1991. The Puu Oo-Kupaianaha eruption of Kilauea, *Trans. Am. geophys. Un.* (*EOS*), **72,** 521–530.

Heslop, S.E., Wilson, L., Pinkerton, H. & Head, J.W., 1989. Dynamics of a confined lava flow on Kilauea volcano, Hawai'i, *Bull. Volc.*, **51,** 415–432.

Kokelaar, P., 1986. Magma-water interactions in subaqueous and emergent basaltic volcanism. *Bull. Volc.*, **48,** 275–289.

Macdonald, G.A., 1972. *Volcanoes,* Prentice-Hall, Englewood Cliffs, New Jersey.

McBirney, A.R. & Murase, T., 1984. Rheological properties of magmas, *Ann. Rev. Earth planet. Sci.*, **12,** 337–357.

Manga, M., Stone, H.A. & O'Connell, R.J., 1993. Interactions between bubbles in magmas, *Trans. Am. geophys. Un.* (*EOS*) *abstrs,* **74,** 333.

Parfitt, E.A. & Wilson, L., 1994. The 1983–86 Pu'u 'O'o eruption of Kilauea volcano, Hawaii; a study of dike geometry and eruption mechanisms for a long-lived eruption, *J. Volc. Geotherm. Res.*, **59,** 179–205.

Sparks, R.S.J., 1978. The dynamics of bubble formation and growth in magmas: a review and analysis, *J. Volc. Geotherm. Res.*, **3,** 1–37.

Toramaru, A., 1988. Formation of propagation pattern in two-phase flow systems with application to volcanic eruptions, *Geophys. J. R. astr. Soc.*, **95,** 613–623.

Vergniolle, S. & Jaupart, C., 1990. Dynamics of degassing at Kilauea volcano, Hawaii, *J. geophys. Res.*, **95,** 2793–2809.

Walker, G.P.L., Self, S. & Wilson, L., 1984. Tarawera 1886, New Zealand – a basaltic plinian fissure eruption, *J. Volc. geotherm. Res.*, **21,** 61–78.

Wilson, L., 1980. Relationships between pressure, volatile content and ejecta velocity in three types of volcanic explosion, *J. Volc. Geotherm. Res.*, **8,** 297–313.

Wilson, L., 1984. The influences of planetary environments on the eruption styles of volcanoes, *Vistas in Astron.*, **27,** 333–360.

Wilson, L. & Head, J.W., 1981. Ascent and eruption of basaltic magma on the Earth and Moon. *J. geophys. Res.*, **86,** 2971–3001.

Wilson, L. & Head, J.W., 1983. A comparison of volcanic eruption processes on Earth, Moon, Mars, Io and Venus. *Nature,* **302,** 663–669.

Wilson, L., Sparks, R.S.J. & Walker, G.P.L., 1980. Explosive volcanic eruptions—IV. The control of magma chamber and conduit geometry on eruption column behaviour, *Geophys. J. R. astr. Soc.*, **63,** 117–148.

Wilson, L., Parfitt, E.A. & Head, J.W., 1995. Explosive volcanic eruptions—VIII. The role of magma recycling in controlling the behaviour of Hawaiian-style lava fountains, *Geophys. J. Int.*, **121,** 215–225, this issue.

Wohletz, K.H., 1986. Explosive magma-water interactions: thermodynamics, explosion mechanisms, and field studies. *Bull Volc.* **48,** 245–264.

Wolfe, E.W., Neal, C.A., Banks, N.G. & Duggan, T.J., 1988. Geological observations and chronology of eruptive events, in *U.S.G.S. Prof. Pap.*, **1463.**