5

The role of volatiles

5.1 Introduction

We saw in the previous chapter how magma moves through the mantle and crust. If this magma contains no dissolved **volatiles** then if it reaches the surface it will always be erupted effusively – simply pouring out of the vent to form lava flows or domes (depending on the chemistry and effusion rate – see Chapter 10). In practice, however, the majority of eruptions which occur subaerially involve some degree of explosiveness. As explained in section 1.2, in volcanology the term "explosive" is used to denote any eruption in which magma is fragmented and ejected from a vent within a stream of gas. In some cases volcanic explosions are transient events (these are described in Chapter 7) but often fragmentation can occur continuously during a steady eruption which might last hours or days (such eruptions are discussed in Chapter 6). So, in both a Hawaiian lava fountain (Fig. 1.1), in which clots of magma up to 1 m or more in diameter are carried up to heights of hundreds of meters above the vent, and a Plinian eruption, in which mainly tiny ash particles are carried to heights of several tens of kilometers, an explosive eruption involves fragmented magma being ejected from the vent within a stream of gas.

In some cases, eruptions are explosive because a volatile substance such as water is mixed with the magma as it approaches the surface. However, in many cases, eruptions are explosive because the rising magma has volatiles dissolved within it. As the magma rises towards the surface and the confining pressure decreases, the volatiles gradually **exsolve** from the magma forming the gas bubbles which are distributed throughout the liquid. It is the connecting together of a network of these bubbles that ultimately causes the continuous body of liquid to break apart or **fragment** into a spray of droplets or clots suspended in the gas (see Chapter 6). This process is similar to what happens when a bottle of any fizzy drink is opened. When these drinks are bottled they have carbon dioxide (CO_2) forced into them at high pressures. At high pressures the CO_2 dissolves in the liquid. When the bottle is opened the pressure is reduced to atmospheric. The solubility of CO_2 in the liquid is lower at lower pressures so when the bottle is opened not all of the CO_2 can remain dissolved and some of it exsolves and forms gas bubbles that expand and make the drink "fizz". In magmas, typically 95–99% of the "mass" of material erupted is liquid rock – at most the gas accounts for only a few percent of the weight; but that small amount of gas represents a very large "volume" as it expands to atmospheric pressure, and is fundamentally important in producing explosive eruptions.

This chapter reviews which gases are commonly dissolved in magmas, how the composition of the magmas influences the amount of dissolved gas, and how gases are released from magmas.

5.2 Volatiles in magma

When magma is in the mantle or the lower crust it contains a range of volatiles dissolved within it. The most common dissolved volatiles are H_2O (water) and CO_2 (carbon dioxide). However, anyone who has visited an active volcano is probably familiar

5.1 Scientists R. Okamura and K. Honma making
ısurements at a fumarole on Kilauea volcano, Hawai'I,
August 27, 1973. (Photograph by R.L. Christiansen,
vaiian Volcano Observatory, courtesy of the U.S.
ɔlogical Survey.)

ː h the fact that volcanoes also release a lot of sul-
ous gases – the often-noticed smell of "rotten
ɟs" is caused by the release of hydrogen sulfide
,S), and frequently deposits of sulfur can be
ınd near vents and **fumaroles** (Fig. 5.1). In fact
· most common sulfur compound released by
canoes is sulfur dioxide (SO_2). Sulfurous gases
particularly associated with basaltic magmas –
ɪasaltic eruption will typically release about 10
ıes as much sulfur as a rhyolitic eruption of the
ɲe size. This is an important factor when con-
ering the effect of volcanic eruptions on climate
e Chapter 12). A wide range of other volatiles
ı be found in varying amounts in magmas, includ-
ɨ hydrogen chloride (HCl) and hydrogen fluoride
ɟ).

The solubility of volatiles in magma

e amount of a given volatile which can be dis-
ved in a magma depends on a number of factors
ːh as the confining pressure, the composition of
ː magma, and the temperature of the magma.
ɛ pressure and the composition (see Table 2.1)
ː generally the most important. To understand
ː behavior of volatiles in rising magmas it is nec-
ɑry first to know something about the solubility
volatiles in different magmas. By carrying out
ɪny laboratory experiments scientists have found

empirical **solubility laws** for various magma-
volatile combinations. Three examples are given
below.

For H_2O dissolved in a basalt:

$$n = 0.1078 \, P^{0.7} \qquad (5.1)$$

where n is the amount of dissolved gas given as a
weight percentage (wt%) and P is the pressure in
megapascals (MPa) acting on the magma. For H_2O
dissolved in a rhyolite:

$$n = 0.4111 \, P^{0.5} \qquad (5.2)$$

For CO_2 dissolved in most magmas:

$$n = 0.0023 \, P \qquad (5.3)$$

The equations above apply only if a single volatile
species is present in a magma. In fact magmas virtu-
ally always contain several volatiles, and the solu-
bility functions are then more complex because the
volatiles interact chemically with one another as
well as with the magma. As soon as the least soluble
volatile starts to exsolve and form gas bubbles,
some small amounts of most of the other species
present will also diffuse into those bubbles.

However, for simplicity, Fig. 5.2 shows the solubil-
ity of H_2O alone in rhyolite and basalt as a function
of pressure and depth beneath the surface. This
diagram illustrates several important points about
the behavior of volatiles in magmas.

• For both basalt and rhyolite there is a general
trend in which the amount of water which can be
dissolved in the magma decreases as the pressure
on the magma decreases (i.e., as the magma rises
towards the surface).
• The solubility of water in rhyolite is considerably
greater than that in basaltic magma. For instance, at
a depth of 5 km beneath the surface the maximum
amount of water which can be dissolved in a rhy-
olitic magma is ~4.8 wt% whereas in a basalt it is
only 3.4 wt% (Fig. 5.2).
• The solubility curves show the maximum
amount of water which can be dissolved within the
magma at a given pressure. This does not mean that
the magma will actually contain that amount of
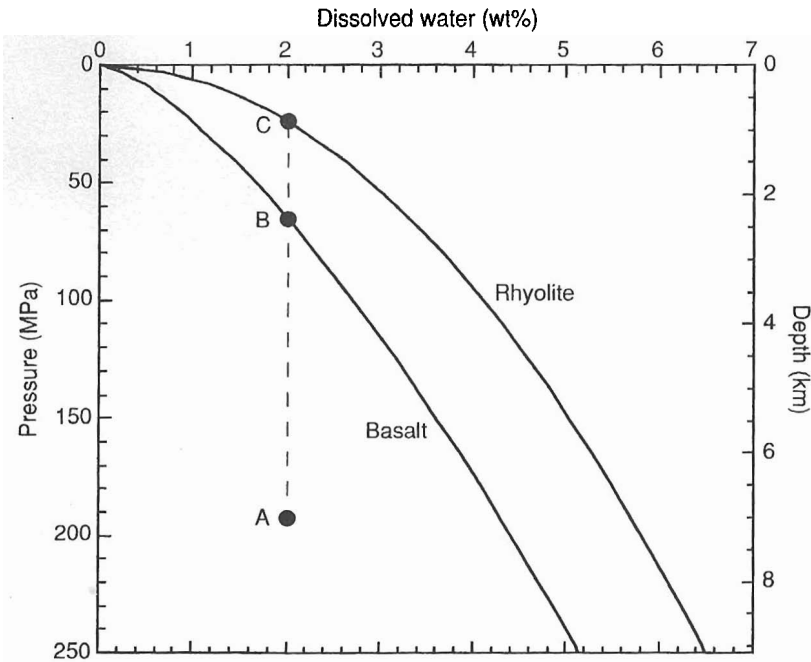water dissolved within it. For instance, at a depth of

Dissolved water (wt%)

Fig. 5.2 The solubility of H_2O in rhyolite and basalt as a function of pressure and corresponding depth beneath the surface when no other volatiles are present in the magma. A magma ascending from a depth corresponding to point A would become saturated in H_2O at point B if it were a basalt but not until it had ascended to the lower pressure marked by point C if it were a rhyolite.

~8 km beneath the surface a rhyolitic magma is capable of containing about 6 wt% of water in solution (Fig. 5.2), but the magma might actually contain significantly less than this. The magma is then said to be **undersaturated** in water, i.e., more water could be dissolved in it if more water were available. In the deep crust and mantle, magmas will generally be very undersaturated in the common volatiles such as H_2O and CO_2 because there are not enough of these available to cause saturation.

If there are two magma bodies, one rhyolitic and one basaltic, each with say 2 wt% water dissolved within it at a depth of 7 km (point A in Fig. 5.2), we can illustrate what happens to the water within the magmas as they rise towards the surface. As the magmas rise nothing happens until point B (Fig. 5.2) is reached. Here the solubility curve for basalt is intersected. At this point the amount of water dissolved in the basaltic magma is 2 wt%, which is equal to the maximum amount of water which can be dissolved in such a magma. At this point the basaltic magma is said to be **saturated** in water. So as both magmas continue to rise the basaltic magma becomes **supersaturated** in water, i.e., it has more water dissolved within it than is allowed by the solubility laws, and water starts to

exsolve from the magma forming bubbles of water vapor. From here to the surface more and more water will exsolve from the basaltic magma as it constantly tries to accommodate the decreasing solubility of the water as the pressure decreases. Exactly the same process occurs for the rhyolite except that the higher solubility of water in rhyolite means that the rhyolitic magma does not reach saturation until point C (Fig. 5.2), i.e., at a shallower depth than that at which the basalt became saturated. From point C onwards the rhyolite also exsolves water and forms bubbles of water vapor just as the basalt does.

Figure 5.3 compares the solubility of H_2O and CO_2 in basaltic and rhyolitic magma. This diagram shows some other important aspects of gas behavior.

• The solubility of CO_2 in basaltic and rhyolitic magmas is very similar (eqn 5.3) and is considerably less than the solubility of H_2O in such magmas.
• The lower solubility of CO_2 means that it tends to exsolve from magmas at much greater depths beneath the surface than H_2O. For instance, if a basaltic magma contains 0.5 wt% CO_2 and 0.5 wt% H_2O, then the CO_2 begins to exsolve at a depth of ~8 km whereas the H_2O only begins to exsolve at a
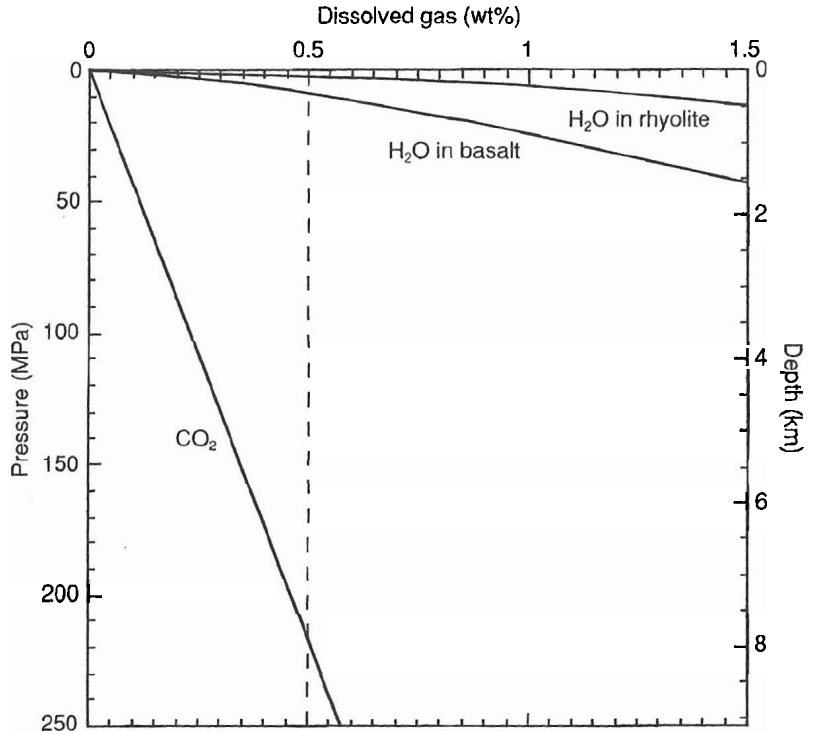
Fig. 5.3 A comparison of the solubility of H_2O in rhyolite and basalt and of CO_2 in either magma as a function of pressure and corresponding depth beneath the surface. The lower solubility of CO_2 means that it tends to exsolve from magmas at much greater depths beneath the surface than H_2O.

depth of ~0.35 km. For H_2O in rhyolite, exsolution begins at an even shallower depth (Fig. 5.3). The situation is actually more complicated when two volatiles are present because as soon as the less soluble volatile starts to exsolve, a very small amount of the more soluble volatile is also released. Fortunately, it is generally accurate enough to treat the solubilities of the two components as though they were independent.

The difference between the depths at which CO_2 and H_2O tend to exsolve can affect the behavior of the magmatic system as a whole. Say, for example, that beneath the summit caldera of a volcano there is a magma chamber with its top at a depth of ~2 km and its bottom at ~6 km, and that basaltic magma is fed to the chamber from below. By the time the magma enters the chamber, the low solubility of CO_2 (Fig. 5.3) will generally mean that the magma has already become supersaturated in this volatile, and so the magma will reach the chamber already containing some CO_2 bubbles.

Bubbles of CO_2 forming in the stored magma will rise through it to the roof of the magma chamber. Often this CO_2 will then seep out through cracks in the overlying rocks causing considerable release of CO_2 in the summit caldera. The magma in the chamber, though, is still likely to be undersaturated in H_2O at all depths within the magma chamber, and so when the magma is eventually erupted it tends to have lost most of its CO_2 but none of its H_2O. As a result, H_2O will be the most important volatile in driving the eruption. This kind of behavior has been observed in real volcanic systems such as Kilauea volcano in Hawai'I. Other effects of exsolution of magma in magma chambers have been discussed in Chapter 4.

5.4 Bubble nucleation

In principle, bubbles should nucleate in magmas as soon as the least soluble volatile species becomes saturated in the melt. However, the process of bubble nucleation is not trivial. By definition it involves the coming together of a sufficiently large number of molecules to form a stable bubble: if the bubble that tries to form is too small, the force of **surface tension** acts to try to shrink the bubble which, at

the molecular level, means pushing the volatile molecules back into the liquid. The spontaneous aggregation of molecules of a volatile species into bubbles in this way is called **homogeneous nucleation**.

The nucleation process is greatly aided if there is some, preferably irregular, surface onto which the volatile molecules can gather to minimize the effects of surface tension, in which case **heterogeneous** nucleation takes place. Thus, nucleation is helped by the presence of solid crystals, and such crystals are commonly present in many magmas, especially if the magma has been stored in a magma chamber before eruption for long enough to have cooled below its solidus temperature, so that at least one mineral has started to crystallize. This use of crystals in magmas as nucleation sites for gas bubbles has analogies with the way water vapor condenses onto dust motes in the atmosphere to form raindrops. Of course, any magma chamber or dike must have walls, and at first sight these are obvious potential sites for bubble nucleation (the next time you hold a glass of champagne look closely at the streams of bubbles nucleating at irregularities in the wall of the glass). However, the magma immediately adjacent to the wall will be relatively cool and viscous, and these factors reduce the ability of volatiles to migrate through the magma to the otherwise attractive nucleation sites.

If there are no aids to nucleation in a magma then there may be a nontrivial delay in the onset of bubble formation, and the magma may become very significantly supersaturated, by as much as ~100 MPa, before bubbles begin to form. The balance between supersaturation pressure ΔP and bubble radius r is

$$\Delta P = 2\sigma/r \qquad (5.4)$$

where σ is the surface tension, typically 0.05–0.1 N m^{-1}. With $\Delta P = 100$ MPa, the initial bubble sizes will be only a few nanometers (1 nm = 10^{-9} m). However, if heterogeneous nucleation on magma crystals is taking place at a small supersaturation, say 1 MPa, then the nucleating bubbles will be more like 1 μm in size.

If the magma has become very supersaturated before bubbles start to form, a large number of bubbles may nucleate at more or less the same time throughout the magma, and this will mean that the distances between the bubbles will be less than if nucleation has taken place in equilibrium with the decreasing pressure. Of course, there still has to be *some* movement of the volatile molecules to reach a nucleation site even if the magma is very supersaturated, and that movement, by diffusion through the liquid magma, takes a finite time. So the greater the rise speed of the magma toward the surface, the more out of equilibrium, and hence the more supersaturated, the magma is likely to become. At all stages in the rise of a magma that is exsolving volatiles there will be a competition between adding more molecules to existing bubbles and nucleating new bubbles. The distance between bubbles is a major factor in determining how efficiently volatile molecules can reach the nearest existing bubble or new bubble nucleation site. Thus the presence or absence of crystals (and also whether there are a large number of small crystals or a small number of large crystals) will have a major influence on how bubbles form and then grow.

5.5 Bubble growth

Once gas bubbles have formed within a rising magma, the bubbles grow progressively through some combination of three processes: **diffusion** of more gas into existing bubbles; **decompression** and expansion of the gas already in bubbles; and **coalescence** of bubbles.

5.5.1 Growth by diffusion

Growth by diffusion involves the migration into a bubble of molecules of the volatile compounds still dissolved in the surrounding magma. The main volatile entering the bubble will be the one that has first become supersaturated and so caused the nucleation of the bubble, but some molecules of any other volatiles present will also enter the bubble. The diffusion process of any one volatile is influenced by many factors, especially the magma composition, the magma temperature, and the mixture of other volatile species present.
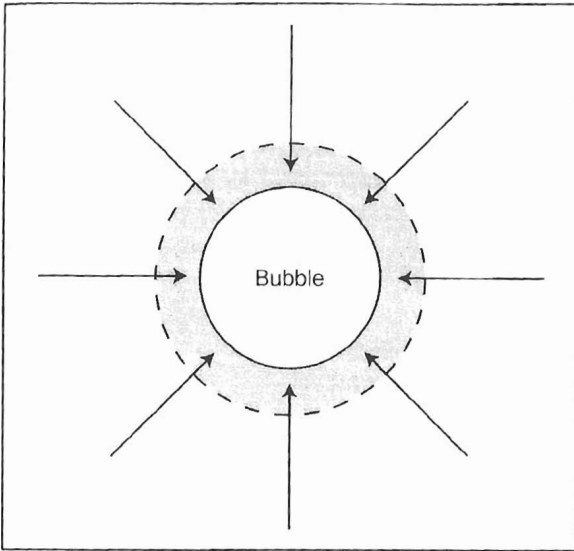
Fig. 5.4 Migration of the molecules of a volatile into a gas bubble from the surrounding liquid increases the concentration gradient in the nearby liquid (shaded area), driving more molecules towards the growing bubble.

When a bubble is very small the addition of even a relatively small amount of gas causes a relatively large increase in the size of the bubble. As the gas molecules in the magma closest to the bubble are the first to be added to the bubble, this sets up a concentration gradient around the bubble in which the number of gas molecules close to it is small compared with the number further away. This concentration gradient drives more molecules into the area of low concentration and thus drives more molecules towards the growing bubble, so further growth can occur (Fig. 5.4). This growth by diffusion is important when the bubble is small but becomes less important as the bubble grows bigger because the percentage increase in bubble volume for the addition of a given number of molecules becomes less as the bubble grows, and also because as the bubble grows there are less gas molecules left in the surrounding magma and so the concentration gradient decreases.

5.5.2 Growth by decompression

Boyle's Law (one of the **Gas Laws**) states that:

$$PV = \text{constant} \tag{5.5}$$

or

$$P_1 V_1 = P_2 V_2 \tag{5.6}$$

where P is the pressure in the gas and V is the volume occupied by the gas, and the subscripts 1 and 2 refer to conditions before and after some change takes place. This means that if the gas pressure decreases, the volume of the gas increases, i.e., the gas expands. So as magma rises and the pressure exerted on it by the surrounding rocks decreases, any gas bubbles within the magma also experience this decrease in pressure and expand in volume accordingly.

For example, if a bubble forms at a depth of ~200 m beneath the surface and grows by decompression until it reaches the surface then the initial pressure on the bubble, P_1, is

$$P_1 = \rho g h \tag{5.7}$$

where ρ is the density of the surrounding rocks, g is the acceleration due to gravity and h is the depth beneath the surface. So for $\rho = 2800$ kg m^{-3}, $g = 9.81$ m s^{-2} and $h = 200$ m, the initial pressure P_1 is 5.5 MPa. The pressure at the surface, P_2, is 1 bar, i.e., 0.1 MPa. Therefore:

$$V_2 = (5.5/0.1)\, V_1 \tag{5.8}$$

where V_2 is the final volume of the gas bubble and V_1 is the initial volume. The volume of a bubble is proportional to its radius cubed, and so the increase in the radius of the bubble in this example is found from

$$r_2 = 55^{1/3}\, r_1 = 3.8\, r_1 \tag{5.9}$$

Figure 5.5 shows the relative importance of diffusion and decompression in bubble growth. Bubbles form at a depth of ~220 m beneath the surface and grow by both diffusion and decompression until they reach the surface. The bubble radius increases by a factor of ~1000 between the nucleation depth (220 m) and the surface (Fig. 5.5). Decompression over a distance of ~200 m can increase the bubble radius by only a factor of about four; thus the growth of bubbles in this case is
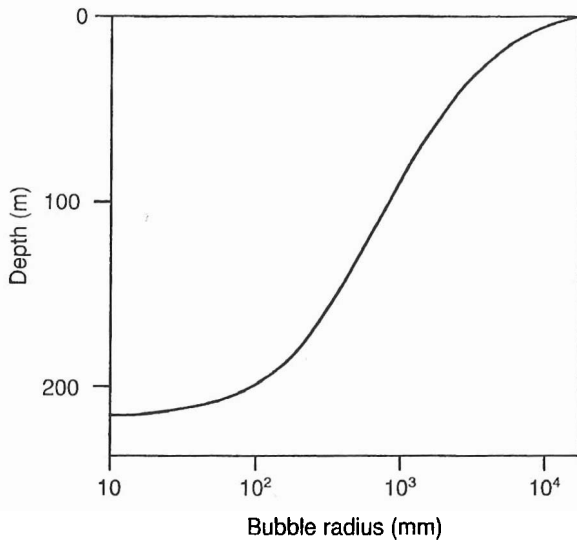
Fig. 5.5 The variation of gas bubble radius with depth beneath the surface in a rising magma. Over this range of depths the growth of the bubble is controlled mainly by the diffusion of volatile molecules from the surrounding liquid into the bubble. (Adapted from fig. 7(b) published in *Journal of Volcanology and Geothermal Research*, Vol 3, Sparks, R.S.J., The dynamics of bubble formation and growth in magmas: a review and analysis, 1–37, copyright Elsevier (1978).)

dominated by diffusion. However, if bubbles form at greater depths, the pressure change experienced by the bubbles is proportionately greater and so the importance of decompression growth is also greater. For example, for a bubble initially formed at a depth of 5 km beneath the surface, decompression alone will increase the bubble volume by a factor of more than 1000 and hence the bubble radius by a factor of 10 by the time the bubble reaches the surface.

5.5.3 Bubble coalescence

The final process by which bubbles can grow is through coalescence. This is only a significant process in certain circumstances, but it is important in transient explosive eruptions such as Strombolian and Vulcanian eruptions (see Chapter 7).

Bubbles within rising magma are always buoyant compared with the magma due to the low density of the gases they contain. This buoyancy means that bubbles are always trying to rise through the overlying magma even as the magma itself rises. The buoyancy force which causes the bubbles to rise through the magma is counteracted by the drag force of friction exerted on the bubble as it moves through the magma. As long as bubbles do not rise too quickly they maintain a very nearly spherical shape and so the buoyancy force, F_B, can be defined as:

$$F_B = (4/3)\, \pi\, r^3\, (\rho_m - \rho_g)\, g \qquad (5.10)$$

where r is the bubble radius, and ρ_m and ρ_g are the magma and gas densities, respectively. As long as the bubble radius does not become extremely big, the drag force, F_D, is controlled just by the viscosity of the magma, η, and is given by:

$$F_D = 6\, \pi\, \eta\, r\, u \qquad (5.11)$$

where u is the rise speed of the bubble. An equilibrium is reached in which the drag force equals the buoyancy force and, thus, the rise speed, u, of the bubble is:

$$u = [(2/9)\, (\rho_m - \rho_g)\, g\, r^2]/\eta \qquad (5.12)$$

From this equation it is apparent that the rise speed of a bubble through the magma is proportional to the square of its radius, i.e., larger bubbles rise disproportionately faster than smaller ones. This difference in the rise speeds of bubbles of different size is a major factor in bubble coalescence. However, the absolute rise speeds, and hence the relative speeds of bubbles, are much greater in low-viscosity magmas than in melts that are very viscous (Table 2.1). Consider a 100 μm diameter bubble rising through a melt with density $\rho_m = 2500$ kg m^{-3} (we can neglect the gas density because it is much less than this). In a basaltic magma with viscosity 30 Pa s this bubble would rise at 1.8 μm s^{-1} whereas in a rhyolitic magma with viscosity 10^5 Pa s its rise speed would be more than 3000 times smaller at 0.54 nm s^{-1}. Now consider a bubble that has managed to grow to a diameter of 10 mm. This will rise at a speed of 18 mm s^{-1} in the basaltic magma and 5.4 μm s^{-1} in the rhyolite. The ratio of the speeds in the two magmas is the same, but the 10 mm bubble
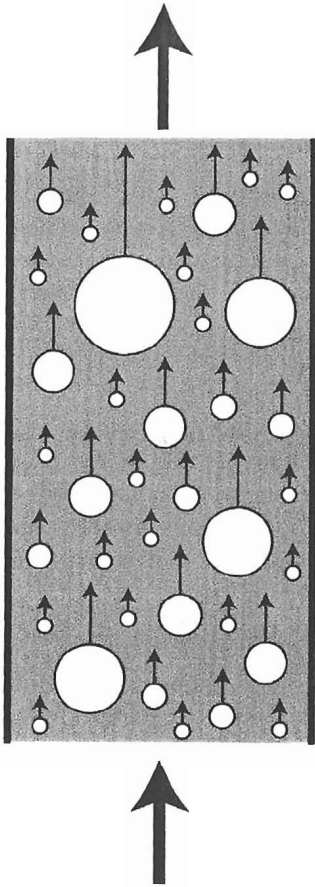
Fig. 5.6 The typical distribution of bubble sizes within magma rising toward the surface. The largest bubbles are those that formed earliest at the greatest depth beneath the surface and have grown by diffusion and decompression; the smallest bubbles are those that have most recently nucleated. The length of the arrows extending from each bubble reflects the relative rise speeds of the bubbles through the magma, largest bubbles having the greatest rise speeds.

rises by nearly twice its own diameter in one second in the basalt but by only about one-twentieth of one percent of its own diameter in the rhyolite.

Figure 5.6 shows schematically what might be seen if a "snapshot" could be taken of the bubbles within magma which is rising towards the surface. At any given time the magma will contain a population of bubbles of various sizes. Once bubbles start to form at the **exsolution level**, new bubbles continue to nucleate until the magma is finally erupted. However, as was seen earlier, the exact

way nucleation proceeds depends on the availability of sites for nucleation and the degree of supersaturation of the magma at any given time. Nevertheless, in general the magma will contain some relatively large bubbles that formed some time ago at deeper levels beneath the surface and which have been growing through diffusion and decompression, and also some very small bubbles which have just formed. A spectrum of bubble sizes will exist between these two extremes (Fig. 5.6).

All of these bubbles are rising relative to the magma because of their buoyancy but the larger bubbles are rising faster than the smaller ones (eqn 5.12). This means that the larger bubbles can overtake the slower, smaller bubbles. When this happens there are two possibilities. The small bubble may be swept around the larger one, effectively trapped in the liquid magma moving sideways and down to let the larger bubble pass, and thus may be left behind as the large bubble continues to move upward. But if the small bubble is close enough to the large one it may be swept into the wake of the large bubble. This wake consists of some liquid magma that is effectively trapped behind the large bubble and is moving up with it through the rest of the liquid. The small bubble is no longer left behind but instead rises slowly through the wake liquid and eventually collides with the bottom of the large bubble and coalesces with it, forming a single, larger bubble. This new, larger bubble then rises even faster, overtakes more bubbles and coalesces with some of them, growing larger and moving even faster and so on. This runaway process can lead to a situation where a single large bubble (called a **slug**) can fill the whole width of the dike or conduit, absorbing all of the smaller bubbles ahead of it as it continues to rise.

In basaltic magmas, a critical factor in determining whether bubble coalescence can occur is the rise speed of the magma. A simple example will illustrate why this is the case. Consider magma rising over a distance of 500 m. If the rise speed of the magma is 1 m s^{-1} then it takes 500 seconds for the magma to rise 500 m. If the magma rise speed is only 0.1 m s^{-1} then it takes 5000 seconds to move the same distance. If a bubble in the magma is rising relative to the magma at a speed of 0.01 m s^{-1} then in 500 seconds it will rise a distance of 5 m through
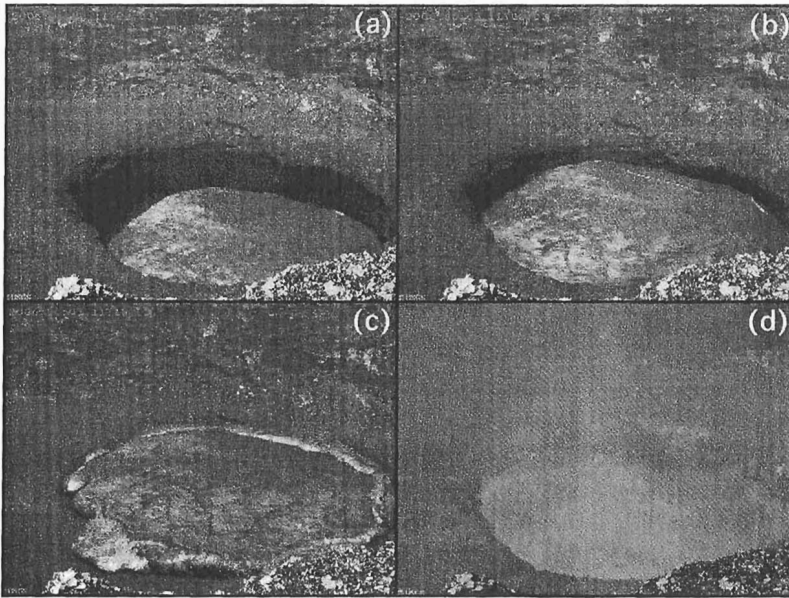
Fig. 5.7 A sequence of four frames extracted from a movie showing the rise of magma in a vent on Kīlauea volcano, Hawai'I, during an episode of gas pistoning. In frames (a) to (c) the accumulation of gas beneath the lava crust causes the lava crust to rise progressively higher in the vent. In frame (d) the gas has escaped from beneath the lava crust by tearing the crust apart in a minor explosion and the level of the lava in the vent has fallen. (Photography by Tim Orr, Hawaiian Volcano Observatory, courtesy of the U.S. Geological Survey.)

the overlying magma, whereas in 5000 seconds it will rise 50 m. Thus, during rise of magma over a given distance, the slower rising magma allows the bubbles to travel further relative to their starting position in the magma. The further bubbles are able to rise through the magma, the greater the opportunity for collision with other bubbles and therefore for coalescence. In the extreme, the magma itself may be stationary and the bubbles rise up through it to reach the surface of a lava pond in the vent. Ascent of the bubbles through such magma gives the initially largest bubbles the greatest possible opportunity for overtaking smaller bubbles and reaching the runaway stage in which single, large bubbles form, filling the entire dike or conduit. In basaltic magmas this effect can manifest itself in vigorous **Strombolian** explosions or in more gentle "gas pistoning" (Fig. 5.7). These styles of eruption are discussed in detail in Chapter 7.

Figure 5.8 shows the influence of rise speed on bubble growth for two different magma gas contents (the amount of gas initially dissolved in the magma at depth) in a basaltic magma that does not become supersaturated. The graph shows that at magma rise speeds of greater than 1 m s^{-1} the final size reached by a bubble by the time it is erupted depends linearly on the rise speed and also on the total gas content. Bubbles grow larger in magmas

with higher initial gas contents and with lower rise speeds. Higher gas contents lead to larger bubbles because the bubbles start to form at deeper levels (supersaturation occurs at deeper levels) and thus the bubbles grow more by decompression during ascent. Smaller rise speeds lead to bigger bubbles because the bubbles have more time to grow by diffusion during ascent. For bubbles in basaltic magmas rising at speeds greater than 1 m s^{-1}, the maximum size that the bubbles reach is typically between 1 and 10 mm and essentially no bubble coalescence occurs. At rise speeds less than 1 m s^{-1}, although the total gas content still influences bubble size, coalescence is the dominant factor determining the final bubble size – the slower the rise speed the greater the final size of the bubble. When bubble growth is controlled by coalescence, bubbles in basaltic magmas can grow to sizes greater than 1 m.

Figure 5.8 shows one specific example of how bubble growth is influenced by rise speed in a basaltic magma. In the case illustrated, the critical rise speed determining whether or not coalescence occurs is about 1 m s^{-1}. The value of this critical rise speed varies as a function of magma gas content and magma viscosity. At larger gas contents the critical rise speed is greater. This is because, as long as supersaturation does not become important, in

Fig. 5.8 The influence of magma rise speed on bubble growth for initial magma water contents of 1 and 2 wt% in a basaltic magma. At magma rise speeds greater than ~1 m s⁻¹ the bubbles grow by diffusion and decompression in the normal way, but at smaller rise speeds there is time for significant amounts of bubble coalescence and very much larger bubbles can be present in the magma reaching the surface. (Modified from fig. 8 in Wilson and Head (1981) Ascent and eruption of basaltic magma on the Earth and Moon. *J. Geophys. Res.* 86, 2971–3001.)
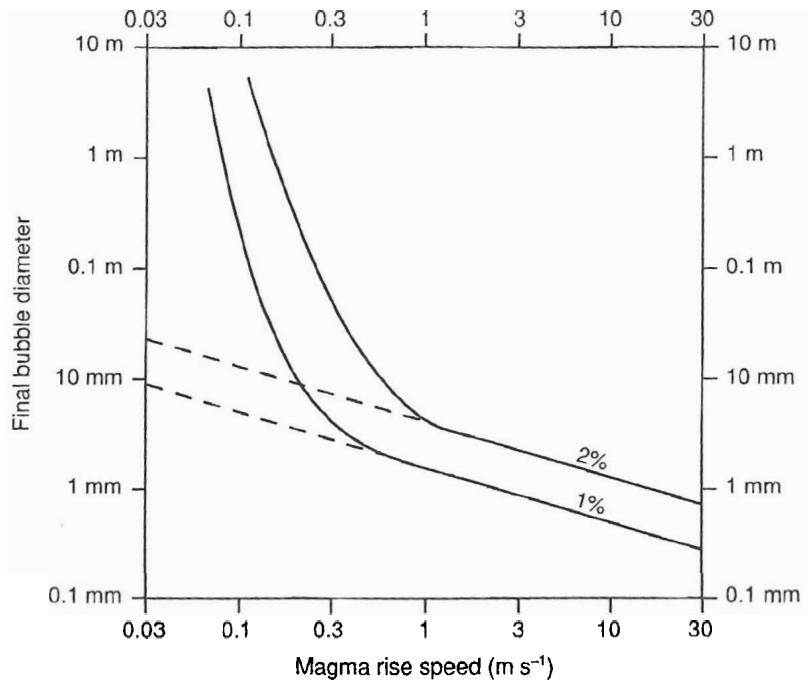
more gas-rich magmas there is a greater number density of gas bubbles in the magma at any given depth, and thus bubble coalescence becomes more likely for a given rise speed. Thus the rise speed must be proportionately larger to prevent coalescence. The critical rise speed is smaller when the magma viscosity is greater. This is because the greater magma viscosity increases the drag on the bubbles and reduces their speed, so even if the rise speed of the magma is small the bubbles move so slowly that little coalesce can occur.

5.6 Magma fragmentation and the influence of volatiles on eruption styles

A critical consequence of the growth of gas bubbles in magmas is their ability to cause the magma to **fragment**, that is, to change from being a continuous liquid containing gas bubbles of various sizes to being a continuous body of gas in which pyroclasts – droplets or clots of the liquid – are carried along by the gas. Fragmentation may occur for any one of a number of reasons, and these tend to be linked to magma composition, but not exclusively so.

An early observation about all types of pyroclasts was that almost all of them contained a network of **vesicles** (see Fig. 5.9), these being the holes left behind by bubbles of volcanic gas that were trapped in the pyroclast as it was erupted. Commonly, but not universally, the bubbles are interconnected, so that the volcanic gas has been lost and replaced by air, and equally commonly the vesicularities of pyroclasts, in other words the volume fractions of the clasts that consist of the bubble spaces, are between 70 and 80%. This led to the idea that fragmentation occurs when the gas bubbles in a magma have grown so much that they become very closely packed, so that the liquid walls between the larger bubbles collapse, allowing the large bubbles to join together. Figure 5.9 shows the kind of bubble growth history that could lead to this. Clearly not all of the bubbles will be connected together, and so the pyroclasts that are formed would be expected to contain a lot of trapped bubbles, just as is observed.

However, there tends to be a difference between the size distributions of trapped bubbles in pyroclasts derived from magmas of different compositions. Pyroclasts of more evolved, highly viscous
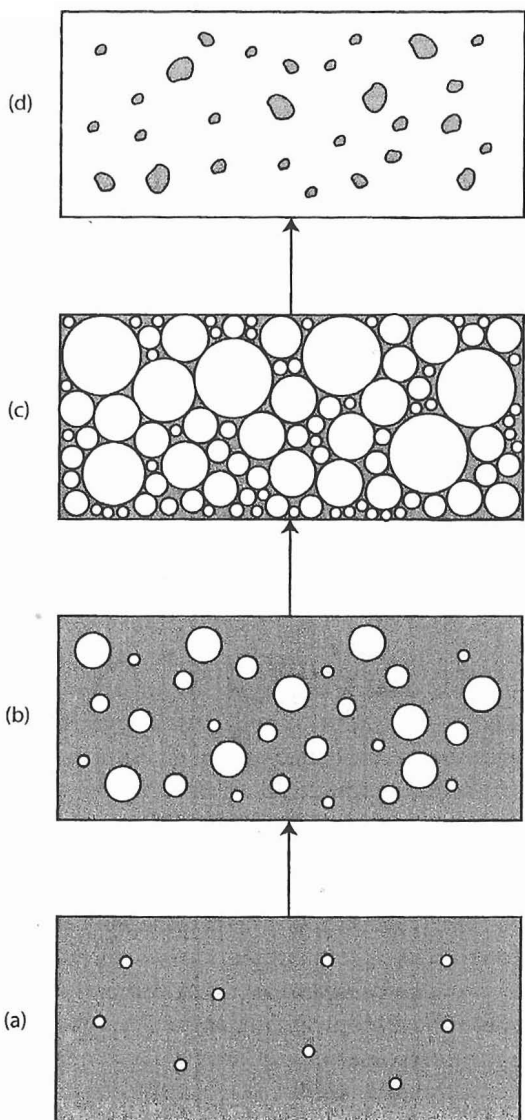
Fig. 5.9 Four stages in the growth of gas bubbles in a magma leading to magma fragmentation and an explosive eruption. Bubbles start to nucleate in (a); older bubbles have grown by diffusion and decompression and new bubbles have nucleated in (b); the bubble number density is so large in (c) that bubbles are extremely crowded and the walls between adjacent bubbles can start to collapse; in (d) so many bubble walls have collapsed that the magma has changed from a liquid containing bubbles to a gas containing liquid droplets.

magmas, called **pumice** clasts (see Fig. 5.10), generally contain a smaller range of vesicle sizes than more basaltic pyroclasts, usually called **scoria** (Fig. 5.11). This is due to the reduction in coalescence,
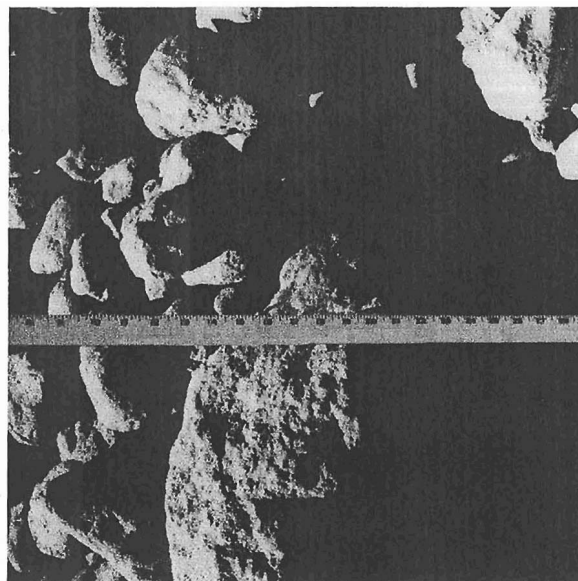

Fig. 5.10 Pumice clasts in the deposit from a pyroclastic density current erupted at Mount St Helens volcano, Washington State, USA, in August, 1980. Scale is in inches. (Photograph by James W. Head III.)

the greater likelihood of supersaturation and hence the nucleation of large numbers of small bubbles, and the reduced ability of volatile molecules to diffuse through viscous magmas as compared with basaltic ones. This suggests that in the more evolved magmas the cause of fragmentation is not just the close packing of large bubbles. Instead, it may often be the fact that it is difficult to force viscous liquid to flow through the narrow films of liquid separating the bubbles. On the time scale of the changing stresses to which the magma is subjected as it accelerates upward through a dike toward the surface (the subject of the next chapter) its rheology is no longer Newtonian, and it develops an effective strength. When the stresses exceed that strength, the magma fractures as though it were a brittle solid; a similar situation was described in Chapter 3, where rheology was a function of environmental stresses in connection with the behavior of rocks in the upper mantle.

There is a final very important factor connected with gas bubbles and magma fragmentation. When the fragmentation process occurs throughout the magma at some specific depth in a dike, every batch of magma rising through the system undergoes the

Fig. 5.11 Scoria clasts in the Montana Colorada cone, Lanzarote. The cone formed during an eruption in April 1736. (Photograph by Elisabeth Parfitt.)

same process, and a relatively uniform and steady stream of gas and pyroclasts emerges through the vent. This is the commonest way in which viscous magmas are erupted. However, if a great deal of coalescence of bubbles occurs, especially to the point where giant bubbles fill nearly the whole width of the dike, then the magma rising though the dike is far from uniform. As the magma between the giant bubbles reaches the surface, it flows out of the vent as lava, containing whatever gas bubbles have avoided being swept up by the giant bubbles. As the giant bubbles reach the surface they up-dome the surface of the lava above them forming a skin. This skin may be plastic if lava is flowing out of the vent fairly quickly and will stretch and tear easily; alternately, if the lava is moving away only very slowly, the skin may have cooled significantly since the arrival of the previous giant bubble and will fracture in a brittle manner as it bursts. The gas

trapped inside the giant bubble is generally at a higher pressure than the atmosphere and so it expands, throwing clots of lava from the torn skin upward and outward in a Strombolian explosion. It was shown earlier that bubble coalescence occurs more easily in basaltic than more evolved magmas, and Fig. 5.8 demonstrates that, even for basaltic magmas, there is quite a sharp divide, in terms of magma rise speed and volatile content, between eruptions in which bubble coalescence is or is not important.

As a result of the above issues, explosive eruptions are split into two main classes. Where magma rise speed is high enough to prevent significant coalescence, gas bubbles stay coupled to the magma within which they form and this leads to "steady" eruptions – eruptions in which a continuous stream of gas and magma clasts is erupted (see Chapter 6). In basaltic eruptions with low enough rise speeds, bubble coalescence dominates and the bubbles decouple from the magma in which they form. Bubbles rising and coalescing lead to a situation in which the distribution of gas through the magma column is uneven. This can lead to a range of be-haviors from relatively minor fluctuations in the intensity of the eruption (like a pulsing in the height of lava fountains in Hawaiian eruptions) through to intermittent, discrete explosions occurring as successive pockets of gas rise through the magma to be erupted. These eruptions are referred to as "transient" and they are discussed in Chapter 7.

5.7 Summary

- Most volcanic eruptions are to some extent **explosive**. In the context of volcanology this means that the magma is torn apart and ejected from the vent as clots or blobs within a stream of gas. Explosive eruptions can be transient events, occurring every few seconds to minutes or hours, or can be continuous, steady eruptions which last hours or days.
- Most explosive eruptions owe their explosive character to the presence of gases dissolved within the rising magma. The solubility of vol-canic gases in magma decreases with decreasing pressure, so as magma rises towards the surface a

point will be reached where the magma becomes **supersaturated** and gas starts to **exsolve** forming bubbles within the magma. The most common gases within magmas are H_2O, CO_2, SO_2 and H_2S.

- The solubility of gases in magmas is dependent on both the type of gas and the composition of the magma. The solubility of H_2O in rhyolite is, for instance, considerably greater than that in basalt, whereas the solubility of CO_2 is considerably less than the solubility of H_2O in either a basalt or a rhyolite. The solubility of volatiles is more complex when more than one species is present because they interact with one another.

- When bubbles first form in a magma they are typically extremely small. Bubbles can grow from their initial size as the result of three processes: diffusion, decompression and coalescence. The relative importance of each process depends on the amount of gas present in the magma, the magma composition and the magma rise speed.

- Bubble coalescence due to large bubbles overtaking smaller ones is important in low-viscosity magmas at low magma rise speeds. The rise speed, and its control on whether significant bubble coalescence does or does not occur, determines the fundamental character of the resulting eruption. If the magma rise speed is low and thus bubble coalescence is important, the gas segregates from the magma and rises through it, causing the kinds of discrete transient explosions described in Chapter 7. If the rise speed is higher then significant coalescence will not occur. In this case the magma and gas stay locked together and this leads to continuous, steady explosive eruptions in a wide range of magma compositions as discussed in Chapter 6.

5.8 Further reading

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5.9 Questions to think about

1 Why are magmas stored in magma reservoirs at depths of a few kilometers below the surface more likely to contain bubbles of carbon dioxide than bubbles of water vapor?

2 What is the main factor controlling whether volatiles exsolve in a way close to being in equilibrium with the decreasing pressure in a rising magma or out of equilibrium with the pressure?

3 What are the three processes by which the average size of the gas bubbles in a rising magma increases?

4 What property of the magma allows bubbles to coalesce more easily in basaltic magmas than more evolved magmas?

5 What is the main control on whether an explosive eruption involves the discharge of a steady or unsteady stream of gas and pyroclasts through the vent?