Chapter 1

Magma Degassing and Fragmentation: Recent Experimental Advances

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ABSTRACT

Recent experimental studies of the physico-chemical processes thought to be involved in the degassing of explosively erupting silicic magmas are discussed. These include bubble nucleation, bubble growth, accelerating two-phase flow, brittle failure and post-fragmentation effects. The nucleation of bubbles in ascending crystal-bearing magmas is likely to be heterogeneous and mode-specific. The growth of bubbles appears to pass through three stages of kinetic control switching between an early viscous control stage, a subsequent diffusive control stage and a possibly ending with a final return to viscous control, at the substantially higher viscosities resulting from the degassing process itself. The two-phase bubbly flow of foaming magma is accompanied by acceleration to the fragmentation event. The fragmentation may occur by brittle failure of the magma as the tensile strength is overcome by internal and/or external stresses. The pyroclasts generated have particle characteristics whose systematic dependence on fragmentation conditions is being investigated experimentally. Postfragmentation effects include subsequent textural modification of even silicic magmas as bubbles grow further and/or collapse and an ultimate thermal history stage of cooling across the glass transition that may be multiple in character and occurs over a demonstrably wide range of cooling rate.

1. INTRODUCTION

The degassing and fragmentation of silicic magmas are processes that contribute to and often control a remarkable range of the chemical and physical consequences of magmatism. They also influence a host of related phenomena including the economic concentration of elements, the evaluation of local hazard potential of explosive and effusive volcanism, the impact of volcanic ash and gases on the Earth's climate and the evaluation of volcanism on terrestrial planets and satellites within the solar system. Taken together both processes encompass a relatively complex series of events in the evolution of silicic magmas beginning with the saturation of the magma in a volatile phase and ending with its explosive ejection from the volcanic vent. The basic controls on the processes that are likely to determine the degassing of magma are generally agreed upon. They are (1) the P-T-X dependence of volatile solubility and partitioning, (2) the kinetics of nucleation and growth for bubbles in the magma, and (3) the development of permeability in the magma and the surrounding country rocks. The basic processes controlling the fragmentation of magma are not generally agreed upon. The possibilities essentially fall into two scenarios, one in which the magma evolves to a textural maturity at high vesicularity (i.e., foams) and thin film instabilities permit magma fragmentation or disaggregation and one in which magma is fragmented brittlely by stresses that exceed the tensile strength of the magma.

This chapter is concerned with recent developments in the understanding of the degassing and fragmentation of silicic magmas. It contains discussions of the processes involved in these late-magmatic systems from bubble nucleation to post-fragmentation processes. The goal of the presentation is to provide the reader with an appreciation of the physico-chemical concepts envisaged to be involved in each step of the process.

The reader is asked to bear in mind that, due to its central role in the evolution and behavior of volcanic systems and their plutonic equivalents, the degassing and fragmentation of magma is an extraordinarily active area of scientific research. Almost one half of the over 100 publications listed at the end of this chapter have been published in the 24 months immediately preceding the submission of this contribution. This chapter attempts to summarize the status as of December 1997. Closely allied topics not covered here but also available in this volume include conduit dynamics (Chapter 3), magma-water interactions (Chapter 2) and compressible two-phase flow (Chapter 7).

2. NUCLEATION

Experiments to determine the nucleation behavior of hydrous calcalkaline rhyolites with high (4-5 wt.%) water contents have been performed by Hurwitz and Navon (1994). They subjected natural samples to hydration by hydrothermal synthesis, pressure drops and dwells and subsequent quenchs. Their results indicate that heterogeneous nucleation prevails in these melts. In crystal-free melts the nucleation of bubbles is absent below pressure drops (ΔP) of 10 MPa, modest between 15 and 70 MPa and vast at > 80 MPa. The identity and to a lesser extent the morphology of crystals in microlite-

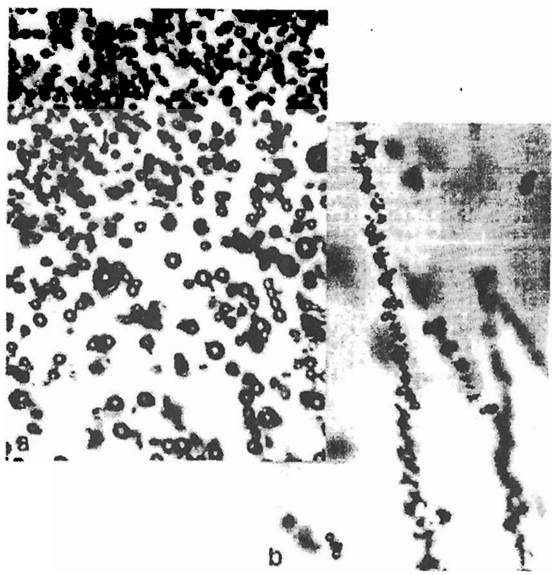

Fig. 1. Bubbles wetting the crystal-melt interfaces of crystalline phases in a calcalkaline rhyolite. The control of pre-existing crystal-melt interfaces on the localisation of nucleated bubbles is evident. Such information is evidence that the nucleation of bubbles in crystal-bearing magmas will be heterogeneous in nature. Reproduced from Hurwitz and Navon (1994).

bearing melts influences nucleation with oxide-melt surfaces providing very efficient nucleation sites, even at $\Delta P < 1$ MPa. Biotite, zircon and apatite nucleated bubbles at $\Delta P = 30$ MPa (Fig. 1). Feldspar appears to be relatively inefficient in bubble nucleation (Oded Navon, pers. comm. 1996). Thus the micro- and macrocrystalline phases in ascending magma are likely to be strong controls on nucleation kinetics. This raises the possibility that the phase equilibria and factors controlling them in ascending magmas may be directly linked to the degree of oversaturation encountered at the nucleation event. Recent experimental isothermal decompressive nucleation investigations of very crystal-poor ("virtually crystal-free") rhyolitic obsidians appear to indicate, in contrast, that significant oversaturations are possible (up to 150 MPa, Mourtada-Bonnefoi and Laporte, 1997).

3. BUBBLE GROWTH

As summarized by Navon et al. (1998) the growth of bubbles in silicic magma may be usefully subdivided into three stages. The initial growth of bubbles at sizes near the critical nucleus size is controlled by viscous stresses because the diffusive transport of water into such small bubbles with very high surface area to volume ratios poses no obstacle to growth. The duration of this initial stage is on the order of the viscous stress relaxation timescale ($\tau_\eta = \eta/\Delta P$). The subsequent growth of bubbles is driven by a slight pressure differential between bubble and melt and rate-controlled by the diffusion of water from the effective melt reservoir of dissolved water adjacent to the growing bubble to the bubble-melt interface. Ultimately, for the restricted reservoir of water available to the multiple bubble events with which we must be concerned, the reservoir becomes depleted and the bubble growth rates slow to generate the sigmoidal curves experimentally observed below. The viscosity of the melt in the latter stages of bubble growth can rise to levels where the viscous relaxation of the liquid again becomes an obstacle to bubble growth, the so-called "viscosity quench".

After a long period of repose following the pioneering experimental work of Murase and McBirney (1973) and the theoretical treatment of Sparks (1978), where the issue of viscous resistance to bubble growth was first raised, quantification of the vesiculation of rhyolitic magmas has received a great deal of attention in the past few years, in experiments (Bagdassarov et al., 1996; Lyakhovsky et al., 1996), in analytical and numerical calculations (Barclay et al., 1995; Proussevitch and Sahagian, 1996; Lyakhovsky et al., 1996;) and in textural analyses of rhyolitic bubble suspensions and foams (Toramaru, 1989, 1990; Klug and Cashman, 1996, 1998).

The experimental basis consists of experiments where the rate of growth of bubbles has been measured both individually, on quenched products using microscopy (Bagdassarov and Dingwell, 1993b; Lyakhovsky et al., 1996) or *in situ*, using video techniques (Bagdassarov et al., 1996) and cumulatively *in situ*, using dilatometric techniques (Bagdassarov et al., 1996; Stevenson et al., 1997).

Theoretical models have focussed on the relative influences of the viscous and the diffusive controls on bubble growth. The Peclet number ($Pe = \tau_d/\tau_\eta = \Delta P r^2/\eta D$) represents the ratio of characteristic timescales for water diffusion ($\tau_d = r^2/x$) and viscous relaxation (see above) in the melt phase. For Pe > 1 the growth rate of bubbles is controlled by the diffusion of water to the melt-bubble interface whereas for Pe ≤ 1, viscous stress relaxation in the surrounding melt controls bubble growth and significant bubble overpressure can result. Proussevitch et al. (1993) developed numerical methods for the simulation of bubble growth for cases where the Peclet number is near 1. Their simulation incorporated treatment of the effects of finite reservoirs of water (neighbour effects) and dealt with the possibilities of viscous resistance of the melt and significant advective flux of the hydrous melt towards bubbles due to the viscous flow. Their model is summarized in Figs. 2 and 3.

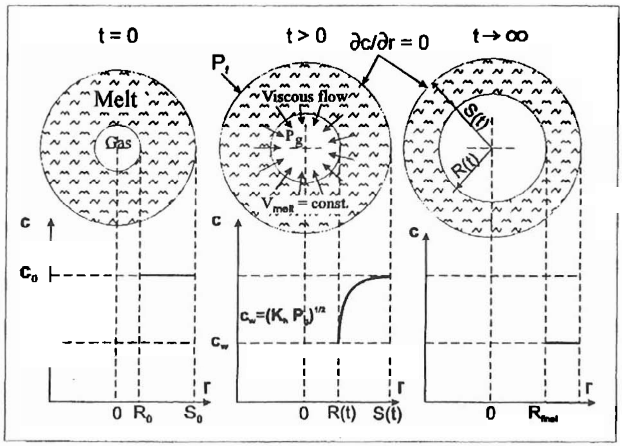

Fig. 2. Sketch of the physical constraints behind the approach to the modelling of bubble growth in viscous liquids of Proussevitch et al. (1993). The bubble is surrounded by a so-called elementary cell whose edge serves as a numerical barrier to mass flow but the cell itself is free to expand. The rate of diffusion and the resulting concentration profile are determined by the saturation concentration at the bubble edge which is in turn determined by the bubble pressure (a sum of ambient, surface tension and dynamic terms). At t_{final} bubble growth ceases as the bubble pressure and oversaturation in the melt reach equilibrium. Reproduced from Proussevitch et al. (1993).

Amongst the primary conclusions of the simulations of Proussevitch et al. (1993) are the following, 1) bubble separation controls the complete growth times and final separations of bubbles; 2) advective flux can generate shorter total growth times for larger bubbles; 3) confirmation that viscosity can play a significant role in bubble growth for dacitic to rhyolitic melts. They also distinguished two fundamental styles of bubble growth, an initially slow, but later very rapid gas exsolution (at low ambient pressures and large difference between initial and final bubble radii) and a relatively gradual degassing and bubble growth (at relatively high pressures, low oversaturations and small difference between initial and final bubble radii). The former would be likely to be fragmented in the early degassing history (textural criterion, see below) whereas the latter is not, in their opinion, relevant to explosive eruptions.

A discussion of the work of Proussevitch et al. (1993) by Sparks (1994) and the subsequent response by Sahagian et al. (1994) settled on the likelihood that the delay in bubble growth observed in the simulations of the former authors, and attributed by them to surface tension effects, had its origin instead in a viscous resistance term which rapidly diminishes in significance when the bubbles increase in radius (i.e., the initial stage of bubble growth noted by Navon et al., see above).

Analytical models for bubble growth during "instantaneous" decompression have been performed by Barclay et al. (1995) for the case of isolated bubbles (infinite melt reservoir) and of neighboured bubbles (melt shell reservoir) in a rhyolitic magma. They

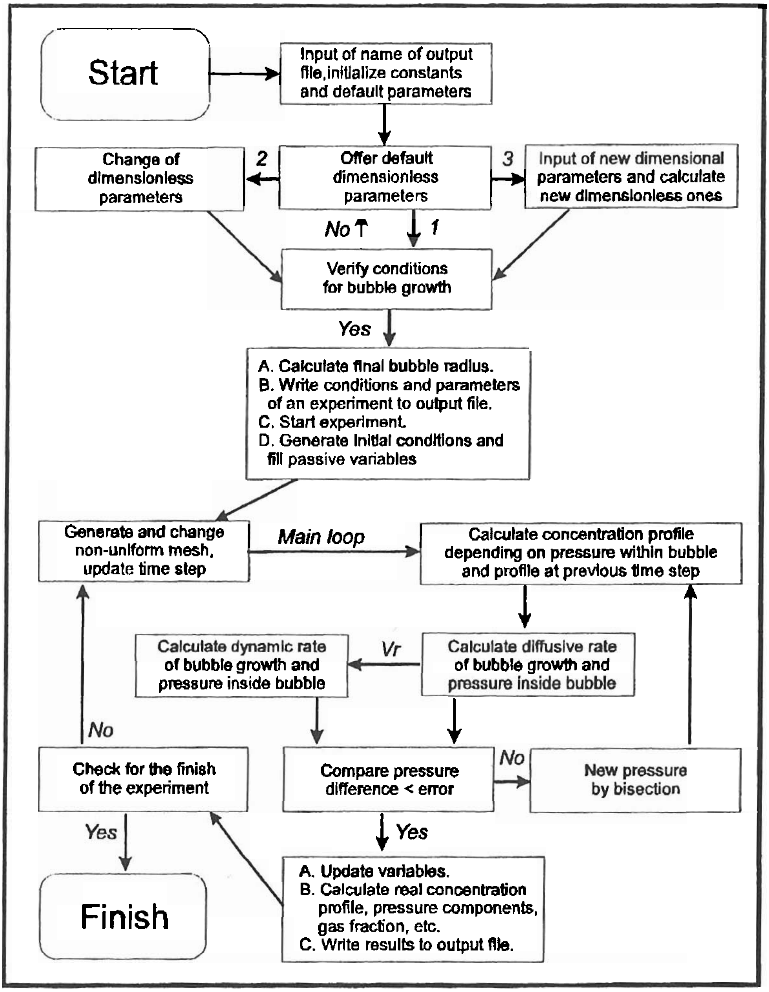

Fig. 3. A flowchart of the numerical scheme used to model bubble growth in viscous liquids. Reproduced from Proussevitch et al. (1993).

conclude that the viscosity of the melt plays no significant role in the bubble growth process until the viscosity reaches 10^9 Pa s.

Proussevitch and Sahagian (1996) recently treated the case of bubble growth in magma ascending at constant decompression rate for basaltic and rhyolitic cases. Their simulations for the rhyolitic case are illustrated in Fig. 4. The major conclusion of their analysis was that the decompression rate plays a very sensitive role in controlling the degassing efficiency. They state that for ascent rates below 0.1 m/s, there is equilibrium degassing, i.e., no significant volatile oversaturation whereas for ascent rates greater than 10 m/s no significant degassing occurs until the magma reaches the vent. In the relatively narrow range of ascent rates between these extremes there is no oversaturation at the surface for 0.1 - 1 m/s but significant oversaturation at 100 m depth; there is significant oversaturation in erupted rhyolite for 1 - 10 m/s but the oversaturation can reach its maximum at very shallow depths (down to a few m).

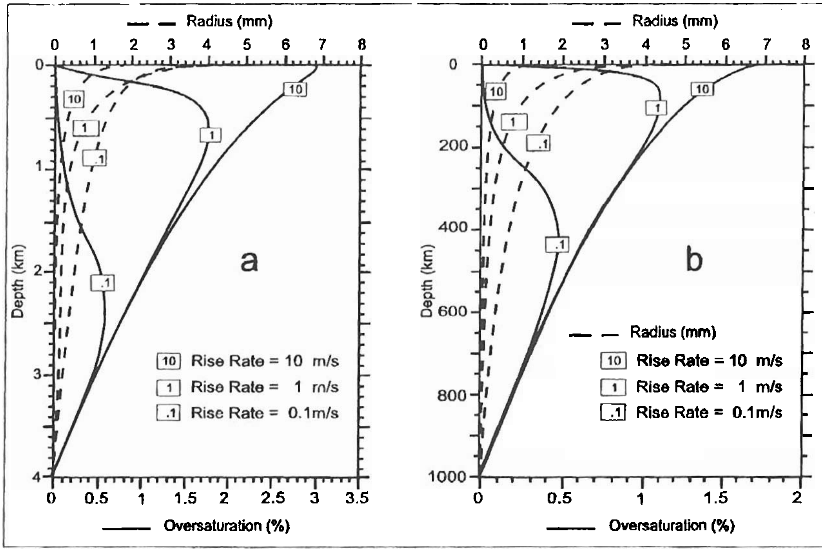

Fig 4. Bubble growth and water oversaturation in an ascending rhyolitic magma. The curves define oversaturation in % as a function of the depth in the system during magma ascent. The labels on the curves refer to the ascent or rise rates. The initial conditions are 4 km (a) and 1 km (b), which correspond to initial water concentrations dissolved in the magma of 3.72 and 1.86 wt.%, respectively. Reproduced from Proussevitch and Sahagian (1996).

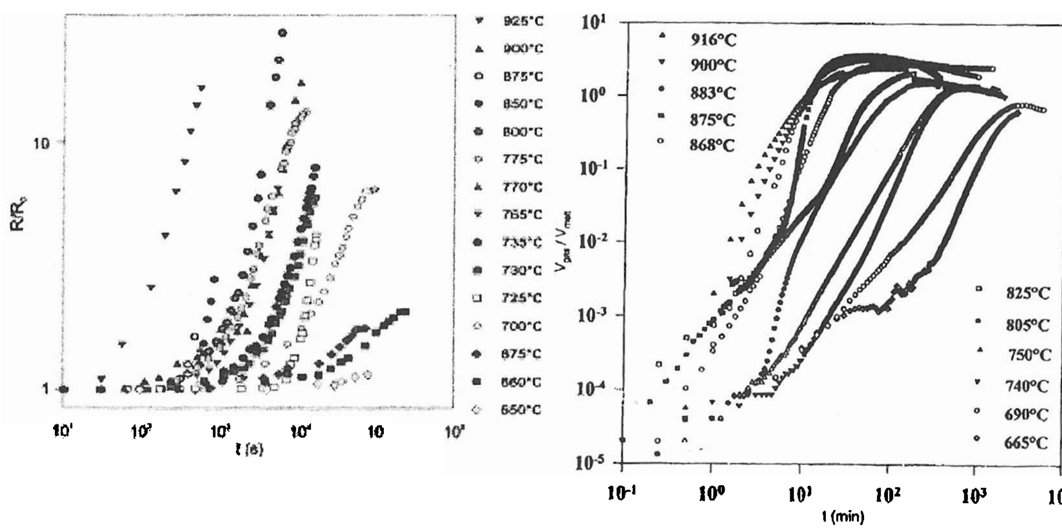

Fig. 5. In situ individual (video) and cumulative (volumetric) data for the growth of bubbles in melts with Peclet numbers near 1. The left dataset are optical data and the right dataset are volumetric data, for the logarithm of the relative radius versus the logarithm of time. The characteristic sigmoidal shape reflects the three stages of bubble growth as discussed by Navon et al. (1998) who replotted this data to obtain a semi-logarithmic dependence of r/r_0 versus time, indicating exponential growth. Reproduced from Bagdassarov et al. (1996).

The experimental analysis of melt vesiculation by Bagdassarov et al. (1996) demonstrated the sigmoidal shape of the individual bubble growth curves expected from bubble-bubble interaction considerations (Fig. 5). They also showed that for the case of viscous melts (Peclet number < 1) the temperature dependence of the induction time and characteristic bubble growth time from the Avrami equation reflect the activation energy of viscous flow for the melt. [The Avrami equation $\Delta V(t) \sim 1\text{-}\exp(-(t/\tau_{av})^n)$ describes the time dependence of volume expansion as a result of bubble growth, where V is volume, t is time, τ_{ac} is the characteristic time for growth and n is an exponent with value between 2.5 and 4.] This strongly implies that the relaxation of viscous stresses controls the growth rate and possibly the nucleation rate of bubbles in this regime. Bagdassarov et al. (1996) were able to demonstrate a transition from this viscously controlled bubble growth regime to a higher temperature regime where the activation energies of the same parameters were close to that of water diffusion, implying diffusion-controlled growth under the conditions of the latter regime. This transition occurs in the viscosity range of $10^{7.5}$ to 10^9 Pa s (Fig. 6).

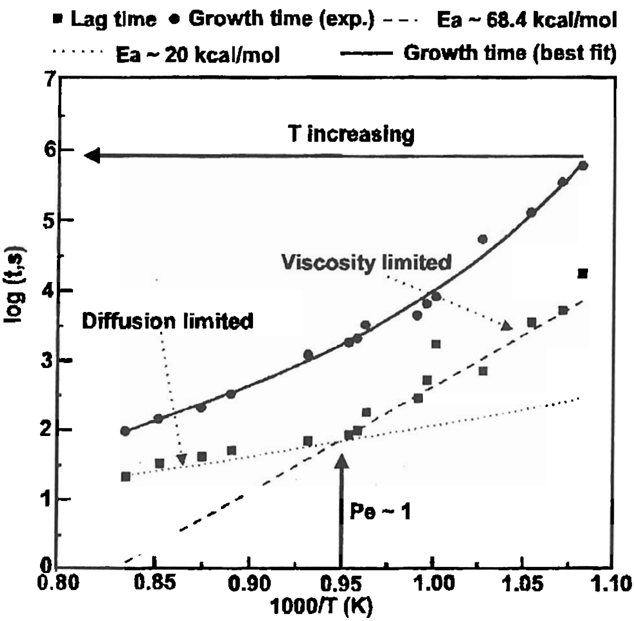

Fig. 6. The temperature dependence of parameters of the Avrami equation (fit to volumetric data) illustrate the transition from viscous to diffusional control of the bubble growth behavior. Reproduced from Bagdassarov et al. (1996).

The growth of bubbles following an instantaneous pressure drop and assumed instantaneous nucleation in the water-rich calcalkaline samples of Hurwitz and Navon (1994), were investigated using optical techniques similar to those of Bagdassarov et al. (1996) by Lyakhovsky et al. (1996). Their analysis indicates that at low number densities and short run durations, a growth model for solitary bubbles describes the growth data well, whereas at higher number densities, and indeed for the majority of their experimental data, the influence of bubble-bubble interaction invalidates the solitary bubble approach. Lyakhovsky et al. (1996) modelled their data in terms of their own modification of the growth model of Proussevitch et al. (1993) achieving good agreement and concluded that in the viscosity range of 10^4 to 10^6 Pa s (corresponding in their experiments to 3-6 wt.% water) that bubble growth is in the diffusional control regime. They emphasize that the supersaturation, controlled (in their experiments) through the initial water content and the final pressure, is a critical factor defining growth behavior because it influences nucleation efficiency and thus the bubble separation. Initial separation in turn ultimately controls the spatial distribution of water and pressure in the melt phase.

Stevenson et al.s (1997) experimental investigation of the degassing of a water-rich natural obsidian demonstrated, in addition to sigmoidal curves of bubble volume expansion reflecting the second and third regimes of bubble growth of Navon et al. (1998), significant fracturing associated with the bubble growth. This fracturing results from the strong increase in the glass transition temperature in the melt phase surrounding the growing bubbles as a result of the diffusion of water from the melt phase. Such fracturing may ensue in highly degassed melts in nature when the residual water contents have dropped below several tenths of a wt.%.

In a very recent analysis of experimental results for relatively viscous (< 10^8 Pa s) melts in the initial stages of bubble growth, Navon et al. (1998) demonstrate that bubble growth rate data can be well-modelled considering the supersaturation pressure and its consequences for viscous relaxation of the melt. They define a timescale ($\tau_{\eta d}$) which corresponds to the transition from an accelerating regime of exponential growth rate to a later diffusion-controlled regime where the growth tends towards a square root radial dependence. Fig. 7 schematically illustrates the transition from exponential to square root scaling of the growth process. They conclude that relatively degassed rhyolites near the Earth´s surface are viscous enough such that the value of $\tau_{\eta d}$ may range from hours to days. Navon et al. (1998) estimate the stored energy involved during the initial viscously controlled growth phase and infer that such overpressured bubble growth regimes may control dome explosivity timeframes in active dome growth events.

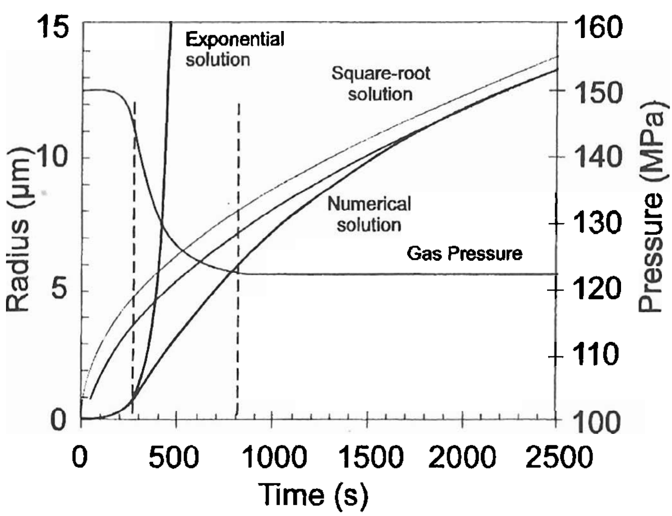

Fig. 7. Bubble growth at constant final pressure. The curves for exponential and square root dependence of the bubble radius versus time are shown together with modelled results for bubble growth illustrating the transition from exponential to square root growth law in the vicinity of $\tau_{\eta d}$ (vertical dashed lines). Reproduced from Navon et al. (1998).

The thermal consequences of the degassing process have been treated by Sahagian and Proussevitch (1996) using the thermodynamic basis for albite-water provided by Burnham and Davis (1971, 1974). Their analysis is divided into the endmember cases of "equilibrium" degassing where the vesiculating melt phase remains at the saturation water content during the entire process and the endmember possibility of disequilibrium degassing where the degassing occurs via vesiculation at constant pressure following an instantaneous pressure drop into the oversaturated liquid state. The two sources of thermal effects are the PV work of bubble growth and the heat of vaporisation or exsolution. They demonstrate that the heat of vaporisation is relatively large at low pressures (up to 20 kJ/mol at 10-20 bar) and decreases with increasing pressure to be negligible at pressures of several kbars. They suggest that at rapid decompression of magma the thermal effects of vesiculation may outrun the ability of the thermal conductivity to homogenize the temperature field, causing significant local undercooling of the melt at the bubble surfaces. Possible consequences of this undercooling are the potential for brittle behavior of the bubble walls (see also Mungall et al., 1996) and a thermal boundary layer impediment to volatile diffusion into the growing bubbles. The former could lead to explosive fragmentation whereas the latter could enhance the degree of oversaturation (and thus disequilibrium) during degassing.

4. ACCELERATING TWO-PHASE FLOWS

The application of shock tube techniques to the investigation of accelerating two-phase flows has been conducted in the past few years using two fundamental apparatus on liquids of very low viscosity. The first is a shock tube operating with H_2O-CO_2 mixtures in which the sample is decompressed by piercing a membrane mechanically causing exsolution of CO_2 and accelerating two-phase flow. The second contains a mechanism by which rapid concentric rotation of a barrier brings highly reactive components (HCl and K_2CO_3) into contact causing explosive reaction and accelerating two-phase flow (a foam containing CO_2 vesicles). The results of experiments with both sorts of devices have been discussed by Sparks et al. (1994) and Mader et al. (1994) and experiments of the latter type have been described more recently in greater detail by Mader et al. (1996). The primary observations in such systems are that the vesiculation of the liquid to generate CO_2 bubbles is accompanied by strong acceleration of the flow which generate a strong longitudinal (extensional) strain in the two-phase flow and that fragmentation follows this acceleration by the rupture of bubble walls in these extremely fluid systems. The growth rate of the column follows a $t^{2/3}$ rather than the $t^{1/2}$ dependence on time predicted by bubble growth models (see above). Mader et al. (1994) interpret this as resulting from enhanced diffusion due to convection via buoyant rise in their experiments. The vesiculation is highly heterogeneous and generates high supersaturations. The fragmentation of the "melt" phase (here water) into a spray in the liquid state demonstrates the feasibility of the mechanism on an empirical basis but at the strain rates generated in these experiments, viscoelasticity of rhyolites would be encountered possibly leading to brittle fragmentation. Clearly the results of these experiments are more consistent with the driving of mass through a fragmentation threshold via acceleration rather than the vertical migration of a fragmentation front in a volcanic conduit.

Further investigations of this sort by Mader et al. (1996) emphasize additionally the potential role of bubble deformation in enhancing growth rates and acceleration and raise the issue of the significance of tube pumice in volcanic products. The tube pumice points fundamentally to the longitudinal nature of deformation in many explosive events and the basal fractures responsible for fragmenting tube pumice into discrete lengths indicate tensile stresses greater than the tensile strength of the magma, thus brittle failure rather than ductile film instability. The question remains as to the frequency of occurrence and significance of the structures in tube pumice for the fragmentation event. If the basal terminations of tube pumice represent primary fragmentation then they may be evident of precisely the sort of accelerating flow witnessed in the shock tube experiments intersecting the glass transition at high longitudinal strain rates in magma in nature.

Obviously the development of these experimental systems for use with higher viscosity materials will improve confidence in the applicability of the results they generate to natural magmas. Efforts in this direction include the addition of polymer solution to the H_2O-CO_2 system (Zhang et al., 1997; Zhang, 1998), by the substitution of gum rosin + acetone (Phillips et al., 1994) and by the very recent foaming and fragmentation experiments of Martel et al. (1998) where magma samples are subjected to decompression foaming followed by fragmentation.

The possibility has been raised by Klug and Cashman (1996) that a heterogeneous distribution of vesicularity within expanding rhyolitic magma might generate the opportunity for significant bubble coalescence at relatively low total vesicularity values via the mechanism of thin film instability. They postulate that pumice represents the highly permeable structures resulting from this coalescence that can survive the fragmentation event which produces volcanic ash layers in which the pumice blocks sit (see however below). Certainly this notion is consistent with the probability that strain localisation will lead to heterogeneities in multiphase magmas provided the strain and time available are large enough. Klug et al. (1998) have recently documented higher porosities and lower number densities in Plinian products of siliceous eruptions in comparison with pyroclastic deposits and infer that higher eruption rates in the case of the pyroclastic events have reduced the time available for bubble growth and coalescence.

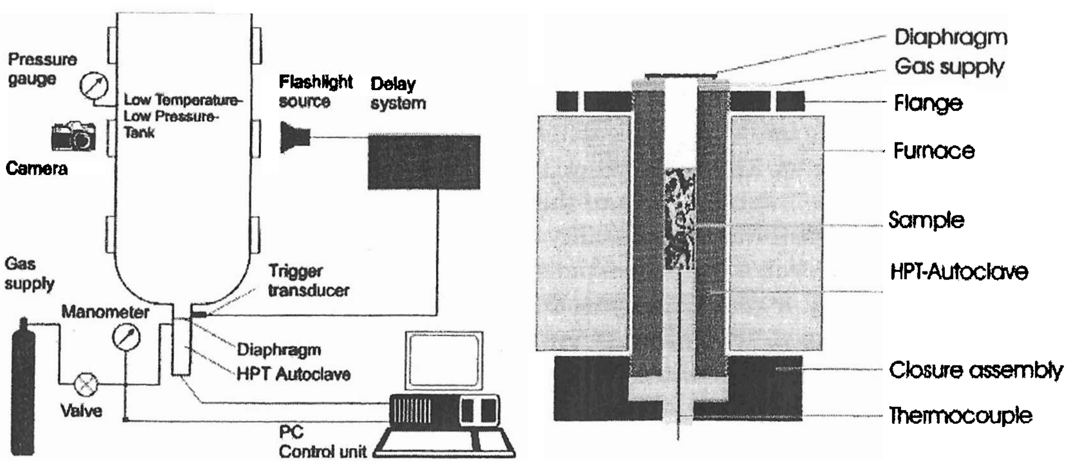

Fig. 8. The fragmentation bomb. A device used for the experimental fragmentation of high temperature, high pressure magma following rapid decompression. The device is composed of a high pressure and temperature section separated from a low pressure and temperature section by a rupture membrane (for a detailed description of the device see Alidibirov and Dingwell, 1996a,b). This figure is reproduced from Alidibirov and Dingwell (1996b).

5. BRITTLE FAILURE

The criteria for the brittle failure of magma are relatively easy to define. The applied stress must be sufficient to 1) push the strain response of the magma into the glass transition region and 2) to overcome the strength of the magma. A new experimental program to investigate the response of magma samples to rapid decompression has been initiated with the development of a so-called "fragmentation bomb" by Alidibirov and Dingwell (1996a,b, 1997a; Fig. 8). This device subjects samples of magma at temperatures up to 900°C and pressures up to 200 bar to rapid decompression resulting from the rupture of a membrane due to a pressure differential. The sample is expelled as a result of the rapid decompression from the high-pressure, high-temperature chamber (HPT) into the low-pressure, low-temperature tank (LPT). The expelled materials can be observed in flight and are completely recovered for analysis after the experiment.

Studies of the response of Mt. St. Helens' dacite under decompression rates in the range of 10^{10} Pa/s indicate brittle fragmentation of the magma as a result at temperatures as high as 950°C. The surfaces of recovered grains, investigated with SEM and optical methods, clearly reveal fragmental geometries indicative of brittle fracture propagation through the sample at high temperatures to cause its failure and fragmentation.

The quantitative description of the grain size distributions of pyroclastics experimentally generated by fragmentation of Mt. St. Helen's dacite have been presented by Alidibirov et al. (1998b). Fragmented samples collected from the fragmentation apparatus were subjected to sieve analysis. The grain size data, in terms of φ units ($\varphi = -\log_2 d$, where d is the diameter in mm) were expressed in terms of the statistics, median diameter ($Md_\varphi = \varphi_{50}$), mean diameter ($M_\varphi = (\varphi_{16}+\varphi_{84})/2$) and sorting coefficient ($\sigma_\varphi = (\varphi_{84}-\varphi_{16})/2$). Typical sieve grain size data are presented in Fig. 9. Two important trends observed in grain size data were: firstly, the mean and median grain sizes increased with fragmentation temperature and secondly, the median diameter decreased with increasing fragmentation pressure.

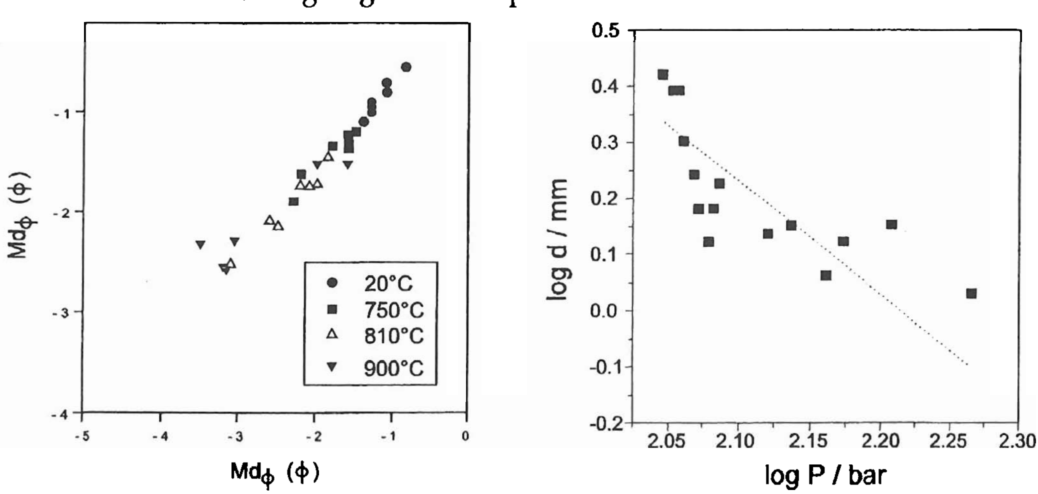

Fig. 9. The temperature- and pressure-dependence of particle characteristics following their generation by experimental decompressive fragmentation. (a) Median versus mean diameter for varying temperature. (b) Median diameter versus initial pressure differential. Reproduced from Alidibirov and Dingwell (1998a).

These experiments clearly demonstrate that magma at temperatures typical of the explosive eruption of dacitic to rhyolitic systems can fragment brittlely in response to rapid decompression and that this mechanism may be responsible for primary fragmentation in explosive volcanism. An important constraint on such a scenario is the decompression rate. The decompression rate defines the volumetric strain rate in the magma and this can be related to the volume viscosity of the material to obtain a predicted onset of brittle behavior using the principles of viscoelasticity noted above. Using this estimation the strain rates likely to be operative in the fragmentation bomb experiments appear to be high enough to fall within a range of response that is predicted to be brittle.

The fragmentation bomb experiments fall at the high strain rate limit of events expected to be operational in explosive volcanism. Because the strain rates in nature are related to the driving stresses through only poorly understood rheological constraints for complex magmatic materials, the estimation of volume strain rates in explosive eruptions is not trivial (see below).

A clear further experimental challenge for the future is the combination of the vesiculation processes observed in the accelerating flow (shock tube) experiments and the rapid decompression fragmentation of the fragmentation bomb experiments. These experiments, carried out on vesicular rhyolitic magmas with higher water contents, will bring us very close to answering the questions surrounding the dominant mechanisms of primary fragmentation in explosive volcanism (Martel et al., 1998).

Some mechanisms for the brittle fragmentation of magma in response to rapid decompression have been discussed recently by Alidibirov and Dingwell (1997b). They describe three possibilities: fragmentation by 1) propagation of an unloading wave, 2) propagation of a fragmentation wave and 3) fragmentation by rapid filtration flow. The propagation of an unloading wave refers to the pressure release wave propagating through a condensed magma, with or without vesicles. If the unloading wave involves stresses greater than the tensile strength of the magma and the wave propagates faster than the magma can viscously respond, then the magma will be fragmented by the tensile stresses. The propagation of a fragmentation wave refers to the case where the rapid decompression of the sample acts upon pressurized gas vesicles (closed porosity) within the magma. In this case the pressure differential between the pressurized vesicles near the dynamic magma surface and the ambient pressure behind the unloading wave may be large enough to disrupt the magma by individual shattering out of vesicles, one after the other. The third mechanism, that of rapid filtration flow, refers to the case of decompression of magma with an open porosity, either original (pre-decompression) or decompression-induced. In this case the rapid flow of decompressing gas through the tortuous open porosity generates considerable frictional forces on the surfaces of the channels, contributing to the fragmentation of the magma. Any combination of the three mechanisms is possible provided the magma contains the textural elements (open and/or closed porosity) or develops them during the fragmentation event. Recent mechanistic investigation of the fragmentation process by Alidibirov et al. (1998a) relies on dynamic pressure transducers to record the pressure at the top and bottom of the fragmenting magma sample as a function of time. The pressure - time signals from such a pair of transducers (Fig. 10) have been analysed by Alidibirov et al. (1998b) to

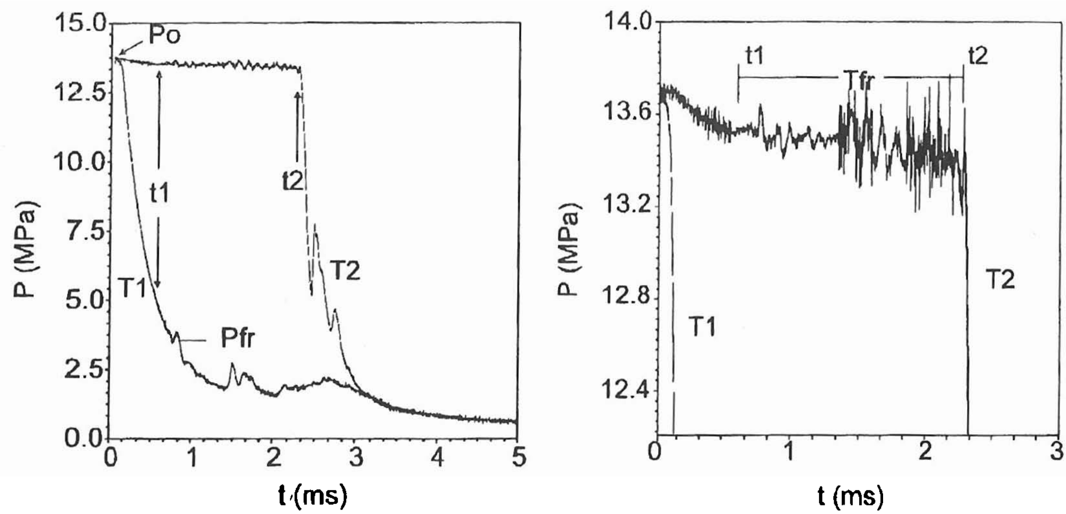

Fig. 10. The pressure histories for a fragmenting magma (20°C, 13.7 MPa) sample near the upper (T_1) and lower (T_2) ends. (a) The onset (t_1) and cessation (t_2) times of fragmentation can be used to estimate the fragmentation velocity. (b) An expanded view of the pressure signal on the T_2 sensor (sample bottom).

indicate that the fragmentation of Mt. St. Helen's dacite proceeds with a decompression rate corresponding to sample fragmentation of 5-23 MPa/ms or a fragmentation velocity of 5-23 m/s. Such rates appear to imply that fragmentation "wave" propagation can be quite slow indeed.

Due to the nature of structural and stress relaxation in melts it is very important for the evaluation of the possibility of brittle failure as a primary fragmentation mechanism to define the strain rates and therefore the decompression rates to be expected in the two-phase flow immediately prior to fragmentation. Certainly the strain rates obtained in shock tube experiments on two-phase accelerating flow are very high but what about the natural case? The ascent velocities of magma during eruption appear to be much too low to induce brittle phenomena during eruption generating primary fragmentation if the flow pressure equals the lithostatic pressure. Recent simulation results summarized by Sparks et al. (1994) indicate that this is not the case. Their Figs. 15 and 16 illustrate the point that there is an enormous increase in the decompression rate to be expected below the fragmentation threshold in a vesiculating magma due to the strong and non-linear increase in melt or matrix viscosity due to dehydration of the melt by this very vesiculation process (see Chapter 3). The decompression rates that can be read from these calculations are very near those for which a predicted onset of non-Newtonian viscosity occurs. The consequence could then be brittle failure.

6. POST-FRAGMENTATION EFFECTS

It has long been accepted that the inference of conditions attendant on silicic magmas at the point of magma fragmentation could be greatly aided by the investigation of natural materials generated in explosive eruptions. Consequently, a number of studies of silicic pumice and ash have documented their textural features (e.g., Heiken and Wohletz, 1985, 1991; Orsi et al., 1992; Dellino and LaVolpe, 1995, 1996). Application of the physical description of pumice and ash to the understanding of the fragmentation process however must consider the possibility of post-fragmentation effects in the exploded materials.

For the case of silicic volcanic ash, if we can assume that the fragmental surfaces have been generated in the primary fragmentation event then it seems relatively secure to use their textures to define the state of the magma at the fragmentation event and to correlate with experimental results and to try and infer the conditions of fragmentation.

For the case of pumice the situation might be more complicated because arguments have been made based on the textures of pumice and inferred magma properties at the time of fragmentation that postfragmentational expansion and/or collapse of vesicles can significantly influence pumice texture (Thomas et al., 1994; Gardner et al., 1996). The obvious example of a breadcrust bomb is an example often used to express this possibility.

The expansion of pumice clasts due to postfragmentation vesiculation has been theoretically analysed by Thomas et al. (1994). They suggest that the primary control on postfragmentational expansion is the melt viscosity and develop a dimensionless number to describe the effect of melt viscosity in terms of the relationship between viscosity, post-fragmentation pressure drop, and decompression rate. Thomas et al. (1994) conclude that for melt viscosities greater than 10^9 Pa s no post-fragmentation expansion is possible whereas for melt viscosities lower than 10^6 Pa s the magnitude of melt viscosity is irrelevant as the pumice equilibrates mechanically with the external pressure. In the range of 10^6 to 10^9 Pa s, the proposal is made that the vesicularity of pumice is influenced by viscosity and, of course, the postfragmentational pressure history. They suggest that temporal variations in pumice vesicularity recorded within successive units of a given eruptive sequence can be used to infer changes in eruption dynamics.

Recently, the analysis of pumice vesicularity has been developed a step further to include the consideration of post-fragmentational pumice (or foam) collapse kinetics together with the expansional analysis of Thomas et al. (1994) by Gardner et al. (1996). The essence of their work is described with the aid of Fig. 11 which illustrates the range of the vesicle to matrix volume ratio V_G/V_L for Plinian pumices of varying chemistry versus estimates of their degassed magma viscosities. The post-fragmentation expansion of the investigated pumices is calculated with the expression provided by Thomas et al. (1994) which predicts expansion up to 10^9 Pa s. Gardner et al. (1996) employ an expression for the collapse of permeable (post-fragmentation expanded) foams due to surface forces which predicts that magmas with viscosities of $< 10^5$ Pa s will suffer collapse and a corresponding reduction in vesicle volume after fragmentation and subsequent expansion. They further argue that the constancy of the lower limit of the volume ratio V_G/V_L at higher inferred viscosities is indicative of the vesicularity typical for the fragmentation event. Were a magma to have $> 10^5$ Pa s viscosity at the fragmentation event and have just achieved permeability due to expansion, then the prediction of the reasoning behind Fig. 11 is that this texture could be preserved in pumice. One apparent paradox remains. Fig. 11 is interpreted to indicate that fragmentation is occurring at

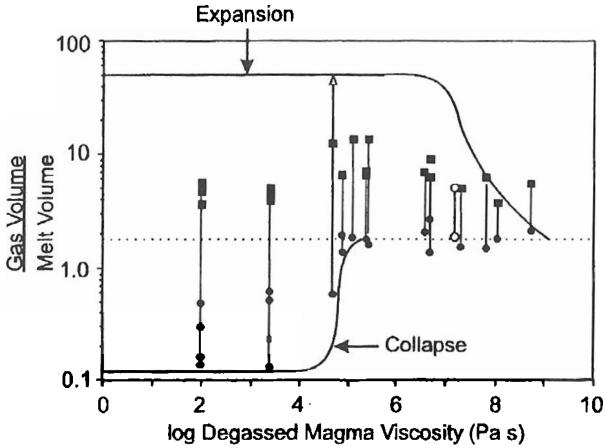

Fig 11. Vesicles/matrix volume ratio V_G/V_L versus estimated degassed viscosity (see text for discussion). Reproduced from Gardner et al. (1996).

≅ 64 vol.% of vesicles, yet Gardner et al. (1996) quantitatively propose the mechanism of expansion to much higher vesicularities without fragmentation of the pumice, as required by high vesicularity samples. The solution to this problem, they suggest, is the role of shearing in fragmentation. Pointing out that thin film rupture at thicknesses of ≅ 1μ requires expansion to > 95% vesicularity for the pumices considered, they propose that film thinning, caused by shearing, remains the fragmentation mechanism and that the inevitably heterogeneous distribution of shear stresses in the conduit thus lead to variable amounts of fragmentation locally. Thus Gardner et al. (1996) subscribe to the textural criterion for fragmentation where the accumulated magnitude of the shear strain, translated into bubble wall thicknesses, generates fragmentation, rather that the dynamic criterion of a critical strain rate, incorporated in the brittle failure mechanism (see above). A similar mechanism of shear-induced heterogeneous bubble coalescence has recently been appealed to by Stasiuk et al. (1996) to explain the distribution of vesicularity within an exposed rhyolitic vent via nonexplosive degassing due to transient permeability developed in the sheared vesicular magma.

7. NON-EXPLOSIVE DEGASSING

As discussed by Cashman and Mangan (1994) the apparent need for significant open system degassing in extrusive rhyolitic centers, based on water content versus D/H systematics has generated significant discussion and modelling of the feasibility of shallow level degassing of rhyolite (Eichelberger et al., 1986; Newman et al., 1988; Jaupart and Allégre, 1991; Taylor, 1991; Fink et al., 1992) and has prompted the experimental investigation of foam permeabilities and foam collapse by Westrich and Eichelberger (1994) and others (see above). The modelling performed to date on this subject has neccessarily involved simplifying assumptions about the relative permeabilities of the magma and the country rock of the vent to set one of them as rate-limiting (e.g., Jaupart and Allégre, 1991). The nature of shallow subvolcanic degassing is clearly an issue of vital importance to understanding the role of late stage degassing in controlling explosivity and fragmentation of rhyolites. The experimental and textural basis upon which our present knowledge of permeability and porosity at the fragmentation event relies must be better constrained by experiments and simulations in the next years. The permeability mechanisms devised by various workers to generate the efficient degassing envisaged by some (e.g., Eichelberger et al., 1986) must be tested in realtime experiments on the decompression, vesiculation, foaming and fragmentation or collapse of magmatic foams. The techniques for these tests are being rapidly developed.

Very recently, Herd and Pinkerton (1996) have raised the important issue that turbulence and shock in magmatic systems may disrupt bubble films before they thin to the textural instability criterion. In fact, the role of "external" or simply non-local stresses in disrupting foam to generate either coalescence or fragmentation is entirely uninvestigated. The best prospect for improvement of this deficit lies in strain impulse rheological studies or shock tube methods. External shock or other rapid stress changes, leading to unsteady eruption rates could be a central aspect of the feasibility of brittle failure mechanisms discussed above.

8. RELAXATION GEOSPEEDOMETRY OF VOLCANIC GLASS

Issues surrounding the possibility of postfragmentational strain in pumice and magma in general are intimately tied to the cooling history of the pyroclasts and in welded layers of volcanic glass in general. Partial quantification of the cooling history of volcanic glass is possible with the use of microscopic or macroscopic indicators of the quenched-in structure of the glassy matrix (Dingwell and Webb, 1990). This has been attempted to date using quenched-in water speciation (Zhang et al., 1996) and using the hysteresis of enthalpy in the glass transition interval (Wilding et al., 1995, 1996). Both methods yield information on the rate of cooling of the magma as it passes through the glass transition interval for the last time in the volcanic history of the sample. This relatively low temperature record of cooling history reveals a remarkable variability when applied to the range of volcanic facies available within a single volcanic center. The example of phonolitic glass cooling rates from the Teide Complex in Tenerife is provided in Fig. 12.

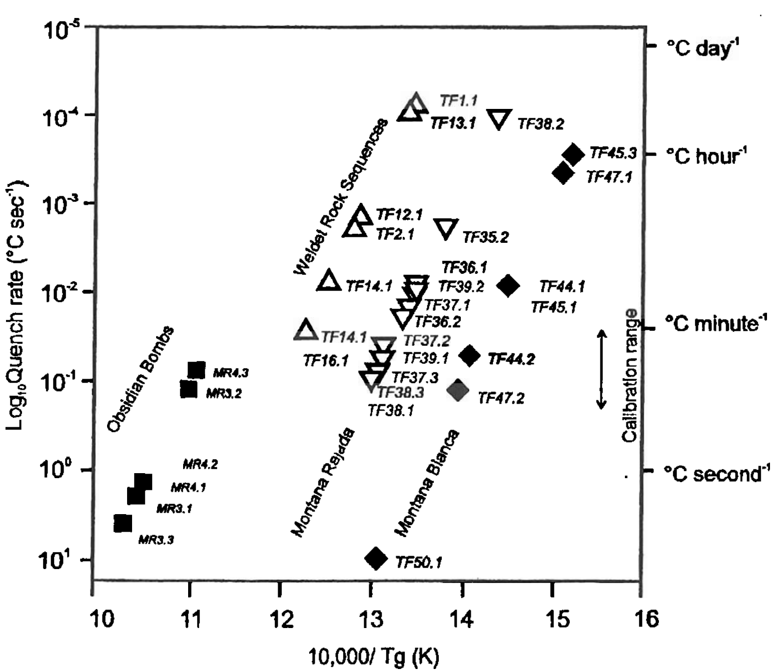

Fig. 12. Volcanic glass cooling rates determined using relaxational geospeedometry. Reproduced from Wilding et al. (1996).

Wilding et al. (1996) also observed enhanced cooling rates in the interior of breadcrust bombs and speculated that the thermal effects of degassing might play a role in generating faster cooling rates across the glass transition interval. The thermal analyses of Sahagian and Proussevitch (1996, see above) and of Mastin (1997) appear to support this notion and clearly such modelling should be applied to very low pressure scenarios to evaluate the possibility in general of whether very low pressure vesiculation can influence cooling and annealing histories in volcanic glasses.

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