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Notes

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Abstract: Gas exsolution and segregation are fundamental controls on eruption dynamics and magma genesis. Basaltic magma loses gas relatively easily because of its low viscosity. However, bubbles grown by decompression and diffusion during magma ascent are too small to segregate. Coalescence, however, can create bubbles big enough for gas to escape from the rising basalt magma. In evolved magmas, such as andesite and rhyolite, high viscosity prevents bubbles rising independently through the magma. The original gas content of magma erupted as lava is commonly the same as that erupted explosively, so that a gas separation mechanism is required. A permeable magma foam can form to allow gas escape once bubbles become interconnected. Magma permeabilities can be much higher than wall-rock permeabilities, and so vertical gas loss can be an important escape path, in addition to gas loss through the conduit walls. This inference is consistent with observations from the Soufrière Hills Volcano, Montserrat, where gas escapes directly from the dome, and particularly along shear zones (faults) related to the conduit wall. Dynamical models of magma ascent have been developed which incorporate gas escape. The magma ascent rate is sensitive to gas escape, as the volume proportion of gas affects density, magma compressibility and rheology, resulting in both horizontal and vertical pressure gradients in the magma column to allow gas escape. Slight changes in gas loss can make the difference between explosive and effusive eruption, and multiple steady-state flow states can exist. In certain circumstances, there can be abrupt jumps between effusive and explosive activity. Overpressures develop in the ascending magma, caused primarily by the rheological stiffening of magma as gas exsolves and crystals grow. A maximum overpressure develops in the upper parts of volcanic conduits. The overpressure is typically several MPa and increases as permeability decreases. Thus, the possibility of reaching conditions for explosions increases as permeability decreases, both due to overpressure increase and the retention of more gas. Models of magma ascent from an elastic magma chamber, combined with concepts of permeability and overpressure linked to degassing, provide an explanation for the periodic patterns of dome growth with short-lived explosive activity, as in the 1980–1986 activity of Mount St Helens. Degassing of magma in conduits can also cause strong convective circulation between deep magma reservoirs and the Earth's surface. Such circulation not only allows degassing to occur from deep reservoirs, but may also be a significant driving force for crystal differentiation.

Degassing of magma is one of the most fundamental processes of volcanism, affecting the dynamics of their eruptions and the evolution of magmas in the crust. Volcanic eruptions are sometimes explosive and at other times more passive – with separation of gas from magma and discharge of lava. Gas exsolution and gas separation can have profound effects on the physical properties of magmas. In particular, loss of water (usually the main volcanic gas) results in large increases in melt viscosity and causes crystallization due to undercooling of the melt. Magma density is also greatly affected by degassing, decreasing substantially if exsolving gas is retained and increasing if the gas escapes. These physical property changes have profound effects on eruptions, as they are the first-order controls on the flow along volcanic conduits, and therefore have a large influence on the geophysical manifestations of volcanism. Thus, understanding of degassing is central to monitoring and forecasting of eruptions.

Degassing has profound effects on the phase equilibria of magmas, because rather small amounts of water dissolved in magmas can reduce their liquidus temperature by hundreds of degrees. Conversely, degassing of magma can cause spontaneous crystallization. While these effects have been known about for many decades (e.g. Tuttle & Bowen 1958; Cann 1970), it is only relatively recently that their significance for understanding volcanic processes has been widely appreciated. Crystallization due to degassing during magma ascent may prove to be as important as cooling in igneous petrogenesis. Additionally, volcanic conduits provide pathways for exchanges between deep magma reservoirs and the Earth's surface, so that gases

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and gas-rich magmas can rise and, in principle, degassed and oxidized magmas can return into deep chambers. The role of degassing and resulting convection exchanges along conduits has not been widely considered as a major mechanism for differentiation of magmas.

This contribution focuses on the role of volatiles in the dynamics of volcanic flows through conduits. Other aspects of gas bubble nucleation and growth, the effects of water on melt viscosity and the more geochemical aspects of volcanic degassing have been well covered in other recent publications (e.g. Dingwell 1998; Navon & Lyakhovsky 1998; Blower *et al.* 2001*a*, *b*; Wallace 2001;). In the last few years there have been significant advances in understanding the role of degassing in conduit flows. I draw attention in particular to modelling research, which shows that conduit flows are highly unstable due to strong non-linearities and feedback loops caused by degassing. Much of the complexity and unpredictability of volcanic eruptions is related to these effects. On the other hand, these same non-linearities related to degassing can result in periodic behaviours, so that there is some prospect of being able to forecast volcanic activity. I will also emphasize that the main geophysical signals monitored at volcanic eruptions are usually controlled by degassing processes and related side effects, such as rheological stiffening of ascending magma and development of high overpressures in volcanic conduits. I also discuss an idea that does not seem to have been widely considered in the evolution of magmas in chambers, namely that in a long-lived conduit system the degassed magma can drain back into the chamber. The liquidus of degassed magma increases markedly with rising pressure, so that this descending magma will either crystallize extensively during descent or will be supersaturated and therefore cause crystallization in the chamber. Finally, I raise the issue of gas contents in magmas, and suggest that in some volcanic systems – notably in arcs – gas contents may be higher than commonly thought. Very wet magmas may degas significantly when emplaced in shallow upper crustal magma chambers, and large amounts of exsolved gas can be present in the chamber prior to eruption.

Gas segregation dynamics

The basaltic case

The separation of exsolving gas from magma is strongly controlled by viscosity. The usual mantra is that basaltic magma has sufficiently low viscosity that gas bubbles can rise at speeds that are comparable with or much larger than the speed of rising magma. This assumption is, however, worth somewhat closer scrutiny. In basaltic eruptions the typical speeds of magma ascent along dykes are of order 1 m/s (Wilson & Head 1981; Sparks *et al.* 1997). The rise speed, u, of a spherical bubble of diameter d can be estimated from Stokes' law as follows:

$$u = d^2 \rho g / 18\mu \qquad (1)$$

where ρ is the magma density, g is gravity and μ is the viscosity. For values of $\rho = 2700$ kg/m^3 and $g = 9.81$ m/s^2, a bubble has to exceed about 14 cm diameter for $\mu = 30$ Pa s and 4.5 cm for $\mu = 3$ Pa s to move significantly faster than the magma. Sparks (1978) modelled the diffusive growth of water bubbles in ascending basaltic magma, and found that bubbles would be in the range of 0.1 to 1 cm diameter for typical rates of magma ascent. This inference is consistent with the observation that most of the bubbles in basaltic scoria are in this size range. Bubbles of this size have speeds in the range 0.5 to 5 mm/s. Thus, the segregation process even in basalt magma is not simply a matter of growing gas bubbles rising faster than magma, even though this will be a viable mechanism in a static magma column or very slowly rising magma. Other mechanisms need to be invoked, such as bubble coalescence, recirculation of degassed magma within the conduit due to fire-fountaining, and convection processes within magma conduits and magma reservoirs.

What is commonly observed in many basaltic eruptions is that magma can erupt both explosively, typically in Strombolian style or Hawaiian fire-fountains, and as degassed lava. In many basaltic eruptions, the two kinds of activity can be observed simultaneously. Thus the fact that efficient gas segregation occurs is certain. Fire-fountaining is one well-established way of segregating gas. Here bubbly gas-rich magma expands into the atmosphere (Parfitt & Wilson 1995), the magma is torn apart, and degassed lumps either fall back to amalgamate into clastogenic lava flows (Swanson & Fabbi 1973) or are mixed back into the rising magma within the conduit. These mechanisms may at least partly explain how degassed lava can emerge slowly at the base of cinder cones, simultaneously with Strombolian or fountaining activity.

It is harder to explain the extrusion of large amounts of degassed lava in cases where the explosive activity is weak or even absent. Thus, ideas have emerged which invoke deeper level segregation processes. Pioneering research on degassing of ascending basaltic magma was carried out by Y. Slezin of the Institute of Volcanic Geology, Petropavlovsk, Kamchatka in the 1980s (summarized in Slezin 1995, 2003). Unfortunately, much of this work is in Russian or in rather poor translations, and is not widely known outside Russia. However, this research pioneered advances in understanding of basaltic eruptions, and showed for the first time that volcanic flows could be intrinsically unstable and could fluctuate between stable steady flow states. The key ingredients of Slezin's model are shown schematically in Figure 1, where two conceptual states of flow are envisaged. In the first state (Fig. 1a) the magma rises slowly and the bubbles have time to coalesce and to grow to a large enough diameter that they can rise at speeds much faster than the surrounding magma. Large bubbles sweep up and assimilate small bubbles, and so this is a potential runaway process. The gas pockets may become big enough to fill the vent (slugs), and so the gas escapes in bursts while the degassed magma quietly effuses. In the second state (Fig. 1b), the flow is so fast that the bubbles do not separate and have no time to coalesce during ascent. An intense explosive flow develops at the surface, with the bubbly magma fragmenting as it nears the surface.

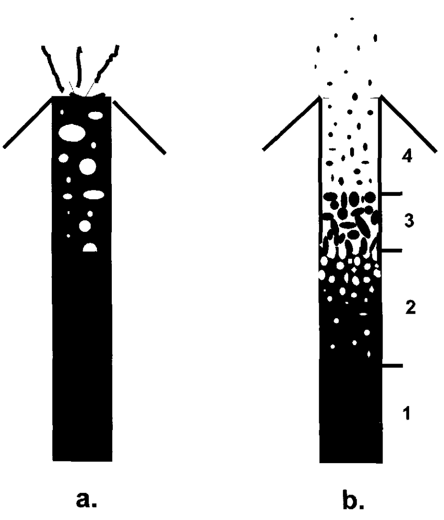

Fig. 1. Schematic of two regimes of basaltic eruptions. In (**a**) the magma rises so slowly that large bubbles of gas can form by coalescence. These bubbles are large enough to rise at speeds much greater than the magma rise speed, and thus gas escapes from the top of the magma column. In (**b**) the magma rises so fast that growing bubbles do not have time to coalesce and are too small to have speeds which are significantly different from the magma. The numbers refer to: 1, the region of magma without bubbles; 2, the region of bubbly magma; 3, the region of concentrated particle–gas dispersion; and 4, the region of an expanded dilute concentration gas and particle dispersion.

Slezin developed a quantitative model of bubbly magmatic flow, which incorporated coalescence. His model involved larger bubbles rising and absorbing smaller bubbles. Conceptually, it is easy to imagine that bubble coalescence and efficient gas segregation will occur when the time-scale of bubble rise and interaction is fast compared with the ascent time of the magma. A typical result is shown schematically in Figure 2, where flow rate along a conduit is shown as a function of chamber pressure. These results are fundamental to understanding volcanic eruptions, and are not immediately intuitive. At first, it might be expected that there would be a smooth monotonic transition between the two end member regimes. However, the steady solutions to the flow equations delineate a sigmoidal shaped curve. The lower branch (A–B) in Figure 2 represents the slow ascent and well-degassed end member, and the upper branch (C–D) represents the explosive flow end member. The intermediate branch (between B and D), however, may not be stable. Further, for a wide range of conduit widths and magma viscosities it is possible to have two steady flows corresponding with the upper and lower branches. Close to conditions of points B and D, the system is

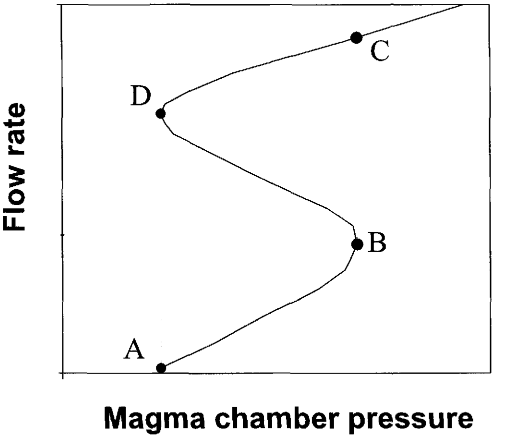

Fig. 2. A general schematic diagram of steady-state flow rate up a conduit against magma chamber pressure, to illustrate the abrupt changes in flow regime that can occur. The two stable branches (A–B and C–D) may correspond with different non-linear physical processes such as bubble coalescence and crystal growth.

intrinsically unstable and there can be jump-like changes of flow behaviour from predominantly effusive to explosive and vice versa. Points like B and D are known as cusps in catastrophe theory. Figure 3 shows some example calculations by Slezin (2003) for a magma with 4.6% water, which show conditions where jump-like behaviours can be anticipated. The parameter conduit resistance is defined as b^2/μ, with b being the conduit width or diameter and μ the magma viscosity. The steady-state curves are characteristically sigmoidal.

The shape of the curves from Slezin's models reflect the strong non-linearities in bubbly flows with strong feedback loops, and can be explained intuitively by a thought experiment. If the chamber pressure is varied, but conduit resistance is kept constant, then similar sigmoidal curves are calculated (Slezin 1995, 2003). So here we consider the horizontal axis to be magma chamber pressure – imagine that the system is on the branch A to B, with magma source pressure increasing. We shall not enquire into the cause of this increase, although there are a number of mechanisms that can lead to such an increase (e.g. chamber replenishment). As pressure increases, the flow rate increases, but the ascent speed is sufficiently slow that growth and coalescence result in large bubbles. For a given parcel of magma, the gas escapes well before the parcel nears the surface. The rising column is thus divided into gas-rich slugs and parcels of partially degassed magma, with the slugs travelling much faster than the magma. Additionally, the overall density of the magma column is high, so that the pressure difference driving flow (the difference between the weight of the magma column and the chamber pressure) is small. At Point B there is no stable steady flow for a very tiny increase in driving pressure. The flow accelerates, there is less time for bubble segregation and coalescence, and the system inexorably moves to the high-flow condition where the gas does not segregate. Here the magma and gas expand together, and very high volumetric fractions of gas develop in the upper parts of the conduit. If the magma is broken apart at some threshold condition as it vesiculates, then the column is divided into a lower region of bubbly magma and an upper region of particles mixed with gas. The conduit system contains much more gas, and so the weight of the magma columns is much less. Thus, the pressure difference driving flow at C is much higher than at B, even though the chamber pressure has not changed. Differences in flow rates between B and C are typically factors of tens to thousands; in other words, the flow jumps from quiet effusion with occasional intermittent bursts of gas slugs, to sustained explosive discharge at very high rates. Similar logic holds at D, where a slight decrease of chamber pressure results in the system dropping to the lower branch with a feedback loop such that, as the flow slows, coalescence and segregation once again become dominant.

Cyclic behaviour is easily envisaged. During the high-discharge explosive phase, volume is

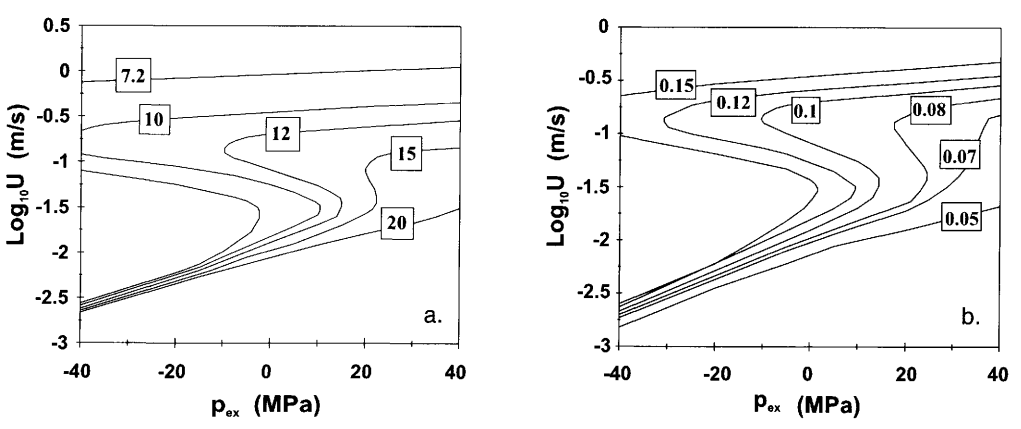

Fig. 3. Calculations after Slezin (1995, 2003) of the steady-state magma speed along a vertical conduit against the magma chamber excess pressure. The calculations are for a magma containing 4.6% water. On graph (**a**) the conduit resistance (see text) is fixed at 10^{-4} m^2/Pa, and curves are shown for different conduit lengths in km. On graph (**b**) the conduit length is fixed at 12 km, and curves are shown for different values of the conduit resistance (multiplied by 10^3).

removed rapidly from a magma chamber, reducing pressure faster than can be restored by recharge. The system then drops to the lower branch, and recharge is then sufficient to build the pressure up once again to the point that sustained explosive discharge can begin again. Periodic alternation of intense explosive discharges and quieter effusive activity have been observed, for example, in the recent eruptions at Etna (Burton *et al.* 2003).

Another mechanism of gas segregation was proposed by Vergniolle & Jaupart (1986, 1990), who showed that the architecture of the magma plumbing system could have an important effect. In experiments, they demonstrated that, in a chamber containing bubbly magma with a flat roof and narrow conduit outlet, gas bubbles would accumulate at the roof to form a foam. They further showed that this foam would periodically become unstable. Thus, the output of fluid from the conduit would be punctuated with pulses of gas-rich magma. They proposed that this was a plausible explanation of the remarkably regular 21-day fire-fountaining in episodes at Pu'u 'O'o on Kilauea, Hawaii. The concepts developed by Vergniolle and Jaupart can be generalized to a common observation on basaltic shield volcanoes. Eruptions from the summit or high-altitude vents commonly are explosive, with large amounts of gas separation, whereas flank and lower altitude eruptions are weakly explosive or just quiet lava effusions. Examples include: the activity of Etna in the last decades, with explosive activity and degassing largely at the summit crater and quiet effusion of degassed lava in flank vents (Burton *et al.*, 2003); and the 1999 and 2000 eruptions of Mount Cameroon (Suh *et al.* 2003), where almost all the explosive degassing took place at near summit high-altitude vents and almost all the lava effused quietly from lower altitude flank vents. This behaviour might be explained by lateral flow of magma through dykes from a central conduit to flank vents, with progressive gas segregation to the roof of the dyke and down flow of degassed denser magma to erupt at lower altitude vents.

Another idea for gas segregation is convective circulation in the conduit (Kazahaya *et al.* 1994; Allard 1997; Stevenson & Blake 1998) with drain-back of degassed magma into the chamber (Dixon *et al.* 1991; Wallace & Anderson 1998). This idea was developed to explain volcanoes like Oshima (Japan), Stromboli (Italy), Etna (Italy), Villarricca (Chile) and Popocatépetl (Mexico), where substantial gas emissions are observed without any eruption of degassed lava. Degassing can continue for many months or years, or even indefinitely as at Stromboli. The quantities of gas released can be orders of magnitude greater than the gas expected to be available in the narrow conduit (Wallace 2001). To explain these observations the idea is that gas-rich magma rises in the conduit and convectively exchanges with degassed magma. Thus, a large volume of magma from the chamber becomes available. The fluid dynamical analysis of Stevenson & Blake (1998) shows that the large gas fluxes are easily attained in narrow conduits.

In view of the development of these ideas of gas segregation processes, it is worthwhile to re-examine another commonly held axiom; namely that a common pattern of basaltic eruptions involves magma with progressively lower gas content with time. The pattern is well established in monogenetic volcanoes, where the early phase of an eruption typically is quite explosive, with sustained discharges and violent Strombolian activity constructing cinder cones. The eruptions then progressively calm down, with lava effusion becoming increasingly important with time and a declining magma effusion rate. Examples include the 1973 eruption of Heimaey (Self *et al.* 1974), the 1949–1956 eruption of Paricutín (Luhr & Simkin 1993) and the 1989–1990 eruption of Lonquimay, Chile (Moreno & Gardeweg 1989). One explanation of such sequences is that the source magma body is stratified with respect to dissolved gas content (Kennedy 1955) or (exsolved) bubble content. Alternatively, the sequence can be interpreted as a consequence of increasingly efficient shallow gas segregation processes with time as the magma ascent rate declines. Gas bubble coalescence, slugging and convective recycling of degassed magma back down the conduit are plausible contributions to these changes, which do not necessarily imply decreasing gas content with time in the deeply sourced magmas.

The silicic case

The problem with silicic magmas is that speeds of gas bubbles are negligible compared with the ascent speeds of magmas, so that segregation of gas from magma due to a significant speed contrast between bubble and melt is not plausible even if bubbles coalesce significantly. Thus, a different mechanism of gas escape is required. It has also been long recognized that silicic magma that erupts explosively typically has similar volatile contents to magma that erupts as degassed lava. There are now many documented examples of degassed silicic and intermediate lavas that must have started ascent with water contents of several per cent as constrained by melt inclusion studies and phenocryst assemblages (e.g. Hervig *et al.* 1989; Barclay *et al.* 1996; Martel *et al.* 1998; Devine *et al.* 1998; Blundy & Cashman 2001). This problem was resolved by Taylor *et al.* (1983) and Eichelberger *et al.* (1986), who postulated that silicic magmas must develop permeability as they vesiculate during ascent. The gas can then escape through the permeable magma foam. Whether the magma erupts explosively or as lava then depends on the competition between magma ascent and gas escape. Slow rates of ascent allow the gas to escape and lava extrusion, whereas gas is retained for fast rates of extrusion and explosive activity results.

Models of magma ascent with permeable foam degassing generate results very like those of Slezin (2003). Jaupart and Allegre (1991) considered models with a horizontal gas mass flux from the permeable magma with constant permeability and pressure difference dependent on the depth in the magma column. They found that for different magma flow rates multiple steady-state flows were possible, and concluded that in some circumstances the transitions from one eruptive state to another (i.e. between explosive and effusive eruption) were extremely sensitive to flow conditions. Woods and Koyaguchi (1994) presented somewhat similar calculations, assuming a hydrostatic pressure in the conduit wall rather than a lithostatic pressure, as in the calculations of Jaupart and Allegre (1991). They reached the same conclusions. Massol & Jaupart (1999) developed models of conduit flow that take account of variations of magma viscosity due to gas exsolution, compressibility effects of gas bubbles and development of gas overpressure due to bubble expansion. They demonstrated that large horizontal pressure gradients can develop across volcanic conduits as a consequence of larger bubble overpressures developing in faster rising magma at the centre of the conduit. If bubbles are interconnected, then gas can escape to the conduit walls.

These models assumed, following Eichelberger *et al.* (1986), that the gas is lost horizontally through the wall-rocks. Jaupart (1998) also suggested that gas was also lost along fracture networks developed in the magma and in the conduit walls, based on field observations at the Mule Creek vent complex (Stasiuk *et al.* 1996). Observations at the Soufrière Hills volcano, Montserrat, suggest that vertical permeable flow can be dominant. In this case, the gas plume emerges from the dome interior (Edmonds *et al.*, 2001), and in particular is focused on shear zones associated with lava lobe extrusion (Fig. 4). At Soufrière Hills, the lava dome is extruded in pulses along shear faults, in which a viscous plug is commonly first extruded followed by a lava dome lobe (Watts *et al.* 2002). The shear faults are rooted along the conduit wall, and much of the gas is observed to escape either along such a boundary or pervasively across the upper surface of the dome. Láscar, Chile, is another case where a high-permeability boundary between magma and a conduit has guided the gas escape vertically. Here, the vigorous gas fumaroles that feed the persistent gas plume are concentrically arranged at the margins of the vent and lava dome (Matthews *et al.* 1997). Such observations support the concept of lateral gas escape from permeable magma to the conduit walls (Massol & Jaupart 1999). If the wall-rocks have significantly lower permeability, then vertical gas escape through the magma or along the conduit walls may be much easier. Wall-rocks can also be self-sealed by precipitation from escaping gases, as seems to have been the case at Montserrat (Hammouya *et al.* 1998). Additional permeability can develop in lava domes and upper parts of the conduit as fracture networks develop (Stasiuk *et al.* 1996; Jaupart 1998; Sparks *et al.* 2000).

An important but incompletely understood issue is the development of bubble connectivity in magmas to allow gas to escape by permeable flow. Theoretical and experimental work suggests that magmas first become permeable at about 30% porosity, when bubbles start to interact and coalesce (Klug & Cashman 1996; Blower 2001). Permeability increases rapidly as bubble concentration rises, and permeabilities as high as 10^{-12} m^2 can develop in magma foams with high vesicularity (60–70%). However, the development of magma permeability is complex. If bubbles grow in a static melt then very high porosities can develop without the thin melt films between bubbles being disrupted (e.g. reticulites). On the other hand, shearing can bring bubbles into contact and promote connectivity (Whalley 1987). Thus, magma permeability may develop preferentially near to the conduit margins (Stasiuk *et al.* 1996). Microlite crystallization can also enhance permeability. Gas bubbles are confined to areas of residual melt that form a distributed phase between the crystals. In a highly crystalline and relatively low-porosity magma, the permeability can thus be high because the distributed residual melt has very high porosity and high connectivity. For example, measurements in Melnik & Sparks (2002) show that, despite low porosity (10–15%), samples of the Soufrière Hills dome have similar permeability to high-porosity pumice (60–70%). In this case, the residual melt component had very high porosity (70–90%).

Fig. 4. View of the Soufrière Hills andesite lava dome, Montserrat, on 21 December 1997, showing a plume of volcanic gas being transported by the wind to the west (right to left). The photograph was taken from the south.

Models incorporating vertical gas flow were developed by Melnik and Sparks (1999) in application to Montserrat. Their treatment developed the earlier models of Jaupart and Allegre (1991) and Woods and Koyaguchi (1994) by including the coupling between the magma permeability and porosity to calculate gas escape rates. The model of Melnik and Sparks focused on slow ascent rates to model dome extrusion, and did not consider flow rates high enough to result in explosive eruptions. The model also included the kinetics of crystallization due to gas exsolution. Thus, calculations for vertical gas escape in a situation where explosive magma fragmentation is permissible have yet to be done. It is, however, certain that similar dynamical regimes will be found. The alternation of Vulcanian explosions and lava dome extrusion in 1997 at the Soufrière Hills Volcano (Druitt *et al.* 2002) is what would be expected in an unstable system where the flow rates are high enough for transitions between eruption style to be sensitively poised.

Observations at the Soufrière Hills Volcano have provided further insights into degassing processes in viscous magma systems. One of the most remarkable features of the eruption has been the cyclic patterns of activity marked by regular patterns of seismicity, ground deformation recorded by tiltmeter and eruptive behaviour (Fig. 5; Voight *et al.* 1999). The cyclic behaviour had typical periods ranging from a few hours to just over a day (Neuberg 2000). In some periods, the system locked on to a very regular period for up to a couple of weeks. Each cycle involved a swarm of shallow hybrid earthquakes with long-period components accompanying inflation of the ground localized around the dome. The inflation cycle would peak, and then a period of elevated activity would immediately follow, accompanying deflation. The nature of the elevated activity would vary. In many cases, there was an increase in rock-falls from the dome, and this was interpreted as evidence of enhanced dome extrusion rates (Voight *et al.* 1999). In some cases, strong ash-venting occurred

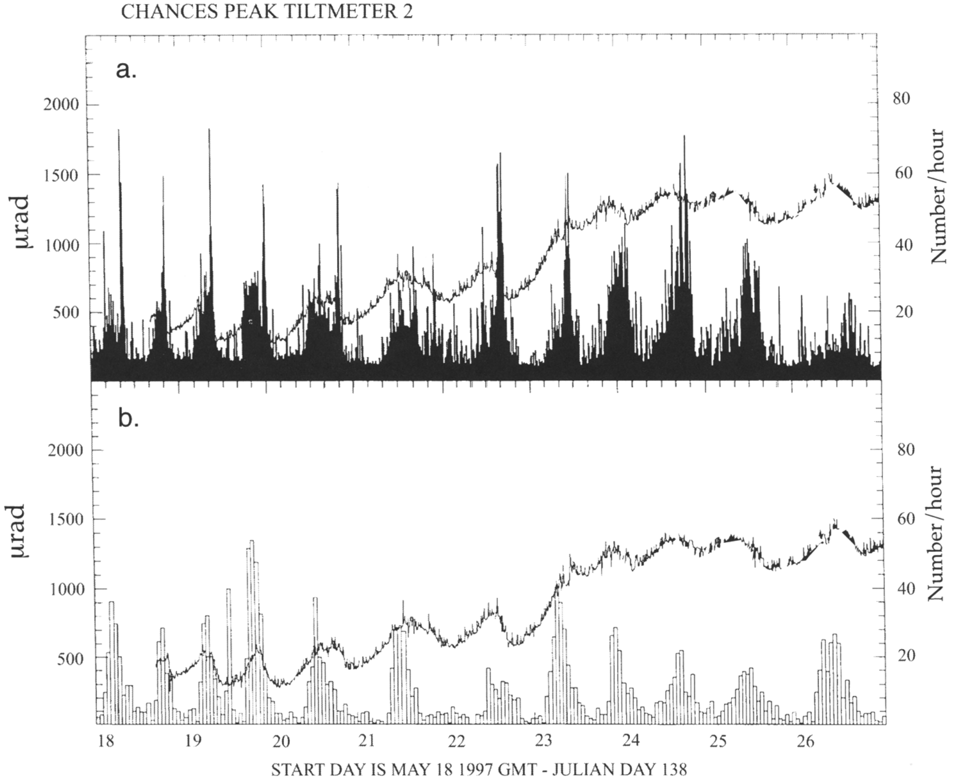

Fig. 5. The tilt pattern at Chances Peak in May 1997 (after Voight *et al.* 1999). The tiltmeter was approximately 400 m from the centre of the dome, with the tilt axis for data shown being approximately radial to the dome centre. The earthquake event frequency in events per hour (right hand vertical axis) at the Gage's seismometer is shown as histograms. The tilt variation in micro-radians (left-hand vertical axis) is shown as the continuous curves. The upper diagram (**a**) shows tilt data along the horizontal axis radial to the dome, and the lower diagram (**b**) shows tilt data along the horizontal axis tangential to the dome. All the instrument output displays the cyclic pattern of deformation and seismicity, with hybrid earthquakes occurring in the inflation periods and rock-fall signals occurring during the deflation periods. Marked episodes of degassing were observed at the peaks in the tilt cycle and during deflation (see Watson *et al.* 2000).

at the peak of the cycle and declined during deflation. Most spectacularly, Vulcanian explosions lasting a few minutes occurred at the peaks, followed by ash-venting for tens of minutes during the deflation. In early August 1997, 11 explosions occurred every 12 hours (Druitt *et al.* 2002). In September and October 1997, there were 75 explosions, with an average period of 9.5 hours between explosions. Watson *et al.* (2000) have also shown that the flux of SO_2 increases over the peak of cycles and, together with the vigorous ash-venting activity, implies that the cycles involve release of pressurized gas from within the dome and upper conduit. The tilt returns to almost the same position over a single cycle, indicating that the deformation is almost completely recovered. Release of pressurized gas provides the most compelling explanation of this recovery. Comparable cyclic patterns of explosive degassing were observed at Mount Pinatubo during lava dome extrusion prior to the paroxysmal phase of the eruption on 15 June 1991 (Denlinger & Hoblitt 1999)

The cyclic patterns at Montserrat can be interpreted as strong evidence for the pressurization of the rising magma (Sparks 1997). The cycles are interpreted as a build-up of a large overpressure in the upper parts of the conduit and the relief of that pressure at the peak of the cycle when some threshold overpressure is reached. In some cases, the relief of pressure involves a surge of lava extrusion. This might be a consequence of the overpressure reaching the yield strength of the non-Newtonian and highly crystalline andesite or some kind of stick-slip mechanism along the conduit wall (Denlinger & Hoblitt 1999). At other times, the pressure can be relieved by the escape of pressurized gases, as manifested in the ash-venting and elevated SO_2 fluxes. These observations thus illustrate a very potent way of degassing magma by vertical permeable flow. In the explosions the overpressure was sufficient to exceed the fracture strength of the magma, and degassing was accomplished in the most spectacular way.

The detailed mechanism of the pressurization and degassing cycles and the controls on their duration are not yet fully understood. As discussed further in the next section, large overpressures can be developed at shallow levels in ascending magma due to rheological stiffening (Sparks 1997; Massol & Jaupart 1999; Melnik & Sparks, 1999). There are a number of possible controls on the time-scales of the degassing cycles. Rheological properties may be important. Highly crystalline degassed magma is highly non-Newtonian. Uniaxial compression tests on samples of Montserrat andesite at eruption temperatures and pressures comparable with those expected in the conduit system have been carried out (A. M. Lejeune, unpublished data) and one of these experiments was presented in Sparks *et al.* (2000). In a typical experiment, the sample had a strongly non-linear response to a constant load pressure (Fig. 6). The initial period involves slow viscous deformation. Deformation accelerates with a decrease in apparent viscosity and development of microcracks, with failure taking place after several hours. Thus, the threshold conditions reached at the peak of a cycle and the time-scale of a cycle may partly be governed by the non-linear response of the non-Newtonian magma to the pressurization. The pressurization itself could be controlled by a number of effects. First, the pressurization may relate to gas exsolution. Bubble formation is governed by diffusion of gas from supersaturated melt and expansion of gas due to pressure decrease during magma ascent. Gas supersaturation can also be coupled to crystal growth (Stix *et al.* 1997; Sparks 1997), which concentrates gas into the melt. Bubble growth and gas pressure are also coupled to magma viscosity, which resists growth. In the upper parts of conduits, viscosities are very high, and considerable gas overpressure can develop (Sparks 1978; Navon & Lyakhovsky 1998; Massol & Jaupart 1999). Finally, the magma pressure is itself a function of flow speed, magma compressibility and viscosity (Massol & Jaupart 1999; Melnik & Sparks 1999) and can regulate gas exsolution and gas pressure.

There remain many uncertainties and outstanding problems in understanding conduit degassing of silicic magmas. The interactions of gas exsolution, gas escape and crystallization in magma flows are complex, because all of these processes greatly affect key magma properties such as density, rheology and compressibility. For example, Massol and Jaupart (1999) demonstrated that large lateral variations of these properties can develop across conduits during magma flow. Most models are one-dimensional, assuming averaged properties across the conduit. Thus, fully realistic models have yet to be developed, and represent a considerable challenge to volcano physicists.

Coupled crystallization and degassing in ascending magma

Although the large depression of the liquidus of magmas due to small amounts of water has been understood for a long time (e.g. Tuttle & Bowen 1958), the implications for volcanology have only recently become well established (e.g. Sparks & Pinkerton 1978; Cashman 1992;

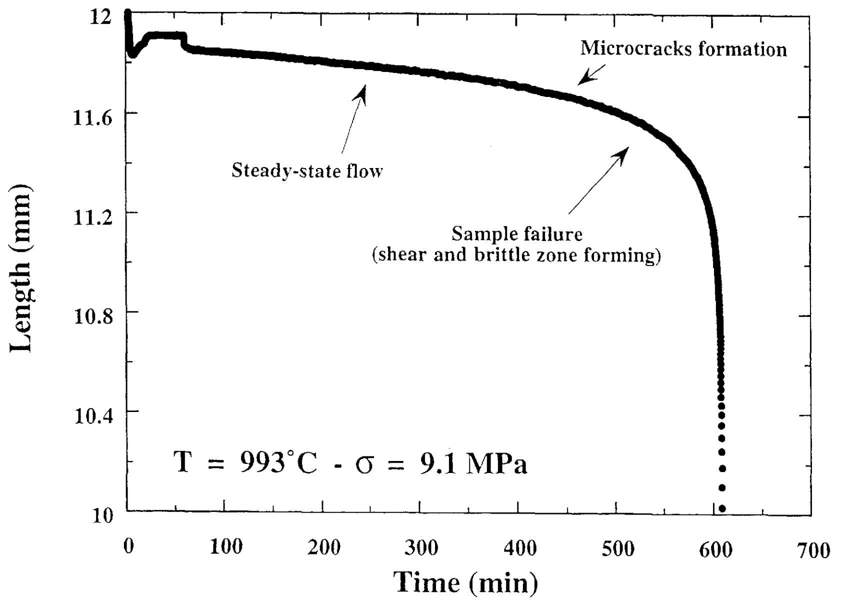

Fig. 6. Deformation curve of a sample of the Soufrière Hills andesite subjected to a uniaxial compression (9.1 MPa load pressure). The experiment has been performed at room pressure and at 993 °C on lava dome material erupted in January 1996. The plot shows the length of a cylindrical core of lava as a function of time (the initial sample length was 12 mm for a 4.8 mm diameter). Initially the sample deforms in a regular manner at a constant strain rate with a viscosity of about 10^{14} Pa s. After a few hours sample deformation accelerates and the apparent viscosity decreases substantially, with an estimated value of 10^{11} Pa s just before failure. The increasing strain rate results from a rising rate of formation of micro-cracks. Failure takes place along a shear surface, which develops from a zone of micro-fractures (after Sparks *et al.* 2000).

Geschwind & Rutherford 1995; Sparks 1997; Hammer *et al.* 1999; Blundy & Cashman 2001; Hammer & Rutherford 2002). The exsolution of water causes magmas to become strongly undercooled, and this induces spontaneous crystallization. As an example, the porphyritic Soufrière Hills magma is thought to start in the magma chamber with a melt content of about 35%, 4–5% water dissolved in the rhyolitic melt phase and a temperature of 850 °C. By the time that it reaches the surface at slow flow rates, the magma has effectively solidified due to degassing with only a few per cent residual melt (Sparks *et al.* 2000). During ascent, the magma changes from a Newtonian fluid, with an estimated viscosity of about 7×10^6 Pa s, to a hot partially molten solid with a strength of around 1 MPa and a power law rheology, in which the apparent viscosity varies with increasing strain rate from 10^{14} to 10^{11} Pa s (Fig. 6; Voight *et al.* 1999; Sparks *et al.* 2000). These degassing-induced changes dominate flow behaviour.

The dynamical model of Melnik and Sparks (1999, 2002) explores the coupling of crystallization, gas exsolution, gas escape and magma flow in lava dome eruptions, applying the results to the Soufrière Hills andesite. A key element of the model is the kinetics of crystallization. Their results show the same kind of sigmoidal relationship between magma extrusion rate and magma chamber pressure as Slezin's models (Fig. 7), but the reason for this structure is quite different. Here the upper branch represents the case where magma ascent rate is too rapid for crystallization to take place and so the viscosity remains relatively low, although the viscosity still increases as the magma ascends due to the effects of dissolved water on viscosity. The lower branch represents the case where the flow is slow and so there is plenty of time for crystallization. In this case, the magma can erupt as a solid. As before, the system can jump between these states of high and low flow rates close to the critical bends in the curves. In this case, however, the transition between fast and slow dome growth is expected to take place slowly, due to the viscous nature of the flows.

A key feature of magma ascent and degassing is that the rheological stiffening caused by degassing results in large overpressures being

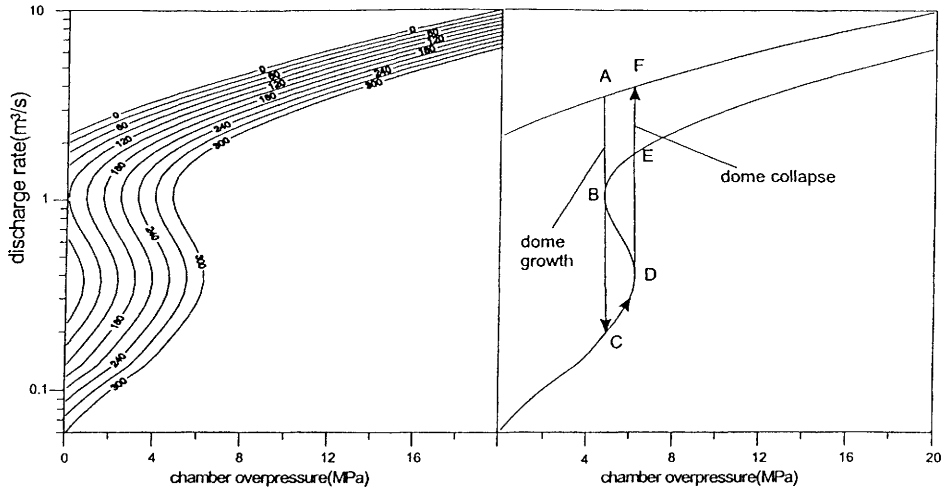

Fig. 7. The relationship between steady-state extrusion rate and magma chamber pressure. The curves were calculated by Melnik & Sparks (1999) for conduit conditions and magma properties appropriate to the degassing eruption of the Soufrière Hills andesite on Montserrat. The curves are characteristically sigmoidal as in Figures 2 and 3. However, here the upper, stable branch is for flows where the rate is too fast for crystallization to occur during magma ascent, and the lower branch is for flows where the flow rate is sufficiently slow that substantial microlite crystallization can occur. The behaviour is related to degassing as the microlite growth is triggered by gas exsolution. In (**a**) curves are shown for different dome heights in metres. In (**b**) the curves for 0 and 300 m from (a) are isolated, and a possible cause of cyclic activity is indicated. Starting at C on the 300 m dome height curve, the magma pressure increases and the discharge rate increases. At the cusp D, the steady-state flow rate must jump to E where the flow rate is much faster and the chamber pressure declines as magma is removed form the chamber. At B there is another cusp, and the steady-state conditions fall to C. If magma chamber pressure then increases, the cycle can be repeated. If at point D there was a dome collapse, then the jump could be to F on the 0 m dome height curve. After the collapse, further eruption may reduce the chamber pressure to A, and the flow rates once again fall back to C.

developed in the upper parts of the volcanic conduit (Sparks 1997; Voight *et al.* 1999; Melnik & Sparks 1999). Overpressure can be defined in two different ways. One definition is the difference between the pressure at some depth and the weight of the surrounding rocks (the lithostatic pressure). Another definition is the pressure difference between the local pressure at some depth and the weight of the overlying column of magma. Which of these overpressures is relevant depends on the circumstances. The second definition is relevant in considering magma ascent from depth and local overpressures that might exceed the strength of the magma and result in explosions. On the other hand, the first definition is relevant if one is interested in the deformation of the volcanic edifice and interpretation of ground deformation. There may not be much difference in the values of these overpressures, although they can be very different in certain circumstances at shallow levels.

In a system with constant viscosity and constant conduit cross-section, overpressure decreases linearly between the magma source region at depth and the surface. However, in a system with large vertical viscosity gradients the overpressure variation is highly non-linear and focused in the uppermost parts of the volcanic conduit where the high-viscosity degassed magma offers most of the frictional resistance to flow. Figure 8 shows the calculations of overpressure variation with height by Melnik and Sparks (1999) for eruption of Soufrière Hills andesite. The key result is that a large overpressure maximum up to several MPa can be predicted at depths of a few hundred metres. The value of the maximum depends on many parameters, and in Figure 8 the dependence on magma permeability is shown. The lower the magma permeability, the higher is the overpressure maximum. Even without crystallization, large overpressures can develop due to the strong decrease of viscosity at water contents below 1% (Massol & Jaupart 1999).

These ideas have been developed further by Barmin *et al.* (2002) to investigate the unsteady

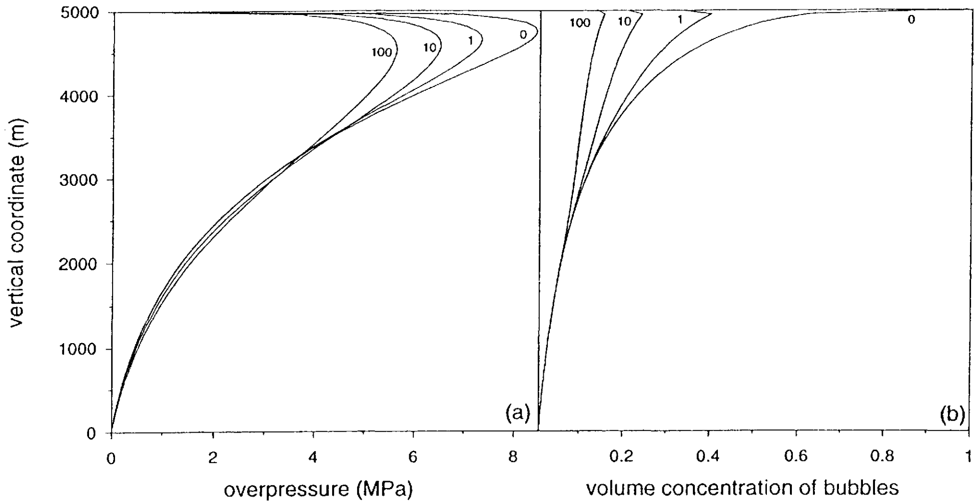

Fig. 8. Calculated profiles of overpressure (**a**) and porosity (**b**) along a conduit. Conditions for calculations are described in Melnik & Sparks (1999) and relate to conduit conditions and magma properties of the eruption of the Soufrière Hills andesite lava dome. Overpressure (the difference between the magma pressure and surrounding lithostatic pressure) reaches a maximum in the uppermost few hundred metres of a volcanic conduit. Curves are shown for values of the permeability coefficient k_0 from values of 0 to 100. This parameter is the constant coefficient in a power law that relates permeability to porosity (see Melnik & Sparks 2002). Values of k_0 are bounded in the range 1 to 10 for samples of the Soufrière Hills andesite (see Melnik & Sparks 2002 for details). Low values of k_0 give unrealistically high porosities at the top of the conduit, and explosive conditions are likely to develop in practice. The calculations here assumed no crystallization in the conduit.

behaviour of lava extrusions with the incorporation of visco-elastic deformation of the magma chamber and conduit wall for the case of an open system chamber being replenished at a constant rate. In the model of Melnik & Sparks (1999), the chamber pressure was kept constant but, with a flexible elastic chamber, the pressure can vary. This model generates a much richer range of behaviours (Fig. 9), with both steady extrusion and highly pulsed periodic extrusion. Barmin *et al.* (2002) have been able to mimic the pulsatory character of the 1980–1986 dacite dome extrusion of Mount St Helens and the 1922–present dome growth of Santiaguito, Guatemala. Such models are moving towards a more complete total system description of eruptions, in which magma chamber dynamics, degassing, degassing-induced crystallization, rheological stiffening, conduit flow and dome growth are coupled together. However, as mentioned earlier, fully realistic models that incorporate two-dimensional effects remain a future goal.

Degassing in magma chamber evolution

It is usually assumed that the principal ways in which magmas differentiate is by heat loss (e.g. Shaw 1985). This is a slow process because heat loss from magma chambers is by conduction, although hydrothermal circulation can greatly increase heat loss. The key point of this paradigm is that crystallization and differentiation are related to a time-dependent loss of heat, which is usually assumed to be sufficiently slow that large bodies of magma can develop and slowly differentiate in the crust. The role of degassing has been given far less attention, although the process is widely recognized and is occasionally mentioned in passing. Degassing, however, can result in large undercoolings and crystallization, and is therefore also able to explain crystallization and differentiation if crystals are separated from the melt. Gas exsolution in magma chambers can also have important dynamic effects: for example magma chamber pressure increases when gas exsolution occurs concurrently with crystallization (Tait *et al.* 1989) and gas exsolution may result in filter pressing of residual melts from crystal mushes (Sisson & Bacon 1999).

One mechanism of magma chamber differentiation that has not been given much attention is the role of convective circulation and drain-back of degassed magma. This has been dis-

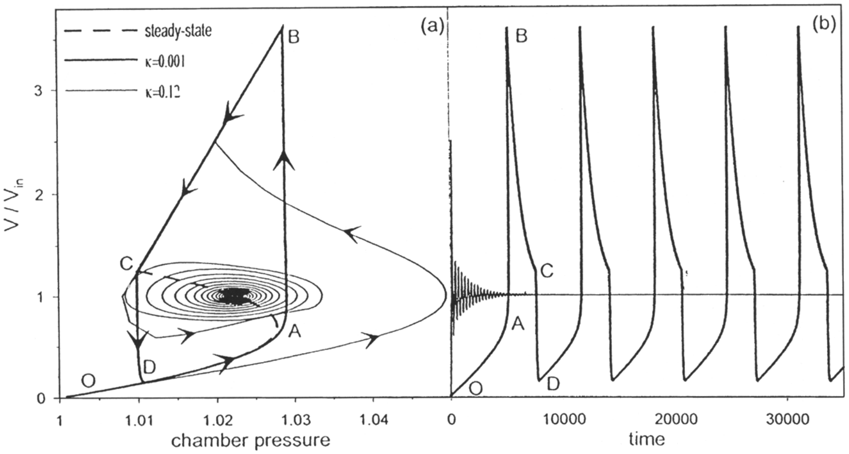

Fig. 9. Models of unsteady cyclic conduit flow and dome extrusion with degassing from an open-system magma chamber, after Barmin *et al.* (2002). In (**a**) the velocity, v, of magma ascent is plotted against magma chamber pressure, with the velocity being normalized by the velocity of input, v_{in}, of new magma into the chamber, and the magma chamber pressure being normalized by the lithostatic pressure at the top of the chamber. The dashed curve (O–D–A–C–B) shows steady-state solutions, and the two curves (thick and thin solid curves) show unsteady calculations for two values of the parameter, κ, which is a dimensionless number defined as $(\pi D^2\gamma/4V_{ch}\rho g)$ where D is the conduit diameter, γ is the rigidity of the conduit walls, V_{ch} is the magma chamber volume, ρ the density of the magma, and g is gravity. Large values of κ are for small chambers with relatively wide conduits and small values of κ are for large chambers with relatively narrow conduits. The equivalent time series for the discharge rate is shown in (**b**). For small κ (=0.001) the magma discharge is highly periodic, with short periods of very high discharge alternating with much longer periods of low discharge along the path A–B–C–D–A. For large κ (=0.12), initial rapid oscillations of discharge rate are damped and the system evolves to a steady-state output. In both cases the time-averaged output rate is equal to the input into the chamber (v/v_{in}=1). The calculations were carried out for magma properties similar to the dacite of Mount St Helens, and details can be found in Barmin *et al.* (2002).

cussed earlier in the context of explaining excess magma degassing in open conduits. However, another implication of the process is that degassed magma draining back into the chamber will develop strong undercooling not only because of degassing at the surface, but because the liquidus of dry degassed magma increases with increasing pressure. For example, a wet basalt with 2% water and a liquidus of 1050 °C has a liquidus of 1200–1250 °C in the chamber after a cycle of surface degassing and drain-back. Consequently, this process can drive substantial amounts of crystallization and differentiation in the chamber. Evidence for this process has been found at Kilauea on Hawaii, from studies of volatiles in lavas and melt inclusions (Dixon *et al.* 1991; Wallace & Anderson 1998). The importance of degassing may have been underestimated, and there is a case to be made that it can, in some circumstances, be the dominant mechanism of magma chamber crystallization.

An alternative view of intermediate to silicic magmatic systems is as follows. Magmas are generated, stored and evolved largely at deeper levels in the crust (Annen & Sparks 2002). The mechanisms of evolved melt generation involve simultaneous crystallization and partial melting as mantle-derived mafic magmas invade the deep crust. In these circumstances, some of the evolved melts may have very high water contents. Such water-rich magmas ascend to shallow depths, and degas during ascent. When they reach upper-crustal depths, they become increasingly viscous due to degassing and crystal growth. They may proceed directly to erupt at the surface, or stagnate to form shallow magma chambers where further degassing results in crystallization (Blundy & Cashman 2000). Wallace (2001) has recently reviewed evidence for large amounts of exsolved gas in many shallow magma chambers. The very detailed study of volatile inventories in the Bishop Tuff magma chamber (Wallace *et al.* 1999) provides strong evidence for the major role of an excess exsolved volatile phase. A major exsolved volatile phase in magmas can explain the otherwise puzzling excess degassing of SO_2 in many eruptions as suggested by several studies

(Gerlach *et al.* 1996; Scaillet *et al.* 1998; Keppler 1999). There are other important implications for magmas having higher volatile contents than has been supposed and, as a consequence, of undergoing coupled crystallization and degassing in chambers. Such notions will influence concepts about formation of ore deposits, leakages of deep volcanic gases in hydrothermal fields, and the dynamics of eruption. In particular, a chamber with large amounts of exsolved gas will be highly compressible, and this will make large differences to the evolution of magma chamber pressure (Huppert & Woods 2002).

Degassing also can have a major role in the dynamics of open-system magma chambers. Replenishment is widely accepted as a major factor in triggering volcanic eruptions (e.g. Sparks *et al.* 1977; Blake 1981, 1984). The presence of exsolved gas in the incoming magma may reverse the usual density relations in which mafic magma is denser than more evolved magma. The presence of exsolved gas can lead to immediate magma mixing (Phillips & Woods 2001), gradual mixing (Eichelberger 1980), catastrophic overturn and mixing (Huppert *et al.* 1982; Turner *et al.* 1982) or gradual transfer of gas without mixing, from more mafic magma emplaced at the base of the chamber to more silicic overlying magma (Wallace 2001).

Magma volatile contents and depth of degassing

In the discussions of magma degassing so far, not much has been said about the typical volatile contents of magmas and the depths at which gas exsolution and escape becomes important. An important issue is: at what depths do magmas become sufficiently permeable for gas to start to escape? Figure 10 shows calculations of the pressures and approximate depth equivalents at which magma reaches 30% porosity for different water contents. A porosity threshold of 30% has been chosen, as this is approximately the value above which significant bubble interaction and permeability development might be expected (Blower 2001). Here, 30% is simply used to illustrate that the depth at which gas escape becomes important will be a strong function of water content.

There are grounds for thinking that water contents of magmas may have been underestimated, particularly in arc magmas. Melt inclusion data and experimental studies on arc basalts, for example, indicate water contents in the range 2–6% (e.g. Sisson & Layne 1993; Moore & Carmichael 1998; Roggensack 2001). Inspection of Figure 10 indicates that significant degassing processes can start at depths of several

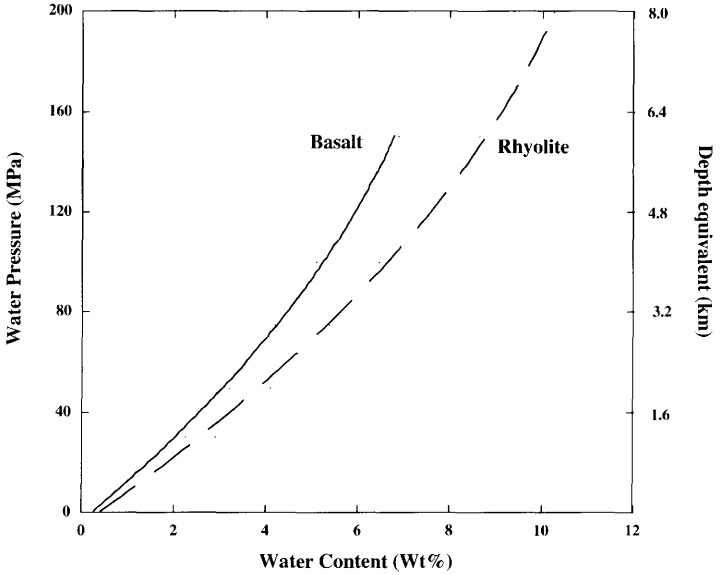

Fig. 10. Plots of water content of magma (melt) against pressure for magma with a vesicularity of 30 vol. %. Curves are shown for basalt and rhyolite with assumed solubility constants of 3 and 4.1×10^{-6} $Pa^{-1/2}$ respectively. Magma density and temperatures are assumed to be 2300 kg/m^3 and 800 °C for rhyolite and 2600 kg/m^3 for 1100 °C basalt. The approximate depth equivalents of the pressures are shown on the right-hand vertical axis, assuming a crustal density of 2500 kg/m^3.

kilometres in water-rich basalts, as also deduced by Roggensack (2001). For intermediate and silicic magmas, numerous studies of melt inclusions and experimental simulations (see Wallace 2001 for review) indicate water contents invariably in the 4–6% range. However, these water content estimates are lower limits, because they are determined by study of phenocrysts that trap melt inclusions and by comparison of phenocryst assemblages with experimental products. A critical problem is that magmas with water contents significantly in excess of 6% may not crystallize during ascent. Experimental studies (e.g. Holtz & Johannes 1994) indicate that very wet magmas (6–12% water) do not crystallize during ascent until the water pressure is sufficiently low, and this happens at pressures where only 6% or less remains dissolved in the melt. If the wet basalts typical of many arcs were to stall and differentiate in the deeper parts of the crust, then evolved intermediate and silicic melt with very high water contents could be generated. Inspection of Figure 10 indicates that these very wet magmas can become very bubble rich at depths of several kilometres, and connectivity can develop in the chamber.

Conclusions

Degassing has a profound effect on the dynamics of volcanic eruptions. Gas exsolution as magma ascends causes large changes in physical properties, such as density and viscosity, and phase equilibria, with spontaneous crystallization resulting from gas loss. The compressibility of bubble-rich magma also greatly complicates the dynamics of flows in conduits and degassing processes. The coupling of gas exsolution, gas separation processes, crystallization and compressibility effects lead to a very rich diversity of flow and eruptive phenomena. The processes of degassing and crystallization are kinetically controlled, and their coupling with flow, largely through their effects on density and viscosity, is highly non-linear. This makes volcanoes classically dynamical systems, in which strong feedback loops, strong unsteadiness, and strong and sometimes catastrophic jumps in flow regime, are intrinsic. The range of behaviours can be from steady discharge, to highly periodical patterns, to more chaotic and complex behaviours.

Volcanic systems can be inherently unpredictable, as a consequence of non-linear dynamics. Perhaps the most important concept, which emerges from this analysis of degassing dynamics, is the discovery that a completely determined volcanic system with all properties known precisely can have multiple stable states. It is not possible to determine which state will actually occur, without knowing the full dynamical history of the system. This concept was pioneered by the research of Y. Slezin, and has subsequently been developed in more elaborate models of magma ascent and eruption dynamics (e.g. Jaupart & Allegre 1991; Massol & Jaupart 1999; Melnik & Sparks 1999; Barmin *et al.* 2002). Of course, none of the parameters that control eruptions are known precisely, and indeed, some critical parameters, such as chamber size, may be highly uncertain. These results indicate that prediction of volcanic eruptions can only be developed in a probabilistic way, in which uncertainties are taken into account. The notion that volcanic eruptions can be inherently unpredictable is not a comfortable truth, and will require some changes in thinking about the goals of volcanology.

On the other hand, these same dynamical models can provide an explanation of beautifully periodic behaviours and some basis for forecasting. Sudden changes of state in an eruption need no longer be surprises or perplexing phenomena. Also, the dynamical flow models provide the conceptual framework for interpreting geophysical, geochemical and phenomenological data monitored in eruptions. For example, models, such as those of Melnik and Sparks (1999), show how large overpressures develop in the uppermost parts of volcanic conduits, thereby explaining several geophysical features.

Another important concept is that degassing may be a major factor in the differentiation in magma. Although it has long been understood that magmas crystallize when they exsolve the major volcanic gas (water), the implications have perhaps been overlooked. The majority of petrological and geochemical studies implicitly interpret data in terms of the concept of a shallow, slowly cooling magma body, where heat loss is the driver for crystallization and fractionation. Of course, heat loss undoubtedly occurs, and there is no suggestion here that it is not going to be an important process in the evolution of igneous rocks. However, degassing is also likely to be important in at least two ways. First, water-rich magmas spontaneously crystallize if emplaced into a shallow magma chamber, and this is clearly an effective way of differentiation to generate evolved melts. Second, the only way to explain huge volumes of degassing with little eruption of magma is by very effective convective exchanges between the surface and deep chambers along conduits (Kazahaya *et al.* 1994; Stevenson & Blake 1998). An overlooked process is that sinking degassed magma must crystallize due to compression, and this can also be a driver for differentiation. Finally, magma volatile contents have evidently been underestimated, and some magma chambers may contain large amounts of exsolved gas (Wallace 2001). The presence of substantial gas in magma chambers will profoundly affect how magmas behave and erupt. I also draw attention to the possibility that water contents of magmas, particularly in arcs, may have been underestimated. Gas loss processes in water-rich magmas can initiate at depth in volcanic conduits and even in the magma chamber. Degassing is not necessarily a shallow process.

First of all I remember Pete Francis as an outstanding scientist and good friend, and I dedicate this paper to his memory. I thank all my many collaborators who have made contributions to understanding degassing. I first acknowledge the close collaboration with staff of the Montserrat Volcano Observatory during the eruption. The ideas and data from that eruption were a team effort, involving a large number of dedicated and outstanding people. I particularly recognize the pioneering research of Y. Slezin, and acknowledge the collaboration with O. Melnik over the last few years. Oleg's mathematical skills and physical insights have been pivotal in developing models of the interactions between degassing and rheological stiffening in controlling conduit flows and eruptive behaviour. P. Wallace is thanked for his comments. C. Jaupart, S. Blake, C. Connor and T. Gerlach are thanked for their helpful reviews. I acknowledge the support of a NERC fellowship.

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