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Magmatic Processes

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INTRODUCTION

The study of magmatic processes has developed from the simple fluid dynamical models of liquids characteristic of the early 1980s into an integrated investigation of the interaction between chemical and physical parameters of multicomponent and multiphase magmatic systems. Emerging from this shift in emphasis is the recognition of the importance of incorporating small-scale observations of melt geometry and local crystallization into large-scale models of magmatic segregation and storage. The models can merge only through understanding the time- and length-scales of melting and crystallization, which in turn control the dynamic feedback between the thermal and chemical evolution of a magma. In this article we attempt to incorporate all the available lines of evidence for the behavior of magmatic systems, including numerical and experimental models in addition to geophysical and geological observations. In this regard, an important publication in this quadrennium is Special Publication No. 1 of the Geochemical Society, a volume in honor of Hatten S. Yoder, Jr. entitled *Magmatic Processes: Physicochemical Principles*, which is a collection of papers on the behavior of magmas from their melting at the source region through their ascent, eruption and emplacement on the Earth's surface. We have attempted to compile a fairly complete bibliography in this area; we regret that space limitations do not permit discussion of all papers listed therein.

MELT GENERATION, SEGREGATION AND TRANSPORT

The mechanisms of melt generation and segregation in the mantle remain a central topic in the study of magmatic processes; a long-term goal is the integration of physical models of melt extraction with geochemical models of melt evolution and tectonic models of melt generation. The rheological limits to extraction will exert a major control on magma chemistry: a first-order question (reviewed in Turcotte, 1987) is the relative importance of (1) magma ascent as diapirs of melt with entrained crystals, (2) extraction and transport of melt from the source region through dikes and veins, or (3) large-scale upward transport of melt by pervasive percolation along grain boundaries. Diapiric transport of a crystal + liquid mixture would allow equilibrium conditions between the crystals and liquid up to the P-T regime of melt segregation; the critical melt percentage at which this can occur (often assumed to be ~50%, based on estimations of limits to melt eruptibility; e.g. Brophy and Marsh, 1986; Bergantz, 1990) is large relative to melt percentages estimated from trace elements in the generation of many basaltic magmas. While there is abundant observational evidence for dikes as the primary mode of magma transport at shallow crustal (brittle) levels (e.g. Spence and Turcotte, 1990), the importance of dike transport at depth is speculative (e.g. Maaloe, 1987; Takada, 1989) and will be controlled primarily by the viscosity contrast between the melt and the country rock (e.g. Emerman and Marrett, 1990) as well as the thermal structure of the conduit (Sleep, 1988). Although fractures generated in the presence of melt may supply a mechanism for rapid transport of material from depth (Spence and Turcotte, 1990), most melt migration in the mantle is believed to be controlled by permeable flow of melt along grain boundaries.

Deep melt migration is described by the differential equations governing porous flow of a melt within a viscously deforming matrix; melt migration is driven either by buoyancy contrast with the surrounding medium or migration under nonhydrostatic pressure gradients, and can be divided into categories of mechanics, thermodynamics and geochemistry (see review by Ribe, 1987). Experimental results now demonstrate that very small amounts (<1%) of texturally-equilibrated partial melt will migrate out of an olivine matrix (e.g. Cooper and Kohlstedt, 1986; Kohlstedt, 1990), with extraction facilitated by a gradient in differential stress (Cooper et el., 1989; Cooper, 1990; Riley and Kohlstedt, 1990; Riley et al., 1990). Studies of multiphase grain percolation in peridotites suggest that modal mineralogy of the solid phases will play an important role in the generation of a connected melt phase, and hence in the critical melt fraction needed for extraction, with melt connectivity for the same volume percent melt increasing with increasing olivine content (Toramaru and Fujii, 1986; Daines and Richter, 1988; Nakano and Fujii, 1989). Early one-dimensional models for melt extraction by matrix compaction (e.g. McKenzie, 1984) and porous media flow (e.g. Scott and Stevenson, 1986; Scott et al., 1986) have been replaced by two-dimensional models for (1) ridges (e.g. Phipps Morgan, 1987; Spiegelman and McKenzie, 1987; Buck and Su, 1989; Scott and Stevenson, 1989), where the driving force for flow is commonly assumed to be the non-hydrostatic pressure created by plate divergence, and (2) for mantle plumes (e.g. Ribe and Smooke, 1987), where thermal buoyancy provides the driving force. Fluid dynamical models for melt migration at ridge crests must account for melt generation over a broad area, and melt focussing into narrow axial magma storage reservoirs at ridge crests - focussing may occur due to corner flow (Spiegelman and McKenzie, 1987), mantle strain-induced anisotropic permeability (Phipps Morgan, 1987), or a laterally stably stratified residuum beneath newly-formed plates (Scott and Stevenson, 1989).

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Interpretive thermodynamic and geochemical models for mantle melt generation focus on determining the pressure, temperature and melt fraction at which the magma is chemically isolated from the crystalline residuum of the source region, and the interaction the melt then has with mantle and crustal material prior to shallow storage and eruption. The amount of chemical interaction will depend on the time- and length-scales of the transport process, which will be very different for fracture-dominated transport, diapiric upwelling and melt percolation (i.e. the physical processes of the system), thus emphasizing the close relationship between the chemical and physical processes of the extraction process (e.g. Nicolas, 1986). Richter (1986) presents a dynamical calculation for modal melting under conditions of local diffusive equilibrium, and subsequent melt segregation from a deformable matrix in order to model trace element fractionation as a function of the extraction process; he demonstrates that while end-member models for melting (perfect fractionation or perfect equilibrium) yield acceptable results, both assumptions may lead to overprediction of the incompatible element content of source regions. Holness and Richter (1989) show that observed variations in MORB chemical signatures with variations in spreading rate can be generated by incomplete extraction of melt in areas of more rapid spreading, and may thus reflect solely the physical process and not changes in the source region. Source heterogeneities are expected, however, from the chemical variability found in mid-ocean ridge basalts, as well as from the heterogeneous structures observed in peridotites (e.g. Reisberg and Zindler, 1986/1987), and would affect the dynamics of melt generation and segregation (e.g. Richter and Daly, 1989) and the trace element and isotopic signature of the source (e.g. Maaloe and Johnston, 1986). Transport of melt through the mantle may alter the chemical signature of the original melt. Chemical (trace element) exchange resulting from melt percolation through an isothermal matrix of mantle composition can be modeled as a chromatographic column (e.g. Navon and Stolper, 1987), with chemical equilibrium achieved as diffusive length scales are small and chemical exchange is maximized. In contrast, Singer et al. (1989) address possible effects of repeated diapiric upwelling in a single conduit system. Both of these models result in (1) large initial variability in chemical signature of the melt and (2) increased homogeneity with time due to both chemical and thermal equilibration of the melt and country rock. Finally, the process of infiltration may lead to patterns of geochemical self-organization (e.g. Ortoleva et al., 1987a,b) - reaction between basaltic magma and ultramafic rock may be important in subduction-related environments, and may be responsible for generating the chemical characteristics of calc-alkaline series magmas (Keleman, 1990; Keleman et al., 1990).

The existence and location of magma storage reservoirs is the result of the density distribution of the crust relative to the magma density (Gudmundsson, 1986a; Ryan, 1987; Glazner and Ussler, 1988). The location of magma storage as dictated by the local tectonic environment will control the shallow evolution of a magmatic system (e.g. Singer and Myers, 1990); additionally, there probably exists a balance between intrusion rate, chamber shape, tensile strength distribution around the chamber, and eruption frequency (Gudmundsson, 1986b, 1988). The use of geophysical observations to constrain magma chamber size and morphology is difficult, as known volcanic centers are by definition sites of prolonged thermal perturbations that result in a reduction of the contrast in physical properties between the magma and the surrounding country rock with its accompanying geothermal system. Potential field data, seismic tomography and the regional seismicity have been used to infer the presence of magma chambers (e.g. Iyer, 1988; Iyer et al. 1990; Rundle and Hill, 1988). Magma bodies under the Cascades volcanoes appear small, and Iyer et al. (1990) conclude that the subvolcanic regime in subduction-related volcanoes is the site of repeated small intrusions, such as the model emerging for the Katmai system from the work of Hildreth (1987). Geophysical data for the existence of magma beneath the caldera at Long Valley is equivocal: based on teleseismic tomography, Iyer et al. (1990) suggest that a magma body of 1000-1500 km^3 may exist under the resurgent dome; however, using higher resolution tomography from local earthquake data, Kissling (1988) prefers a model where the velocity contrast is the result of a plexus of small batches of partially molten material, a model consistent with the geochemical arguments of Sampson and Cameron (1987) for small magma batches in the Inyo system. A similar picture of small melt storage regions is emerging from seismic tomography studies of active spreading centers (e.g. Toomey et al., 1990). Results of tomographic inversion commonly show anomalously *high* velocities directly below volcanic centers, interpreted as solidified magmatic intrusions, underlain by broader low-velocity regions of partial melt (e.g. Foulger and Toomey, 1989; Toomey and Foulger, 1989). While a comprehensive picture of rapid melt removal and short residence times in crustal magma chambers is consistent with much of the recent observational and experimental data, such a model challenges traditional representations of magma chambers as large, long-lived vats of differentiating magmas.

Transport of magma from these storage reservoirs to the surface is commonly achieved by dike propagation - in the past four years there has been extensive work on the effect of this transport on the resulting eruptive products, and, conversely, on the information that can be obtained about transport conditions through a study of these products, as covered in the publication *Mafic Dyke Swarms* (Halls and Fahrig, 1987). In this volume, the mechanics of dike emplacement are reviewed by Turcotte (1987), while Pollard (1987) illustrates techniques to combine theoretical treatments of fracture mechanics with careful field measurements necessary to a comprehensive understanding of the energetics of dike formation.

New thermal models for dike cooling demonstrate that the duration of cooling will depend both on the distribution of latent heat across the crystallization interval (e.g. Delaney, 1987; 1988; Ghiorso, 1990) and the duration of flow (e.g. Bruce and Huppert, 1989; Fabre et al., 1989), as complex patterns of solidification and meltback may result from prolonged flow duration and released latent heat of crystallization (Huppert and Sparks, 1989). Such relationships could be expected to affect not only resultant magma chemistry and rock textures, but may also control eruption dynamics. While these models provide important new guidelines for understanding field relationships, they remain incomplete due to our inadequate knowledge of the complex relationship between heat transfer, crystallization and magma rheology coupled with possible feedback to the system from metamorphic reactions and melting of the country rock. For example, Kille et al. (1986) show that the country rock composition (in particular, the melting temperature of different country rock strata) controlled macroscopic and microscopic (grain size) features of dikes in southwest Mull.

The transport mechanism may exert a substantial control on the chemistry of the erupted products. Magma mixing in conduits

has been explored by Koyaguchi (1987) and Koyaguchi and Blake (1989), who show that the eruption of an initially stably stratified system can lead to mixing in the conduit. Experiments modeling flow through a conduit with a viscous outer fluid and low viscosity inner fluid (e.g. Blake and Campbell, 1986; Freundt and Tait, 1986) show entrainment across the interface boundary, with the degree of mixing increasing with increasing Reynolds number, fluid velocity and distance. Such a mechanism may be operative under conditions either of lateral stratification within the dike (e.g. Blake and Fink, 1987) or in small magma batches (e.g. Koyaguchi and Blake, 1989) and will result in mixed magma products. In contrast, Carrigan and Eichelberger (1990) propose that two magmas flowing in the same conduit may unmix to yield low viscosity (basaltic) layers near the margins and a plug-like flow of the silicic component in the center; they suggest that viscosity segregation during viscous flow may be responsible for zonation in eruptive products irrespective of the pre-eruption configuration. Evidence for physical interaction during the concurrent eruption of material drawn from separate magma reservoirs exists in the compositional range of magmas erupted during the 1912 eruption in the Valley of Ten Thousand Smokes, where contemporaneous eruption of rhyolitic, dacitic and andesitic magmas has been explained by tapping of separate though related magma bodies under Mts. Trident and Katmai (Wallmann et al., 1990). Conduit *geometry* may also place important constraints on flow behavior in eruption fissures (Ida and Kumazawa, 1986; Giberti and Wilson, 1990).

Field studies on dike rocks continue to provide important constraints on dike flow dynamics. Flow directions can be obtained by macroscopic field observations (e.g. Baer and Reches, 1987; Smith, 1987), petrologic data relating dike rock chemistry and mineralogy to possible source areas (e.g. Macdonald et al., 1988), and through AMS fabric studies (e.g. Knight and Walker, 1988). Such studies indicate that flow in dikes is predominantly lateral, a conclusion consistent with observations of inferred shallow dike emplacement and resulting fissure eruptions in active volcanoes (e.g. Gudmundsson, 1987; Ryan, 1987; Sigurdsson, 1987; Wilson and Head, 1988; McGuire and Pullen, 1989). Both remote stress and magmatic pressure control this shallow lateral propagation (Rubin and Pollard, 1987) - the depth of transition from predominantly vertical to predominantly lateral flow remains a topic of investigation (e.g. Emerman and Marrett, 1990). Dikes may also act as "planar storage reservoirs" (Wilson and Head, 1988) in active basaltic systems, and may thus provide the location for shallow differentiation and mixing in subsequent eruptions (e.g. Helz, 1987a; Mangan, 1990).

The drilling program at Inyo Craters, eastern California, has provided a three-dimensional view of a shallow silicic eruptive system (Eichelberger et al., 1988; Kerr, 1988), allowing evaluation of the relationship of dike geometry to intrusion mechanism (e.g. Mastin and Pollard, 1988; Reches and Fink, 1988) as well as determination of magmatic degassing and crystallization at different stratigraphic levels and cooling regimes (e.g. Manley and Fink, 1987; Westrich et al., 1988; Swanson et al., 1989). The geometry of different magma types in the feeder dike, the conduit and on the surface suggests a complex relationship between the pre-eruption location of high- and low-Si end member magmas, changing eruption rates and the dynamics of magma mixing during transport. In particular, observed mafic-to-silicic (rim-to-center) chemical zoning in the conduit has been explained by Vogel et al. (1987; 1989) using relationships derived by Blake and Fink (1987) for eruption from a stratified dike; they calculate draw-up distances of 70-400 m for plausible variations in eruption rates. An alternative explanation for the observed associations between two magma types is viscosity-controlled segregation of magma due to flow in a conduit regardless of the pre-eruptive geometry of the source magmas (Carrigan and Eichelberger, 1990).

MAGMA CHAMBERS

The style and vigor of magma chamber processes have been the subject of some controversy; this is a measure of both the difficulty of formulating *testable* hypotheses for real magma chambers and of discriminating among competing processes as represented in the rock record. The last four years have seen a major shift in the modeling of magmatic processes from a study of simple convecting fluids to the more realistic treatment of magmas as multi-component systems, with physical properties (e.g. viscosity, density) that vary stongly with time and location, and crystallization occuring at both the chamber margins and in the interior. It is clear that parameterizations of fluid dynamics for thermodynamically simple, low viscosity systems must be applied cautiously, indeed if at all, as crystallization physically partitions the magma body and generates buoyancy in complex ways. Many of the generic aspects of magma chambers are discussed by Marsh (1989a) and Morse (1988). Crystal settling, magma mixing and convection continue to be the dominant physical processes thought to yield petrologic diversity, although physical models of compaction and buoyancy-driven extraction in partially-solidified zones at chamber walls (e.g. Shirley, 1986; 1987; Spera et al., 1989) have led to the examination of the geochemical implications of such a process (e.g. Barnes, 1986; Morse, 1986a; Langmuir, 1989).

Crucial to all the models presented below are well-constrained measurements of the physical properties of different magmas throughout the crystallization interval. New density measurements and compilations for silicate melts are presented in Lange and Carmichael (1987); a review of volumetric properties of silicate liquids (including density, compressibility and thermal expansion) can be found in Lange and Carmichael (1990). Magma viscosity varies greatly with changes in volatile content and included phases (crystals or bubbles), and is a difficult property to measure experimentally, thus making empirical correlations prone to error. Inclusion of temperature-dependent viscosity in fluid dynamical models may greatly alter predictions for magmatic convection, however, (e.g. Oldenburg and Spera, 1990a,b), and for eruption styles (e.g. Blake and Ivey, 1986a,b; Bonatti and Harrison, 1988; Wolfe et al., 1990). Basalts show a decrease in viscosity from 2.5 Pa s at 0.8 GPa to 0.2 Pa s at 3.5 GPa along the anhydrous solidus of peridotite (Kushiro, 1986). Non-Newtonian behavior of subliquidus basalts under conditions of shear is the result of solid-solid interactions; observed viscosities are in good agreement with those calculated from the Einstein-Roscoe equation for a serial size distribution of solids (Ryerson et al., 1988). Measurements on multiphase magmatic suspensions in rhyolitic magmas also show non-Newtonian behavior that can be modeled as a power law fluid; the power law exponent increases with increasing temperature largely due to expansion and deformation of vapor bubbles, and a normal stress coefficient is extracted for rhyolitic magma (Spera et al., 1988). Field measurements on changes in magma rheology due to crystallization during flow (e.g. Lipman and Banks, 1987; Rowland and Walker, 1988) also help to constrain magma viscosities.

One of the notable features of magma bodies is that crystallization can occur simultaneously at the margins and in the interior. This leads to a partitioning of the body into a mushy zone near the contact where a rigid or plastic region of crystals exists with an interstitial melt phase, and a slurry in the interior where crystals reside in an expanse of melt. The key element in the mush-slurry distinction is that the transition from a rigid or plastic mush to a convectable, mobile slurry is a rheological one. The duration of convection in a solidifying magma was explored in the work of Brandeis and Marsh (1989, 1990) and Marsh (1989b), where they introduced the concept of the convective liquidus based on experiments with liquid paraffins. The convective liquidus is defined as the isotherm associated with the termination of convection, which may be very nearly that of the actual liquidus. The convective liquidus is difficult to determine *a priori* and will be a function of the ensemble crystal-melt properties. One implication of their work is that temperature differences and length scales that are available to drive convection in multi-component systems do not necessarily follow from the the initial temperature contrast or chamber geometry, thus emphasizing the need for integrating realistic thermodynamic and kinetic parameters across the crystallization interval. Most of the temperature and concentration differences will exist across this more rigid mushy zone (Kerr et al., 1989), hence all the fluid processes will be explicitly coupled to the transient history of crystallization. Experiments by Kerr et al. (1989) with initially superheated aqueous solutions demonstrated a very different convective history from the paraffin experiments described above, with vigorous compositional convection (i.e. convection driven by buoyancy of low density melts generated in the process of crystallization) that kept the interior well stirred; they attribute the vigorous convection to temperature differences of undercooling. In contrast, experiments by Tait and Jaupart (1989) on higher viscosity aqueous solutions indicate that compositional convection is inhibited for parameter ranges typical of many magmas and hence the paraffin experiments may provide the more geologic relevant end member. Both the paraffin and the aqueous crystallization experiments indicate that the progress of solidification is largely consistent with predictions of a conductive cooling model.

The efficacy of crystal settling as a means of driving fractionation has been the subject of laboratory and numerical experiments. Experiments with small volumes of crystals demonstrate that particles can settle despite the action of convection (Martin and Nokes 1988; 1989), although the calculations and experiments of Weinstein et al. (1989) reveal that retention of crystals in a convecting fluid (under laminar flow conditions) is sensitive to the site of nucleation. Koyaguchi et al. (1990) show, however, that when concentrations of crystals exceed ~3 vol. %, they can initially stabilize a convecting fluid that subsequently undergoes a rapid overturn or destabilization as settling proceeds. Elements that are missing from these studies of crystal settling which are present in real magmas include: (1) the convective regime will be strongly influenced by processes at the margin (in the mushy zone), and (2) the tendency of real crystals to undergo resorption and/or growth while being swept around in the chamber. Experiments with aqueous solutions by Martin (1990) demonstrate that the partitioning of crystallization between walls and interior is sensitive to the rate of heat loss and liquid viscosity. The petrological implications for partitioning of the magma chamber have been explored in Marsh (1988a,b). For example, the settling of crystals in plumes that originate in the mushy zone at the margins may allow sorting during descent, while growth, resorption and the ability of crystals to "escape" the moving crystallization front will control both the rate of crystal accumulation at the chamber floor (and hence the rate of growth of the lower mush zone) and the rate at which plume formation occurs at the chamber roof.

Stratification of magma chambers by generation and flow of evolved melt during crystallization is undergoing reappraisal based on experimental and computer models of crystallization from viscous fluids. Numerical models of double-diffusive convection provide parametric expressions for convection in the absence of crystallization (Clark et al. 1987; Spera et al. 1986). Experiments with crystallizing aqueous solutions show various forms of stratification, depending on whether crystallization is taking place on the floor, roof or sloping walls (Huppert et al. 1986b; 1987; Leitch 1987, 1990; Martin and Campbell, 1988; Martin, 1990). Viscous fluids, however, yield substantially different fluid structures and macrosegregation than less viscous salt solutions, as demonstrated by Tait and Jaupart (1989) and Martin (1990). These results suggest that a critical re-evaluation of models based on experimental work on crystallization and double-diffusive convection in aqueous systems is necessary. Mixture theory has been used to model solidification and double-diffusive convection during crystallization in constant property binary systems (Bennon and Incropera, 1987a,b; Beckerman and Viskanta, 1988; Thompson and Szekeley, 1989). These formalisms were extended by Oldenburg and Spera (1990a,b) to examine solidification in the nearly ideal eutectic system of diopside-anorthite. Although there are difficulties in prescribing constitutive expressions and permeability functions for the flow of evolved melt in the mushy zone (see Daines and Richter, 1988) experimental work indicates that a Carman-Kozeny relationship, which relates porosity to permeability, may be appropriate (Bergantz, 1990). Computer intensive modeling of the type done by Oldenburg and Spera (1990a,b) may prove to be the best way to combine the strongly varying transport properties of magmas with realistic crystallization and fluid flow scenarios, but they must be coupled with accurate thermodynamic and kinetic descriptions of silicate melts.

Finally, although magma mixing remains an often invoked means of producing petrologic diversity, the specific driving forces for mixing remain poorly understood. The fundamental kinetics of mixing of two magmas has been considered by Sparks and Marshall (1986) and Frost and Mahood (1987); they show that subliquidus magmas may behave immiscibly if there exists a large viscosity contrast between the mixing populations. These results imply that homogeneous mixing of two magmas is probable only for magmas of similar compositions, or for large amounts of a mafic (hot) end member (Frost and Mahood, 1987). Numerical experiments of mixing of a silicic magma overlying basalt (Oldenburg et al., 1989) include the effects of variable viscosity and double-diffusion. Complex, time dependent modes of convection appear, with significant lateral heterogeneity that may yield a vertically stratified pyroclastic sequence upon eruption. Spera et al. (1990) demonstrate that many of the features ascribed to magma chamber processes, such as compositional gaps and mixed magma horizons, can be due to the eruptive processes itself. In this regard, the possible origins of enclaves in plutons continues to spark debate - a special section on granites was published in J. Geophys. Res., v. 95, no. B11, 1990. Interpretation of inclusions as refractory source material (e.g. Chen et al., 1990) is not as popular as the interpretation of

inclusions as fragmental accumulations of minerals crystallized from previously mixed magmas (e.g. Vernon, 1990; Dodge and Kistler, 1990; Pin et al., 1990), mixing of magmas (e.g. Stimac et al., 1990; Larsen and Smith, 1990; Davidson et al., 1990), or a combination of pre-entrapment hybridization and post-entrapment modification (Linneman and Myers, 1990). Evidence for the loss or gain of interstitial liquid makes interpretation of enclaves as direct representations of original mafic melts suspect (e.g. Linneman and Myers, 1990; Christiansen and Venchiarutti, 1990). Ayrton (1988) provides an intriguing synthesis of the many enclave enigmas. He calls for mixing of mafic magmas at the margins of a ballooning pluton, in the style of a failed ring-dyke. Dorais et al. (1990) and Eberz and Nicholls (1988) also conclude that many enclaves may originate from the injection of a basaltic magma into a more silicic magma chamber.

ACTIVE VOLCANOES

Investigations of volcanoes provide a realtime database for understanding all aspects of active magmatic systems. Since the Puu Oo eruption of Kilauea Volcano, Hawaii, has dominated this quadrenium as the most active volcano in the United States, we review here, as example, investigations of recent eruptions of Kilauea and Mauna Loa volcanoes that have a direct bearing on understanding of processes of magmatic differentiation in basaltic systems. Summaries of available knowledge on Hawaiian volcanoes have been published in USGS Professional Paper 1350, *Volcanism in Hawaii* (edited by Decker et al., 1987), and USGS Professional Paper 1463, *The Puu Oo eruption of Kilauea Volcano, Hawaii: episodes 1 through 20, January 3, 1983, through June 8, 1984* (1988).

A fundamental question in magmatic differentiation is the relative importance of fractionation and mixing in creating the range of lava compositions observed. Recent studies of eruptive products from the past 40 years of activity at Kilauea suggest that different processes dominate under different conditions. Olivine-controlled fractionation is common for most of the less-differentiated magmas, with examples being the 1967-1968 Halemaumau and Hiiaka eruptions (Nicholls and Stout, 1988) and the Puu Oo eruption after a single vent was established (Wolfe et al., 1987; Garcia et al., 1987; Wolfe and Garcia, 1988). In addition, small amounts of olivine fractionation occurred during short term (8-65 days) repose periods between vent eruptions at Puu Oo (e.g. Wolfe et al., 1987). Differentiation involving plagioclase and clinopyroxene in addition to olivine occurs in isolated melt pockets along the rift zone, as seen in the initial material erupted from the Puu Oo eruption (Wolfe et al., 1987; Garcia et al, 1989), and the 1955 East Rift Zone lavas (Ho and Garcia, 1988; Russell and Stanley, 1990). Both of these examples are consistent with magma residence in "planar magma storage reservoirs" (Wilson and Head, 1988) within rift zones; dimensions for the dike feeding the Puu Oo vent of 2.5 km in height, 1.6 km in length and 2.5-3 m in width have been determined from horizontal ground deformation measurements (Hoffmann et al., 1990). Mangan (1990) has estimated a residence time of approximately 10 years for the shallow differentiate erupted during the 1959 Kilauean eruption by measuring sizes of olivine phenocrysts in the erupted products. Injection of more primitive magma into a differentiated pocket of melt may result in mixing of magmas (Helz, 1987a; Garcia et al., 1989) or may act as a hydraulic plunger causing eruption of melt without mixing (Ho and Garcia, 1988) - physical controls on these two styles of dike intrusion are not well understood.

The deeper part of Kilauea's conduit system is concentrically zoned, and inferred to contain a high permeability core (Ryan, 1988); a deep melt body may also persist in the East Rift Zone (Delaney et al., 1990). Studies of Kilauea and Mauna Loa volcanoes indicate that a balance exists between magma supply rates, differentiation rates and eruptions. For example, Mauna Loa volcano commonly produces homogeneous lavas, as typified by the constant composition and eruption temperature of the 22-day 1984 eruption (Rhodes, 1988). Bulk compositions are located at the reaction point involving olivine, clinopyroxene, plagioclase and pigeonite, suggesting that the composition is buffered by a fairly continuous magma supply with further differentiation (away from olivine control) possible only when pockets of magma are isolated from the primary supply (in this context, Lockwood and Lipman, 1987, note that the eruption rate and chemistry of Mauna Loa lavas changed significantly after the 1868 M7.5 earthquake). In contrast, an increase in average supply rate relative to crystallization/settling results in increasingly mafic compositions (Rhodes, 1988). While it has been noted that long term increases in magma supply rate may lead to quasi-steady-state eruptions (e.g. Tilling et al., 1987a; Casadevall et al., 1987), the cause of such variations relative to melt generation and transport from the source area are unknown.

Hawaiian lava lakes provide a natural laboratory for the study of the physical and chemical changes in a cooling basalt. The best studied of these lava lakes is Kilauea Iki, reviewed by Helz (1987b). The complexity of the processes observed on this small scale are sobering when compared with the simplicity of available models of magmatic cooling and differentiation. These processes include (1) fractionation of basalt liquid to small amounts of rhyolite, (2) control of the effective modal proportion of olivine on the succession of liquid compositions due to the rapid re-equilibration of olivine with the liquid, (3) diapiric low-ρ melt upwelling as an efficient mechanism of chemical and heat transfer, occurring without a textural signature due to low crystal contents (Helz et al., 1989), (4) gravitative settling of olivine and rise of vesicles and vuggy olivine-rich bodies, (5) emplacement of segregation veins of ferrobasalt composition in partially-solidified crust, a mechanism for both heat and mass transfer. Hawaiian lava lakes have also provided an ideal natural crystallization experiment. Crystallization rates for plagioclase and ilmenite from Makaopuhi lava lake have been estimated by the application of crystal size distribution (Marsh, 1988b) analysis to drill core samples (Cashman and Marsh, 1988). Growth rates of $5\text{-}10\times10^{-11}$ cm/s (plagioclase) and $3\text{-}5\times10^{-10}$ cm/s (ilmenite) and nucleation rates of $1\text{-}30\times10^{-3}/\text{cm}^3\text{s}$ indicate that even under conditions of rapid surface cooling, lava at depths greater ~10m crystallized at slow rates, which in turn suggest very small degrees of undercooling.

PETROLOGIC CONSTRAINTS

Magmatic differentiation is a complex process involving varying degrees of crystal fractionation, mixing and assimilation. While crystal fractionation still explains much of the variation seen in the products of single eruptive centers (e.g. Cox and Mitchell, 1988), minor assimilation and compositional convection (the ascent of the light melt fraction created by partial solidification) are usually needed to explain details of differentiation trends (e.g. McBirney et al., 1987; Reagan et al., 1987; Grove et al., 1988; Bacon and Druitt, 1988; Druitt and Bacon, 1989; Brophy, 1989), although the direct coupling of crystallization and assimilation has been questioned (Grove et al.,

1988). There is abundant evidence for pre-eruption stratification in all types of magmatic systems, and mixing on all scales (e.g. Koyaguchi, 1986a,b; Newman et al., 1986; Brophy, 1987; Camp et al., 1987; Pyle et al., 1988; Bloomfield and Arculus, 1989) as well as mingling of different magma types on eruption (e.g. Bacon, 1986; Sparks, 1988; McGarvie et al., 1990; Robin et al., 1990). As alluded to above in the sections on magma chambers and dike transport, fundamental questions remain as to the physical means by which these relationships arise. "Magma mixing" is used to describe processes from the scale of macroscopic comingling to complete homogenization of the liquid phase; suggested mechanisms of mixing include forceful injection (e.g. Brown and Becker, 1986), convective overturn (e.g. Green, 1988; Nixon, 1988a,b), compositional convection (e.g. Bacon and Druitt, 1988), volatile concentration/exsolution (Tait et al., 1989) and mixing at a double-diffusive boundary layer (e.g. Koyaguchi, 1986b). Recognition of mixed magmas is commonly on the basis of disequilibrium phenocrysts (e.g. O'Brien et al., 1988), which Vogel et al. (1989) suggest are *a priori* evidence for mixing. However, detailed studies of phenocryst stratigraphy show that zoning is ubiquitous and complex, and generally does not concur with a single mixing event (e.g. Pearce et al., 1987). Additionally, rates of phenocryst resorption or rim growth will be strongly controlled by the phase equilibria of the mixed assemblages (Ussler and Glazner, 1989), and the degree of both thermal and chemical supercooling (e.g. Bacon, 1986; Sparks and Marshall, 1986; Geist et al., 1988). Controls on the effect of these variables are few, and recent experimental evidence suggests that isothermal decompression (an expected result of eruption) can create similar effects (e.g. Helz, 1987a).

In an editorial introduction to a debate on the origin of granitoids, W. Casey (EOS, v. 79, no. 9, Feb. 27, 1990) stated that "...anatexis is perhaps the most effective process in inducing chemical reorganization within existing crust," although the characteristic timescale for this process may be tens of millions of years (Zen, 1988). The "ponding" of mafic magma in the deep crust (including young mafic crust that is isotopically indistinguishable from mantle) with subsequent partial melting and mixing appears to be a potentially important means of generating evolved melts. One geochemical paradigm for this process is the MASH (melting-assimilation-storage-homogenization) hypothesis of Hildreth and Moorbath (1988). Physical models for coupled solidification-melting yield melt fractions whose volume and compositions appear to be in agreement with some silicic systems (Bergantz, 1989a,b; Huppert and Sparks, 1988; Fountain et al., 1989). Field evidence for the transport mechanism may lie in the textures of granitic bodies (e.g. Hutton, 1988; Miller et al., 1988). More work remains to be done to elucidate the basic time and length scales of the interaction of basalt with the crust, and to fully understand the role of water in the generation of granites (e.g. Wickham, 1987; Whitney, 1988). Laboratory experiments that determine the volume fraction melt as a function of temperature (e.g. Beard and Lofgren, 1989; Conrad et al., 1988; Rutter and Wyllie, 1988) are needed. Measurements on the simple binary nepheline-sodium disilicate system suggest that the melting process may be decoupled from heating, probably due to high melt viscosities and sluggish reaction kinetics (Yoder, 1990).

Magma supply rates must ultimately control both the duration of crustal magmatic reservoirs and the periodicity of eruptive events (e.g. Mahood and Baker, 1986; Thy, 1987; Wetzel et al., 1989; Defant and Nielson, 1990). While rates of magma generation may be tectonically controlled (McKenzie and Bickle, 1988), rates of ascent are related to crustal structure (e.g. Newman et al., 1986; Baker, 1987; Grunder and Mahood, 1988), which may also influence recognized patterns of complex crustal reservoirs (e.g. Hildreth, 1987; Donnelly-Nolan, 1988; Nixon, 1988). Since large-volume ash-flow tuffs and layered basic intrusions continue to be popular models for magmatic systems, we briefly review the field and petrologic work on these systems that relate directly to elucidation of magmatic processes.

Large-volume ash-flow tuffs provide valuable information on the stratification and processes occurring in silicic magma reservoirs. Differentiation in most large ash-flow deposits is consistent with fractional crystallization (e.g. Grunder and Mahood, 1988; Halliday et al., 1989) with some degree of both assimilation and mixing (e.g. Musselwhite et al., 1989), and may result in the generation of large quantities of new crust (e.g. Riciputi and Johnson, 1990). The controversy (e.g. Grunder and Boden, 1987; Stormer et al., 1987) generated by the suggestion of Whitney and Stormer (1985) that the large-volume Fish Canyon Tuff equilibrated at 9 kb just prior to eruption has been resolved by the experimental work of Johnson and Rutherford (1989a,b), that shows the Fish Canyon Tuff hornblendes to have equilibrated at 2.4 kb. Chemical zonation in these systems may be reflected in the inverted sequences of deposits, in variations in quenched fiamme within the deposits, or in volatile gradations measured in phenocryst melt inclusions. Zonation styles include continuous zonation restored by fractional crystallization (e.g. Druitt and Bacon, 1988); continuous volatile zonation (Anderson et al., 1989); step function changes in bulk composition but continuous gradations of liquid compositions (Bacon and Druitt, 1988); abrupt step function zoning (Carey and Sigurdsson, 1987; Fridrich and Mahood, 1987; Boden, 1989; Schuraytz et al., 1989; Sigurdsson et al., 1990), sidewall crystallization (e.g. Sawka et al., 1990), isotopic zonation resulting from assimilation (Johnson, 1989), and chemical homogeneity (e.g. Dunbar et al., 1989; Self et al., 1988). Estimates of the time persistence of silicic systems also vary, from long term stable stratification of the precaldera (0.7 my; Halliday et al., 1989) and postcaldera (0.5 my; Johnson and Rutherford, 1989) lavas of the Long Valley system, to the lack of zonation observed over 50,000 yrs. in the Taupo system (Dunbar et al., 1989). Although Sparks et al. (1990) interpret the isotopic data of Halliday et al. (1989) to be the result of many small rapid melting events, rebuttals by Halliday (1990) and Mahood (1990) are in agreement with time estimates for periodicity seen in the Bandolier magmatic system of 0.27-0.37my (Styx et al., 1988; Spell et al., 1990) and with the compilation of Wolfe et al. (1990) suggesting that the time required for the development of significant zonation in large-volume silicic magmas is $>10^5$ yrs. Isotope systematics suggest, however, that phenocryst crystallization in many ash-flow tuffs must postdate assimilation and predate eruption by < 100 yrs. (Johnson, 1989).

Layered basic intrusions continue to pose intriguing questions about the development of large basic magmatic systems (see *Origins of Igneous Layering*, ed. I. Parsons, 1987), although growing evidence for significant postcumulus modification makes the direct interpretation of magmatic processes from these rocks more difficult (e.g. Boudreau, 1987; Hunter, 1987; Mathison, 1987; McBirney, 1987; McBirney and Russell, 1987; see review by Lesher and Baker, this issue). Problems associated with determining magmatic processes from layered intrusions are illustrated by the controversy provoked by the interpretation of the Skaergaard Intrusion presented by Hunter and Sparks (1987) and

countered in a series of Discussions and Replies (Brooks and Nielson, 1990; Hunter and Sparks, 1990; McBirney and Naslund, 1990; Morse, 1990). Fundamental questions raised in this interchange include (1) the relationship between plutonic and volcanic rocks, and by extension, their processes, especially if the eruption of magma can affect the evolution of the system as a whole (Morse, 1990); (2) determination of volume relationships, especially of the concurrently crystallizing roof, sidewall and floor regimes that are critical input for models of heat transfer due to convection (McBirney and Naslund, 1990; Hunter and Sparks, 1990); and (3) when the extent of roof melting is largely a function of the heat transfer mechanism envisioned and the evidence for large-scale assimilation is lacking due to the erosional level of the intrusion, how to reasonably choose among the models presented?

Many of the other unanswered questions regarding the physical evolution of layered basic intrusions revolve around establishing the balance between rates of crystallization (cooling) and rates of magma chamber replenishment (growth). Models of layered intrusion formation by a combination of fractionation and periodic replenishment are common (e.g. Naldrett et al., 1987; Robins et al., 1987; Wiebe, 1987a,b; Ballhaus and Glikson, 1989; Stewart and DePaolo, 1990) - periodic replenishment by mafic magma must, then, change not only the chemical structure but also the thermal structure of the magmatic system (Campbell and Turner, 1986; 1989; Huppert et al., 1986). Large amounts of new material must also increase the overall chamber size, and the relative rates of crystallization, replenishment, stoping, or engulfing of roof rocks will then affect patterns of crystallization and assimilation (e.g. Pedersen, 1986; Wilson and Engell-Soresen, 1986; Wiebe, 1988; Habekost and Wilson, 1989); the extent to which processes of crystallization and assimilation are decoupled remains a question (e.g. Campbell and Turner, 1987). Irvine (1987) reviews possible explanations for commonly observed sedimentary structures, which are in turn tied to the balance between magmatic cooling, crystallization and convection (e.g. Conrad and Naslund, 1989). Variations in settling rates and crystallization regimes may also affect original porosities of cumulus piles and resultant textures (e.g. Campbell, 1987). Finally, competing processes of nucleation and growth that control textures and ease of compositional convection are themselves controlled by convection (Morse et al., 1987; Donaldson and Hamilton, 1987), cooling rates (Morse and Allison, 1986; Bedard, 1987;), sidewall crystallization (Parsons and Brown, 1988), and changing concentration of volatiles (Parsons and Becker, 1987). These processes in turn have a profound influence on the texture and geochemistry of the resulting rocks (e.g. Langmuir, 1989).

CRYSTALLIZATION AND VESICULATION

As illustrated throughout this review, processes and rates of crystallization and dissolution control the dynamics of magmatic processes. A fundamental observation about rock textures is that with the exception of megacrysts and products of vapor-phase crystallization (e.g. pegmatites), variations in grain size throughout a dike, sill, lava flow sequence or intrusion are small (2-3 orders of magnitude), suggesting that crystallization times in these systems are similar (e.g. Cashman, 1990a). In this section we review the many experimental, theoretical and observational contributions of the past four years that represent progress toward understanding crystal nucleation, growth and dissolution in complex silicate systems. Recent reviews of the burgeoning engineering literature on the heat and mass transfer attendant with phase change can be found in Yao and Prusa (1989), and Viskanta (1988), and are collected in *Structure and Dynamics of Partially Solidified Systems* (ed. D.E. Loper). Among the more notable contributions to the physicochemical theory of multi-component phase change are the works of Hills, Roberts and Loper (Hills, Loper and Roberts, 1983; Hills and Roberts, 1988a,b; Loper and Roberts, 1987; Roberts and Loper, 1987) and those from a group at Purdue University (Bennon and Incropera, 1987a,b; Beckermann and Viskanta, 1988; see discussion on magma chambers above). These works have provided the much needed formalisms for the quantitative (computer) modeling of crystallization.

Combined kinetic and thermal models of crystallization in one-component systems initiated by Brandeis et al. (1984) suggested that the latent heat evolved through crystallization serves to buffer crystallizing systems close to equilibrium temperatures (i.e. small degrees of undercooling) for most cooling conditions. An extension of this model includes parameterization of the equations describing the time history of crystallization and one-dimensional conductive cooling of a dike or sill (Brandeis and Jaupart, 1987a,b,c,d) - it predicts (1) very short time scales for crystallization (2×10^5 sec for dike margins, 10^8 sec for intrusion interiors) and resulting small length scales for the crystallization (mushy) zone, and (2) final textures that are much more sensitive to the form of the nucleation rate (as a function of temperature) than to the growth rate. Spohn et al. (1988) include the effect of crystallizing a binary (eutectic) system (i.e. allowing crystallization to occur over a temperature interval); they find that solidification times and crystallization rates differ from those calculated for the one-component Stefan models due to smaller undercoolings maintained in a multi-component system and the extended temperature range of crystallization. This concurs with the prediction by Brandeis et al. (1984) for one component systems and is an important result in terms of the applicability of equilibrium thermodynamic models (e.g. Ghiorso, 1987) to crystallization of dynamic systems. It is important to note, however, that the crystallization models described do not include possible effects of convection or a realistic distribution of latent heat throughout the crystallization interval. Both of these complications *are* addressed in the equilibrium (no kinetics) model of Ghiorso (1990), who demonstrates their importance in modifying the cooling history of very large intrusions.

Experimental work on synthetic and natural rock systems continues to provide valuable information on crystallization kinetics and is reviewed in Cashman (1990a). Muncill and Lasaga (1987) show that crystal growth in the synthetic An-Ab system is interface-controlled for conditions of small undercooling; growth at larger undercoolings is non-linear with time, and apparently diffusion-controlled. Measured growth rates in the anhydrous An-Ab system can be modelled using the formalism of Lasaga (1982); this model does not work for the hydrous system, however, and points to the inadequacy of the Stokes-Einstein relation in relating diffusivity to viscosity (Muncill and Lasaga, 1988). Baker (1990) finds that the Eyring equation more closely approximates measured diffusivities and viscosities in rhyolitic and dacitic melts. Dissolution experiments by Zhang et al. (1989) point to the complexities of the dissolution process, including the documentation of common uphill diffusion. They thus caution against inferring bulk magma composition from crystal zonation, and apparently confirm the importance of a crystal's local environment (e.g. Pearce, 1987). Added confirmation for this observation lies in the documentation by Bacon (1989) of the

common occurence of accesory phases at the boundary (i.e. in the depletion zone) of growing crystals. Finally, classic diffusion-controlled growth (e.g. dendrites) is shown by Fowler et al. (1989) to be fractal on certain length scales; while quantitative analysis of fractal dimensions could ultimately supply cooling rate constraints, no calibrations are currently available for direct application to magmas.

An empirical approach to the determination of *in situ* crystallization rates was introduced by Marsh (1988) and Cashman and Marsh (1988) - as demonstrated by chemical engineers (e.g. Randolph and Larson, 1971), measured crystal size frequency distributions for crystals grown from a melt show a negative correlation between frequency and size that may result from either growth dispersion (different crystals of the same size having different growth rates) or from the small probability that any given crystal will advance from one size class to the next larger (Maaloe et al., 1989). Crystal growth rates determined from measured size distributions are small (10^{-10} - 10^{-11} cm/s; Cashman and Marsh, 1988; Cashman 1988; 1990a,b) and probably represent growth at very small degrees of undercooling. Measured crystal growth rates appear to be relatively insensitive to the specifics of the precipitating phase, the apparent undercooling (within limits) and to the melt composition (with the exception of extremely inviscid magmas, e.g. Peterson, 1990). Conversely, nucleation rates are strongly dependent on undercooling, as evidenced by the range of nucleation rates estimated for plagioclase phenocrysts, microphenocrysts and microlites of Mount St. Helens dacite (Cashman 1988; 1990b). This observation concurs with the prediction of Brandeis and Jaupart (1987) concerning the control exerted by the initial nucleation rate on final rock textures, and has the following ramifications for modeling magmatic processes: (1) growth rates appear constant at 10^{-10} - 10^{-11} cm/s for a wide range of melt and crystal compositions and cooling conditions, and can thus be used as growth rate estimates in magma chamber models, (2) comparison with experiments shows that these rates are consistent with crystallization at very small degrees of undercooling (e.g. Maaloe et al., 1989) - equilibrium thermodynamic models from melt evolution may thus be appropriate except for conditions of extreme undercooling (e.g. small dikes), and (3) while nucleation exerts the primary control on rock textures it is the process that is least well understood.

The kinetics of bubble nucleation and growth control processes of vesiculation and mechanisms of magmatic degassing, which in turn are related to magmatic transport dynamics and volcanic eruption styles. To apply such models quantitatively, we need not only theoretical models of vesiculation but also solubilities of volatile species as a function of pressure, temperature, melt composition and volatile composition. Most of the models relating volatile exsolution to eruptive style have been developed for basaltic systems, where the low viscosity of the melt allows significant migration of a volatile phase, and transitions between different two-phase flow regimes may control eruptive style (Vergniolle and Jaupart, 1986; Jaupart and Tait, 1990). Triggering of volcanic eruptions by fractional crystallization and resulting oversaturation of volatiles will be controlled primarily by the solubility relations of the volatile species involved, with eruptions much more likely for water-saturated rather than CO_2-saturated magmatic systems (Tait et al., 1989b); this eruption mechansim also appears to hold for the dome growth episodes at Mount St. Helens (e.g. Cashman, 1988; Anderson and Fink, 1989; Fink et al., 1990) and may be responsible for patterns of crustal uplift and subsidence at Yellowstone caldera (Dzurisin et al., 1990) and patterns of microgravity change at Poas volcano (Rymer and Brown, 1987). Continuous degassing can result in discontinuous eruption behavior depending on the balance between bubble accumulation in a foam layer at the roof of a chamber and the flow of bubbles up a conduit (Jaupart and Vergniolle, 1989; Vergniolle and Jaupart, 1990). The magmatic gas content and magma volume flux from the conduit then determines the dynamic structure of lava fountains, which in turn control the nature of the deposits produced by them (Head and Wilson, 1987; 1989).

Equations describing bubble nucleation and growth from a melt are similar to those of crystallization, except that simplifications can be invoked if vesiculation is assumed to involve only one or two diffusing species, spherical shapes and uniform interface properties. An additional growth parameter must be incorporated, however, to allow for expansion due to decompression. The analysis of Sparks (1978) continues to provide the foundation for much of this modeling, despite admitted simplifications (diffusional and decompressional growth of a single bubble in an infinite medium). A more elaborate numerical model has been developed by Toramaru (1989), who calculates the evolution of bubble size distributions during vesicualtion in ascending magmas with constant velocities. Bubble size distributions measured in basaltic scoria and in rhyolitic pumice show some correlations with melt composition and eruption style, but do not yield well-constrained kinetic parameters for vesiculation (Toramaru, 1990). While experimental constraints on degassing of silicate magmas are extremely limited, quantitative test of vesiculation (e.g. Lubetkin and Blackwell, 1988) and condensation (e.g. Schmelzer and Ulbright, 1987; Schmelzer and Schweitzer, 1987) in simpler systems provide a foundation for much of this work.

Experimental and observational constraints for the vesiculation models presented above are limited. Experiments in basaltic systems are facilitated by the calibration of FTIR as a tool to measure dissolved CO_2 in volcanic glasses (Fine and Stolper, 1986). Experimentally-determined solubilities of CO_2 in mid-ocean ridge basalts (MORBs) is approximately a linear function of pressure at the low pressures expected in MORB magma reservoirs (Stolper and Holloway, 1988). Measured CO_2 solubilities in MORBs from the Juan de Fuca ridge are higher than predicted, and may represent incomplete magma degassing due to rapid ascent (Dixon et al., 1988) and the difficulty of nucleating CO_2 bubbles at low pressures (Bottinga and Javoy, 1989; 1990a,b). However, there is also evidence that basalts that reside in magma storage reservoirs for any length of time undergo significant degassing (e.g. Gerlach, 1989a,b; Fisher and Perfit, 1990; Dixon et al., 1990), evidence in agreement with the prediction that MORBs may be saturated with CO_2 when generated in the mantle, and thus exist as a separate phase prior to accumulation in the magma chamber (Bottinga and Javoy, 1989; 1990a,b). Thus ocean basalts containing anomalous amounts of volatiles (e.g. Gill et al., 1990; Sarda and Graham, 1990) may represent magmas transported directly from depth; in fact, Sarda and Graham (1990) measured negative exponential vesicle size distributions in mid-ocean ridge "popping rocks" that are identical to the linear crystal size distributions of volcanic rocks described above, and suggest continuous nucleation and growth of bubbles during magma ascent.

The past four years have seen great advances in both measurements and thermodynamic models of water solubilities in

aluminosilicate melts, building on the work of Silver and Stolper (1985) and answering the plea of McMillan and Holloway (1987) in their review of the available solubility data for water in aluminosilicate melts at that time. Evidence for Henrian behavior of water at total water contents up to several weight percent (i.e. within the observed range for natural silicate melts) allows thermodynamic modeling of water solubility (Silver and Stolper, 1989; Silver et al., 1990). In addition, water speciation in silicic glasses is temperature-dependent and may be used as a glass geothermometer, although blocking temperature is a function of quench rate (Stolper, 1989). Wasserburg (1988) demonstrates that when both molecular water and OH species are considered, transport behavior (diffusion) is nonlinear at intermediate and dilute water concentrations. Pre-eruption volatile contents of ash-flow tuffs have been estimated by measurements of H_2O and CO_2 contents of phenocryst melt inclusions (e.g. Anderson et al., 1989; Dunbar et al., 1989; Hervig et al., 1989). Results from these studies suggest that volatile saturation occurs at least at the top of large silicic chambers throughout much of their evolution, although details of degassing mechanisms involved with eruptions of silicic systems remain elusive and have just started to be addressed for smaller silicic systems (e.g. Eichelberger et al., 1986; Newman et al., 1988; Anderson and Fink, 1989). For example, volatile exsolution in the Inyo Craters system may have occurred in two stages, with extensive degassing due to decompression (Eichelberger et al., 1986; Eichelberger, 1989) prior to degassing by second boiling as the result of crystallization and volatile supersaturation of the melt (Westrich et al., 1988; Swanson et al., 1989). Continued work in this area should provide important constraints for interpretation of measured volatile data plus a better understanding of the relationship between rates of degassing and timescales of structural relaxation of viscous melts (e.g. Dingwell and Webb, 1989; Kress et al., 1989).

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