REVIEWS in MINERALOGY
Volume 26

CONTACT METAMORPHISM

DERRILL M. KERRICK, Editor

Series Editor: Paul H. Ribbe

MINERALOGICAL SOCIETY OF AMERICA

Chapter 2 — George W. Bergantz

PHYSICAL AND CHEMICAL CHARACTERIZATION OF PLUTONS

INTRODUCTION

The purpose of this chapter is to inventory and describe some of the generic features of intrusive systems which pertain to the understanding of contact metamorphism. Magmas are the sources of heat, mass and mechanical energy that yield contact metamorphism and associated deformation, and an appreciation for the manner in which the intensive and extensive variables vary during plutonism may aid in understanding the temporal and spatial details of contact metamorphism. A wide variety of thermal histories and corresponding styles of self organization are possible given the variety of initial and boundary conditions attendant with plutonism. This review concentrates on the physical and chemical processes in plutons. The scope of this review is limited largely to granitoids; literature on mafic systems or mid-ocean ridges is not included.

The purpose of this review is to provide quick access to the current work and paradigms relating to plutonic systems. The first priority is to direct the student of contact metamorphism to the literature useful in constraining magmatic processes. Given the extremely broad nature of the subject and the space available, it was decided to sacrifice detail for scope; a functional understanding of magmatic processes will require further study of the works cited herein. Petrogenetic schemes or compositional-tectonic associations are not discussed. The importance of petrogenesis in the broad characterization of plutons is recognized, however, a detailed treatment of this topic is outside the scope of this review. Petrogenesis and regional tectonics are treated in a number of timely summaries and the interested reader is encouraged to look there. First among them are the general reviews of plutonism by Pitcher (1978, 1979, 1987), Whitney (1988), Zen (1988), and Chappell and Stephens (1988). A number of excellent compilations of magmatism in a regional context are available: reviews of Andean magmatism (Atherton and Tarney, 1979; Harmon and Barreiro, 1984; Pitcher et al., 1985) also see Hildreth's (1987b) review of Pitcher et al. (1985), studies of magmatism in North America (Anderson, 1990; Ernst, 1988), and compilations of papers discussing magmatism around the Pacific margin (Kay and Rapela, 1990; Roddick, 1983). Anthologies of related interest are those edited by Vielzeuf and Vidal (1990) which addresses granulites and crustal evolution, and Mereu et al. (1989) on the physical properties and processes of the lower crust.

Contact metamorphism as a conjugate system

The transfer of heat and mass from the magma to the country rock comprises what is known as a conjugate, or coupled, system (Bejan, 1984; Bergantz and Lowell, 1987). The important characteristic of conjugate systems is that the quantitative modeling of the transfer of heat and mass from the magma to the country rock requires explicit consideration of the heat transfer systematics on *both* sides of the contact zone. A direct physical analogy of this is the window of a house on a cold day. Heat is being brought up to the window by whatever processes may be operating in the room, such as a forced air furnace or perhaps natural convection from a wood stove. The heat is being transferred through the window to the cold thermal reservoir outside, and conditions outside the house will determine how efficient the environment is at removing the heat from the window. Thus, the temperature of the window reflects the balance of the heat transfer processes on each side of this coupled system. If the window feels warm to the touch, the heat transfer in the room is able to keep up with the losses to the environment and so the rate-limiting step is the thermal resistance associated with transfer in the outside environment. If the window feels cold to the touch, which is more often the case, that indicates that the rate-limiting processes are associated with heat transfer in the room bringing heat up to the window. The point is that the temperature of the geological window, which is recorded in the contact metamorphism and is time dependent for any case of geological interest, provides a constraint for the possible processes that can occur in the coupled magma-country rock system.

The implication of this is that the temperature and spatial extent of contact metamorphism can possibly provide some insight into the coupled nature of the processes operating on both sides of the contact. This suggests that one might be able to "invert" the geologically determined conditions of metamorphism to choose among the possible processes of heat and mass exchange. It is not difficult to imagine a variety of interactions between the pluton and the country rock: (a) conduction on both sides of the contact with or without phase change being explicitly included (Bowers et al., 1990), (b) conduction in the magma with hydrothermal convection in country rock (Cheng, 1978, 1981; Norton and Taylor, 1979; Parmentier, 1979, 1981); and a study by Johnson and Norton (1985) that deserves more attention, (c) convection on both sides of the contact (Bergantz and Lowell, 1987), (d) cracking of the solidifying magmatic rind which permits hydrothermal fluids to cross the contact (Carrigan, 1986; Lister, 1974), or (e) conduction in the country rock with convection in the magma, with the possibility of simultaneous crystallization and melting (Bergantz, 1991). It is likely that more than one of these processes may operate during an episode of contact metamorphism, particularily if the magma is subject to open system behavior, e.g., additional magma enters the system. A knowledge of the contact metamorphism, such as the spatial distribution of maximum temperatures, provides a constraint with which the magmatic history must be consistent. This has some appeal for those working in magma dynamics, as it is very difficult to constrain processes in the magma from the temperatures recorded in the pluton.

Undoubtedly, this difficulty is largely due to the fact that plutons represent the end result of what is a complex chemical and mechanical history. This complexity is manifested a number of ways: in the ubiquitous disequilibrium mineral textures formed during both growth and subsolidus conditions, the isotopic evidence for assimilation and other open system behavior, mineral fabrics formed during magmatic flow with superposed near or sub-solidus deformation features, and in the diverse sequence of chemistries and eruptive styles exhibited by extrusive systems. The characterization of plutons must borrow substantially from the study of volcanic systems where magmatic conditions can be reconstructed without the veil of subsolidus transformations. Even some of the simplest magmatic systems appear to have complex histories. For example, scientific drilling at the Inyo Dome, California, thought to be a simple shallow silicic system revealed a complex relationship between high and low Si magmas, mixing and transport (Vogel et al., 1989). Complex patterns of re-intrusion, magma mingling and mixing, and segregation can all occur in what is thought to be a single magma chamber as suggested in a number of studies; examples include Vesuvius (Civetta et al., 1991) and the chamber that existed below Mount Mazama (Bacon and Druitt, 1988). A fascinating degree of complexity in space and time of magma types and eruptive styles has been documented at Katmai, Alaska, where a plexus of small, compositionally distinct magma chambers are postulated (Hildreth, 1987a). A study that attempts to explicitly address the compositional, spatial and temporal relationships between plutonism and volcanism is the study of Lipman (1988). Even though volcanic rocks provide the best circumstances from which magmatic intensive variables can be estimated, it can be difficult to demonstrate equilibrium (Frost and Lindsley, 1991). In addition, the depth at which the pluton forms can radically impact the cooling and crystallization history by controlling the timing and extent of volatile exsolution (Swanson et al., 1989; Westrich et al., 1988).

The implication of this is that the magmatic conditions that existed at the time of contact metamorphism may be different from those preserved in the pluton, and establishing the conditions which existed when the pluton was a viable "chamber" is subject to uncertainty. Another complication arises in interpreting magmatic history in light of the contact metamorphism: more than one magmatic history can yield the same metamorphic history. In fact it appears that thermal contact metamorphism is consistent with the simplest scenario imaginable: instantaneous intrusion followed by conductive cooling. This is discussed in more detail below.

The most important aspect to consider when addressing the thermal evolution of plutons is the process of solidification. The enthalpy flux that ultimately yields contact metamorphism is invariably accompanied by an increase in crystallinity in the pluton. This can occur at the margins and the magma chamber may form a rind of crystals that propagates inward as cooling continues, perhaps conductivly in a manner analogous to the Hawaiian lava lakes. Alternatively, the crystallinity may increase in the center or in a distributed fashion in the melt, and the pluton will solidify uniformly, perhaps undergoing sustained and vigorous convection. These two end-members can yield very different calculated thermal histories, depending on the assumptions involved in parameterizing the heat transfer (Bergantz, 1991). The solidification process will also have a dramatic effect on the physical properties of the magma namely, the viscosity and the density of the magma will change as the crystallinity and volatile content increase. This in turn influences the processes driving the heat transfer and a feedback is created that is difficult to generalize. Laboratory and numerical experiments of solidification, as shown in Figure 1, reveal a rich diversity of solidification and cooling histories depending on the composition of the liquid,

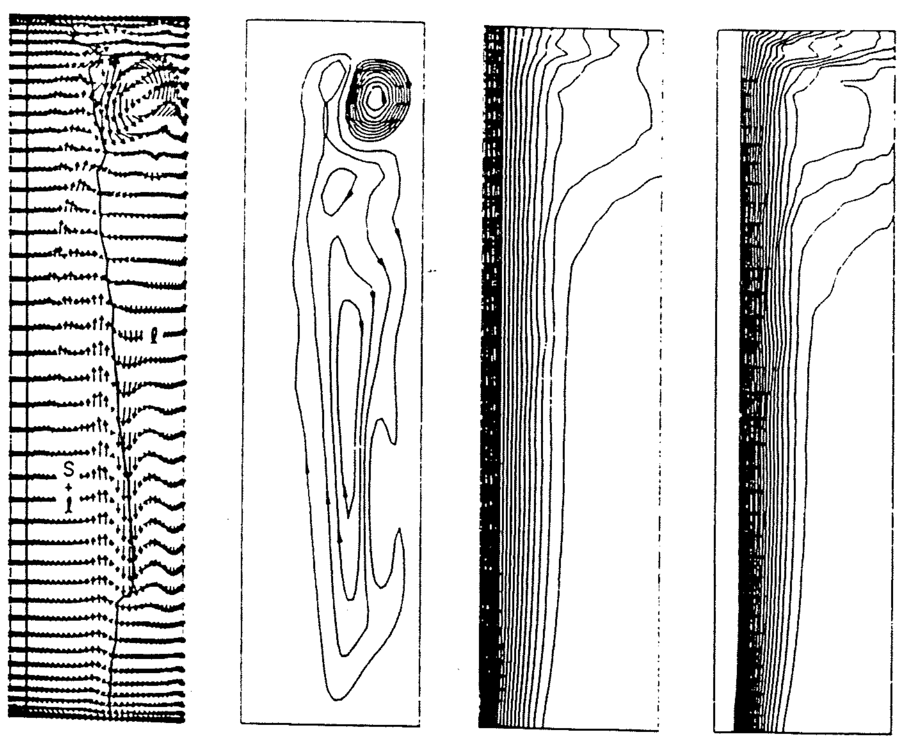

Figure 1. Output from the numerical simulation of the solidification of a binary, taken from Bennon and Incropera (1987). The left wall of the figure is kept at a constant temperature that is below the solidus, the right wall is insulated. The first panel shows the progression from all solid to a solid-liquid mixture into a pure liquid. The arrows indicate direction and magnitude of flow, note the upward flow near the right margin of the mush due to compositional effects on buoyancy. Also note the complex velocity field. The second panel gives the streamlines, the third is the temperature field and the fourth is the isocomposition contours. Although not directly applicable to magmatic systems, numerical modeling of this kind is currently being adapted to geologically relevant conditions.

and the physical properties of the system (Beckermann and Viskanta, 1988; Bennon and Incropera, 1987; McBirney et al., 1985; Oldenburg and Spera, 1991). These studies provide a look at the current state-of-the-art in the formulation and modeling of solidification processes.

Although none of the laboratory and few of the numerical experiments use materials that are directly analogous to magmas, they reveal some of the generic processes that occur in plutons. One of the most important of these is the partitioning of the body into a mushy zone near the contact where crystals and melt form a self supporting framework and an adjacent slurry where crystals reside in an expanse of melt. This partitioning has two important implications for contact metamorphism: (a) the contact between a rheologically viable magma and it's solid container propagates inward with time away from the original intrusive contact, thus the solidifying margins of the magma also undergo contact metamorphism in the sense that cracking and interaction of fluids is possible, (b) the conjugate nature of the heat transfer requires explicit consideration of the phase changes in the magma (Bergantz, 1991), and the rate of propagation of the solidification front and any convection in the magma can only occur at rates consistent with the rate of heat loss through the country rock. Thus, the heat transfer can be conceptualized as a heat transfer "circuit" with thermal resistances in a series arrangement (a common analogy used in engineering textbooks) and the magma cannot pump out heat any faster than the country rock can carry it away. In fact the optimal cooling time for a perfectly mixed (convecting) hot fluid body surrounded by a conducting medium is only twice the cooling time if cooling by conduction alone. This partitioning of the magma chamber into a solidifying margin and an adjacent slurry in the interior, which may be subject to re-intrusion, precludes the use of simple dimensionless heat transfer parameters when describing the progress of solidification or the rate of heat loss from the body; the phase change process must be explicitly considered.

This review discusses those elements of plutons that influence the physical processes that yield contact metamorphism. The focus is largely on heat transfer and less on mass transfer. Magmatic fluids (volatiles) clearly play an important role in contact metamorphism, as evidenced by ore deposits (Burnham, 1979b). However, quantitative models of the "second boiling" process coupled with the mechanical processes of country rock fracture, with attendant changes in permeability, have yet to be developed. The emphasis on heat transfer is consistent with the classical treatment of contact metamorphism, reflecting in part the quantitative accessibility of this approach. These and other generic aspects of the physical evolution of magmas are discussed in the reviews of Marsh (1989a) and Morse (1988). Establishing the physical history of a pluton requires a careful consideration of the magmatic intensive and extensive variables as the rate at which the pluton loses heat will depend on these in combination. We consider these below.

INTENSIVE VARIABLES

Sequence of crystallization

Establishing the crystallinity at the time of intrusion and the subsequent sequence of crystallization of the magma is difficult but is often one of the only ways to practically bracket magmatic intensive variables. A knowledge of the sequence of crystallization is used to infer the temperature, pressure, water content and rheological properties of the magma. In practice, the order of crystallization is estimated petrographically from crystal morphologies and other textural criteria and then compared to laboratory and computer experiments of solidification. The uncertainties associated with this approach originate in the difficulty of interpreting plutonic textures and in comparing these with experimental systems whose components only partially match those of the pluton being studied.

Since the seminal work of Tuttle and Bowen (1958) a number of experiments have

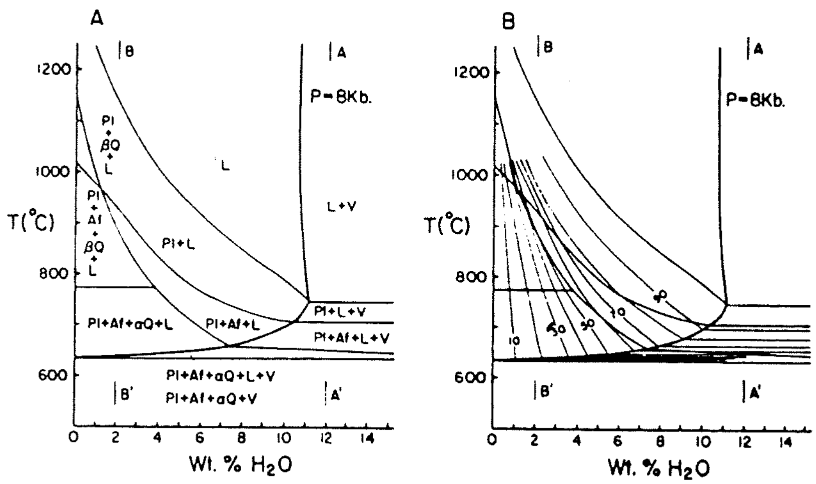

Figure 2. Temperature-XH_2O diagrams for synthetic granite compositions. From Whitney (1988). These diagrams permit an estimate of the onset of saturation and the paragenetic sequence. The contours in the right hand panel give approximate percent of melt present. Information of this kind is crucial to developing transport and thermal models of magmas.

been done to establish phase relations in the granitic system (Luth, 1976; Wyllie, 1988). Both solidification and melting experiments for common granitoid *rock* compositions have been done under a variety of pressures and water contents (Huang and Wyllie, 1986; Naney, 1983; Stern and Wyllie, 1981a,b; Whitney, 1975). Of particular interest are the experiments of Naney (1983) and Huang and Wyllie (1986), whose experiments include the ferromagnesian silicates. A revealing and useful format for presenting the experimental data is found in plots of temperature vs. weight percent water for a given composition at a given pressure. Curves are drawn to illustrate saturation of a given phase and thus one can follow the sequence of crystallization for a given water content and even estimate when the system becomes saturated. Whitney (1988) provides crucial additional information: curves of volume percent liquid. An example of this is shown in Figure 2. These are critical to transport modeling of the solidification process (Bergantz, 1990). Melting experiments of broadly granitic materials have been done to simulate the process of granite generation by crustal fusion (Presnall and Bateman, 1973; Wolf and Wyllie, 1989; Wyllie, 1977) which seems to require very high temperatures to generate melt fractions in sufficient quantity to form extractable magmas. This has motivated melting experiments of metaigneous (Beard and Lofgren, 1991; Rushmer, 1991) and metapelitic protoliths (Patiño-Douce and Johnston, 1991; Vielzeuf and Holloway, 1988) which generate large amounts of melt at temperatures like those expected in lower crust subject to intrusion by basaltic magma (Bergantz, 1989). Thus, for a given composition and water content, the phase relations for both the crystallization and melting of granitic systems are broadly known, although much work needs to be done with other volatile species and to further refine the thermodynamic database so that the solidification progress can be modeled numerically.

Once sufficient experimental data exist to establish the thermodynamic properties of the crystal-melt system, computer algorithms can be developed which allow one to simulate crystallization in some detail (Ghiorso, 1985; Ghiorso and Carmichael, 1985, 1988a; Nekvasil, 1988b; Nielsen, 1990). Each of these authors uses a different approach to the numerical treatment of crystallization, and each algorithm gives good agreement between predicted and naturally occuring assemblages for certain compositional ranges; the formulation of Nekvasil is specifically designed for silicic systems. One element missing from these numerical crystallizers is the ability to model the presence of hydrous phases.

This is simply due to the absence of the appropriate thermodynamic data and is not an inherent difficulty in the numerical approach. The numerical approach to crystallization is extremely powerful in that it permits one to couple transport models of time dependent heat and mass transfer to geologically relevant conditions (Bergantz, 1990). One example of this is the work of Bowers et al. (1990) which uses the algorithm of Nekvasil to estimate the temperatures at which the phases appear and their contributions to the time dependent enthalpy changes in the magma. Computer models of phase changes coupled with conjugate heat and mass transport provide an important and exciting direction for the elucidation of the generic features of contact metamorphism.

Although laboratory and numerical experiments allow predictions for the appearance of the phases as the pluton cools, it is more difficult to estimate the crystallinity during actual ascent and at the time of intrusion. Extrusive rocks contain up to ~50% crystals (Marsh, 1981). It can be difficult, however, to confidently know when in the magmatic history the crystals grew (compare the phase diagrams in Whitney (1975) for 2 and 8 kbar). Unlike lavas or ash flows which often reveal chilled margins and an estimate of the crystallinity at the time of eruption can be obtained, textures in plutonic rocks invariably represent re-equilibration during prolonged periods near solidus temperatures. Apart from pegmatitic and obviously volatile-rich aplitic apophyses, plutonic rocks most often show a uniformity in grain size, often right up to the contact. There can be little question that nucleation is heterogeneous and there is little evidence for the magma having been in a superheated condition. Crystal growth rates appear constant at 10^{-10} - 10^{-11} cm/s for a wide range of melt and crystal compositions and cooling conditions as discussed in the comprehensive review by Cashman (1990). These growth rates are consistent with very small amounts of undercooling which indicates that undercoolings are a negligible part of the thermal budget. The heat transfer systematics, whether conductive or convective, will not be influenced by them.

Volatiles

Volatiles have a substantial influence on magmatic systems: they can substantially lower solidus temperatures and vary the sequence of crystallization, depolymerize the melt structure and hence reduce viscosity and density, influence the ascent history by the onset of saturation, and induce complex patterns of repeated fracturing and partial quenching of the solidifying pluton. Despite the importance of volatiles in understanding magmatic history, it has been very difficult to rigorously quantify the thermodynamic state (speciation) of volatiles in a silicate melt and also very difficult to model the multiphase behavior of a crystal-melt-volatile system from a continuum mechanical approach. These problems are exacerbated when considering actual plutons as one can often only crudely guess the original volatile content, however fluid inclusions may permit estimates of magmatic volatile composition. There are four ways to estimate volatile content (Clemens, 1984): (1) by direct measurement, (2) geological inference, (3) thermodynamic calculation, and (4) experimentally. Of necessity, much of our understanding of magmatic volatiles comes from the study of volcanic systems and extrusive rocks, although it is difficult to confidently extrapolate from eruptive conditions in volatile stratified magma chambers to plutonic conditions. Summary discussions of the role of volatiles in granitoid magmatism can be found in Burnham (1979a,b) and Whitney (1988).

Melt inclusions in phenocrysts and the dissolved water in volcanic glass provide a means of determining the water, carbon dioxide, fluorine, sulfur and chlorine contents of magmas at the time of eruption. Water is the most abundant volatile component: measuring 5 wt % in the Fish Canyon Tuff (Johnson and Rutherford, 1989b), 4.3 wt % in the Taupo volcanic center (Dunbar et al., 1989; Hervig et al., 1989), 4-6 wt % in the Bishop Tuff (Anderson et al., 1989), 2-4 wt % in the Bandelier Tuff (Sommer and Schramm, 1983), and 4.1 wt % from Obsidian Dome (Hervig et al., 1989). Many of these volcanic systems appear to have a gradient in water content suggesting that a vertical gradient in volatile content existed in the magma chamber. Understanding the manner in which water is speciated in the melt has been more difficult. The influential works of Burnham (1979a,b) proposed using a solution model based on the simple system albite-water. This model has appeal in that it provides a useful and straightforward means of quantifying the thermodynamic state of water, although there are complexities and questions regarding speciation that remained to be addressed. In their summary article, McMillan and Holloway (1987) demonstrate that molar water solubility increases with decreasing silica in binary and pseudobinary silicates. Silver et al. (1990) note that hydroxyl groups are the dominant hydrous species at low water contents, and that increasing silica content and K over Na leads to an increase in the molecular water relative to hydroxyl. The Henry's law behavior of water provides a means of putting the thermodynamic modeling of water in silicate melts on a firmer basis.

Given the chemically complex nature of granitic melts, estimating water content in practice is often done by comparing the petrogenetic sequence as determined petrographically with phase equilibria as determined by laboratory experiments or computer models (Maaloe and Wyllie, 1975); see discussion on the sequence of crystallization given above. Another way to bracket water content is to evaluate the role of water in melt-forming reactions as done by Wyllie et al. (1976), and Clemens (1984). None of these methods yield a precise estimate of volatile content and the presence of hydrous phases can only give minimum values.

Carbon dioxide and fluorine are also important volatiles in magmas. The solubility of CO_2 in silicate melts appears to be low (Holloway, 1976), hence many magmas may be saturated in CO_2 for much of their history which will influence the activity of water. This in turn could have a profound effect on the crystallization and ascent history. Experiments by Peterson and Newton (1990) demonstrates that CO_2 in silicate liquids can delay the crystallization of biotite, mafic material is retained in the melt. Stolper et al. (1987) discuss the mechanisms of CO_2 dissolution. Anderson et al. (1989) note that there is an inverse correlation between water and CO_2 concentration in the Bishop Tuff; the CO_2 concentration varying from 0.005 to 0.035 wt %. Assessing the role of CO_2 in the thermomechanical history of any particular granitoid pluton is difficult, few experiments with CO_2 saturated and water bearing multi-component silicate melts have been done. Fluorine contents can range from tens of ppm to several percent (Bailey, 1977) and is usually held in biotite and hornblende; the F/OH ratio in biotite was used by Ague and Brimhall (1988b) to infer contamination of granitoids in the Sierra Nevada. Dunbar et al. (1989) report fluorine contents of about 450 ppm in rocks from the Taupo volcanic center. Manning (1981) considered the liquidus phase relationships in the water saturated Qz-Ab-Or system and found that the minimum liquidus temperature fell 100°C from that of the fluorine free system. Manning posits that there may be a continuum between magmatic and hydrothermal conditions in mineralized granitoids. Pichavant and Manning (1984) report occurances of up to 3.2 wt % fluorine in tourmaline granites and topaz granites. They provide evidence that these granites formed from highly differentiated residual melts. Pichavant (1987) considered the influence of boron on the phase relations. The boron concentration does not significantly influence the phase relations in the haplogranite system; the boron content of magmas does not exceed about 1 wt %.

Based on thermodynamic calculations (as opposed to thermomechanical), it is possible to evaluate the balance between the composition of the magmatic source material and initial depth, the ascent distance, and initial volatile content. Sykes and Holloway (1987), Hyndman (1981) and Marsh (1984) examine the energetics during ascent and demonstrate that water must be considered to yield realistic estimates of magmatic conditions. This limits the range of initial conditions and source compositions of those magmas which can ultimately reach the surface and become erupted: relatively hydrous melts are prohibited from travelling far from their source as the solidus is encountered at depth (Fig. 3). As crystallization proceeds in a magma, volatile elements can be preferentially partitioned into the melt phase until saturation occurs. The pressure increase accompanying boiling can induce fracturing of the crystallizing margins of the pluton,

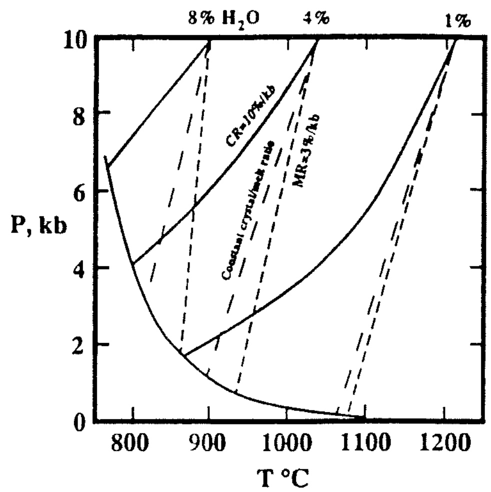

Figure 3. Ascent trajectories calculated by Sykes and Holloway (1987) for model system albite-H_2O. Dashed lines represent a constant crystal/melt ratio and dotted lines represent a melting rate of 3%/kbar.

providing a means for the volatiles to escape. Further solidification permits this process to continue yielding a second and subsequent boilings (Burnham, 1979b). There is no doubt that such processes are important in the thermal history of shallow plutons; a complete quantitative description of this process awaits development.

Estimating temperature and pressure

There are a variety of methods to estimate pluton temperature and pressure; Zen (1989) discusses the generic features of the two approaches that are commonly used to estimate conditions in plutons. One approach is extrinsic, where pluton temperatures and pressures are determined from the character of the country rock. There can be large uncertainties in this method, as discussed by Hodges and McKenna (1987). The second is intrinsic, which relies on the specific features of the pluton itself. With both methods it is difficult to confidently determine when in the history of the magmatic system the temperature and pressure were set. The temperatures and pressures recorded by magmatic mineral assemblages are usually those of consolidation of the magma and may not represent the same point in time at which contact metamorphism was achieved (Zen, 1989); see discussion on open systems below. In addition, confidently estimating magma temperature and pressure in plutons is difficult due to the re-equilibration of the relevant phases, and the inevitable loss of volatiles. Temperatures thus usually represent minimums and may actually be related to secondary, subsolidus processes. There is a growing appreciation that estimating conditions in volcanic rocks, which ostensibly represent the best sample of a thermodynamically contiguous system, requires a careful assessment of the equilibrium assumption (Frost and Lindsley, 1991). Nonetheless, a number of intrinsic geothermo-barometers have been applied to plutonic rocks and we will discuss the systematics of a few of them below. For a general review of the principals behind their application see Bohlen and Lindsley (1987).

There are two methods generally used to determine temperature in plutons. One approach is to compare the paragenetic sequence of precipitation of minerals with the phase experiments discussed above, e.g., Naney (1983), and as done by Hill (1988). This is obviously a rather crude way to estimate temperature, however it has the advantages of being inexpensive and usually straightforward. The other approach is to evaluate the compositions of coexisting phases that are thought to represent equilibrium conditions such as coexisting plagioclase and potassium feldspars (Whitney and Stormer, 1977), amphibole and plagioclase (Blundy and Holland, 1990) and/or coexisting oxides (Frost and Lindsley, 1991; Whitney and Stormer, 1976). The experiments of Elkins and Grove (1990) provide the latest attempts at calibration of a two feldspar thermometer following the approach of Fuhrman and Lindsley (1988) and others (Ghiorso, 1984; Green and Udansky, 1986). The experiments of Elkins and Grove (1990) were done at 700°-900°C and 1-3 kbar under water saturated conditions. In most cases their measurements agreed with their thermodynamic model to within 20°C and agree well with Fe-Ti oxide temperatures obtained from volcanic rocks where the temperature could be independently constrained. The oxides are particularily difficult to work with as they commonly have exsolved and the assumptions involved in recreating the equilibrium assemblage often are untenable.

Magmatic hornblende has been proposed as a geobarometer. Hammarstrom and Zen (1986) noted that the total Al content of hornblende from calc-alkalic plutons increases linearly with increasing pressure of crystallization. This led to the development of an *empirical* barometer based on the assemblage plagioclase + quartz + potassium-feldspar + biotite + amphibole + titanite + Fe-Ti oxides (magnetite or ilmenite). There are two groups of calibrations: those of Hammarstrom and Zen (1986) and Hollister et al. (1987) which are empirical and rely on corroborating pressures as determined by the country rocks or the presence of magmatic epidote, and those of Johnson and Rutherford (1989a) which were derived from reversed experiments with f_{O_2} buffered and vapor present . At present it would appear that the Johnson and Rutherford (1989) calibration would be preferred as the laboratory conditions provide a more confident estimate of total pressure. The experiments of Rutter et al. (1989) on partially melted tonalite demonstrate that the total Al content is very sensitive to the specific mineral assemblage and hence caution must be used when garnet or other phases are present. Blundy and Holland (1990) propose a different substitution and argue that the Al substitution is temperature dependent and that amphibole equilibria are not appropriate for geobarometry. They propose an amphibole-plagioclase geothermometer instead. In a regional study of granitoids, Vyhnal et al. (1991) explore both the temperature and pressure dependence of Al in hornblende and conclude that it may be difficult to discriminate both temperature and pressure effects because they are both linked to the solidus. They propose that this might occur due to the occurrence of more than one substitution reaction. It thus seems that the hornblende geobarometer should be used with caution, and that much work remains to be done before the hornblende geobarometer has a more complete thermodynamic basis.

The presence of magmatic epidote has been used as a geobarometer by Zen and Hammarstrom (1984) who argue that magmatic epidote is a high pressure, near solidus phase that appears as a reaction product involving hornblende and the melt. The laboratory experiments of Naney (1983) produced epidote at a pressure of 8 kbar and epidote has been observed in the chilled margins of dike rocks and rhyolitic lava flows (Dawes and Evans, 1991; Evans and Vance, 1987) establishing that it can occur as a near-liquidus mineral. The difficulty in using the presence of epidote as a geobarometer is in establishing whether the plutonic epidote is indeed magmatic and under what water and total pressures the epidote grew and last equilibrated. Oxygen fugacity also exerts a strong control on epidote stability. Dawes and Evans (1991) note that there are three types of magmatic epidote in the dacitic dikes of the Front Range, which were emplaced at 2 kbar (or less) and provide evidence that epidote formed at pressures of about 8 kbar. As with magmatic hornblende, a complete thermodynamic characterization of epidote awaits development and hence its use as a geobarometer is subject to some uncertainty.

Physical properties

The ability of magmas to transmit heat to the solidification front will depend on the thermophysical and transport properties of the magma. The two properties that vary the most, and also influence the heat and mass transfer the most, are the viscosity and the

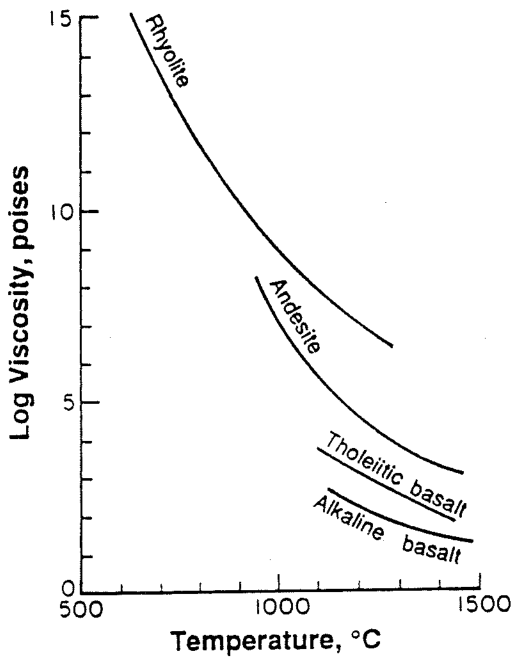

Figure 4. Viscosities of some common igneous rocks. From McBirney and Murase (1984).

density. Both will vary strongly as crystallization proceeds and they generally vary in an opposing manner: as the density of the liquid goes down, the viscosity of the melt goes up, although the presence of volatiles complicates this simple picture. Volatiles can depolymerize the melt structure and induce non-monotonic changes in melt density as cooling and crystallization proceeds. Before we begin a discussion on physical properties, the reader should be mindful of what the viscosity and density in the cited works refers to: the property of the melt, or the ensemble melt plus crystals. General reviews of the viscosity of magmas can be found in McBirney and Murase (1984), Ryan and Blevins (1987), and Ryerson et al. (1988). The work of Lange and Carmichael (1990) gives the most recent formulations for computing melt densities. One useful summary of the thermophysical properties, conductivity, specific heat, etc., is the compilation in Touloukian et al. (1981).

Viscosity in silicate melts is a strong function of crystallinity, temperature and composition (Fig. 4). For example, the viscosity for basaltic compositions may vary three orders of magnitude over three hundred degrees, for rhyolites (dry) it may be nine orders of magnitude over six hundred degrees. It has been observed that the strong dependence of viscosity on temperature, for a fixed composition, can be given by an Arrhenius type of relation:

$$\eta = \eta_0 e^{E/RT} \tag{1}$$

where η is the shear viscosity (the Greek letter μ is also commonly used), η_o is a constant, E the activation energy, R the gas constant and T the absolute temperature. In practice, the values of η_o and E are not known for many compositions and alternative computational methods have been developed. The viscosity of a crystal-free melt (and hence compositionally invariant) liquid can be calculated from the algorithm of Bottinga and Weill (1972) at any temperature. This algorithm is based on summing the contributions to viscosity from the proportions of the partial molar oxides in the melt. We note in passing that the dynamic, or shear, viscosity is often reported in units of poise (gm cm-1 s-1) or Pascal seconds (kg m-1 s-1); the conversion is 10 poise to everyl Pascal second (Pa s). Current custom discourages the use of cgs units and the reader is encouraged to use Pa s as the proper unit of dynamic viscosity.

As cooling of a magma proceeds crystals grow and their presence will influences viscosity. To evaluate the viscosity of a crystal-melt mix, a suspension, requires that the aspect ratio, *volumetric* concentration and relative particle size distribution of the crystals be known. Metzner (1985) reviews a number of aspects related to the viscosity of suspensions and concludes that for melts with a liquid viscosity of greater than 10^2 Pa s, the presence of particles does not induce an order of magnitude change in viscosity (at a fixed composition and temperature) until the volume fraction exceeds about 0.5. These results are for a suspension of uniform spheres. Experiments on picritic compositions by Ryerson et al. (1988) reveal that the viscosity of the suspension was independent of the crystallinity up to about 0.25 volume fraction. Marsh (1981) argues that a volume solid fraction of about 0.5 represents a rheological locking up point or critical melt fraction, an idea we develop in more detail below. Metzner (1985) reviews a number of functional forms for suspension viscosity and proposes the following expression:

$$\eta_r = \left[1 - \left(\phi/_{A}\right)\right]^{-2} \tag{2}$$

where η_ρ is the relative viscosity which is the ratio of the crystal-free viscosity at a given temperature and composition, to the actual suspension viscosity, ϕ is the volume fraction and A is a constant whose value depends on the particle size distribution and particle aspect ratio; values of A vary from 0.68 for smooth spheres to 0.18 for materials with an aspect ratio of 27.

In principle, absolute particle size plays no role in expressions for the viscosity of a suspension; this is in keeping with the usual continuum assumptions that attend the definition of viscosity. However, the relative particle sizes can influence the viscosity dramatically, for example Metzner (1985) notes that at a volume fraction of 0.66, the relative viscosity varies from 1200 for a unimodal size distribution to 23 for a tetramodal size distribution!

In practice then, the calculation of viscosity can proceed as follows: the viscosity of the crystal-free liquid is calculated at a given temperature and composition by using an algorithm such as that given in Bottinga and Weill (1972). To account for the additional effects due to the presence of crystals, the calculated crystal free viscosity is multiplied by expression (2) to yield the final expression for viscosity.

The strong variations in viscosity that accompany an increase in crystallinity naturally give rise to a concept of an effective "locking" viscosity where the crystallinity is so high that the crystal-melt mixture no longer exhibits fluid-like behavior. This mechanical threshold has been considered in a number of studies, see the review in Bergantz (1990), and the percentage of melt remaining at the time this threshold is reached is generally known as the critical melt fraction. Although the critical melt fraction will depend on a variety of factors, including strain rate and temperature, a value of about 50% is frequently invoked. This is consistent with the observation of Marsh (1981) that one rarely sees lavas with more than about 50% crystals suggesting that higher crystallinities preclude eruption and, by inference, transport. These ideas are developed in more detail below.

Changes in the density of magmas provides the mechanisms which drive petrologic diversity and control the distribution and temporal patterns of plutonism. There are two ways to consider magmatic density as cooling proceeds, either as a pure liquid where the crystals are assumed to be removed immediately or as a crystal-melt mixture where the density is the ensemble density of both the crystals and melt. The second of these requires that the solids phases and amounts be precisely known. An algorithm to actually calculate

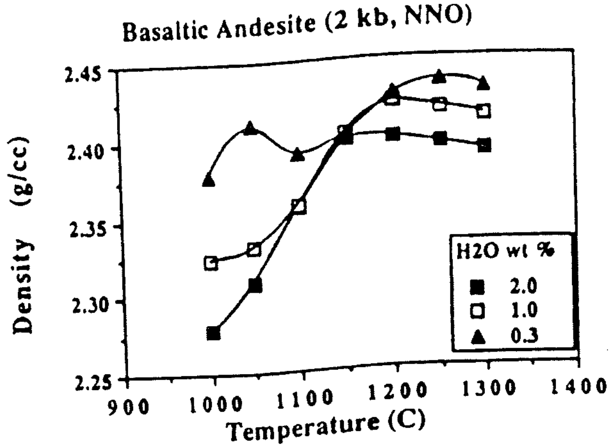

Figure 5. Density of basaltic andesite residual liquid as a function of three water contents. From Lange and Carmichael (1990). Note the non-monotonic behavior of the density with temperature and the pronounced role water can play in producing complex density variations.

density can be devised much like that used to calculate viscosity, a simple equation relating volume of the liquid to the partial molar volumes of the components can be written and density calculated accordingly. This is comprehensively discussed by Lange and Carmichael (1990). They also point out that the density of a magmatic liquid can change in a variety of ways as crystallization proceeds: changes in temperature induces crystallization and the selective removal of components from the melt into the solid phases can leave the melt either more or less dense, depending on the phases involved (Fig. 5); the fluid structures attendant with crystallization were also explored by (McBirney et al., 1985).

EXTENSIVE VARIABLES

Size and shape of plutons

From field relationships alone, the size and shape of a pluton is difficult to confidently determine. This has motivated the use of geophysical techniques, principally gravity, to delineate the subsurface continuity of plutonic exposures, aspect ratio, and depth to basement below plutons. Spatial variations in density yield gravity anomalies which can indicate geochemical zonation at depth, hidden apophyses and roots, or the existence of plutons where there is no outcrop or surface expression of metamorphism. Despite the ambiguities inherent in the interpretation of gravity measurements, it is currently the best geophysical technique for delineating plutons in the subsurface (Bott and Smithson, 1967; Vigneresse, 1990).

Plutons are typically elliptical to sub-circular in map view with near vertical contacts. This is in qualitative agreement with the forms suggested by fluid dynamic laboratory experiments of diapiric instabilities. These experiments in turn have guided the interpretation of the gravity data and the agreement of measured gravity anomalies with with those predicted from simple geometries is often surprisingly good. Silicic plutons are typically associated with negative anomalies on the order of tens to hundreds of mgals (1 gal = 1 cm s^{-2}), although positive anomalies over plutonic outcrops may reveal mafic components of a plutonic suite at depth (Waskom and Butler, 1971). Variations in gravity can also be used to determine the geochemical structure of dominantly mafic intrusions where the anomaly is positive (Gupta and Sutcliffe, 1990; Stephenson and Thomas, 1979). One of the largest sources of uncertainty in constructing gravity models of plutons is the possibility of lateral and vertical variations in density. Almost all plutonic suites demonstrate mineralogical zoning of some kind, and this will introduce three dimensional effects whose importance is difficult to determine *a priori* (Bott and Smithson, 1967; Vigneresse, 1990). These variations have been documented in a stock in the Sierra Nevada (Peikert, 1962) where an algorithm for vertical extrapolation of density data was presented. If the boundaries of a pluton are gradational and/or have similar densities to the country rock, the pluton size as determined by the gravity measurements will be difficult to resolve and will probably represent minimum values.

The estimated thickness and morphology of a pluton seems to vary with tectonic regime (Bott and Smithson, 1967; Hodge et al., 1982; Vigneresse, 1990) although there are a number of assumptions implicit in this. Local structure and state of stress influence the final form of a pluton, and some generic elements emerge: most plutons have thicknesses that vary from five to thirteen km. Examples include the Aulneau batholith (Brisbin and Green, 1980) which has a range from 4.5-7 km, the Tuolumne Intrusive Suite of the Sierra Nevada which has an estimated thickness of up to 9 km (Oliver, 1977), the Round Lake batholith with a maximum thickness of 10 km (Gibb and van Boeckel, 1970), the Ardura Granite with a thickness of 5 km (Young, 1974) the Nigerian Younger Granite province with depths estimated of up to 12 km (Ajakaiye, 1970), plutons of the English Lake District which extend to 9 km (Lee, 1986; Locke and Brown, 1978), the Mortange pluton has depths of from 5-10 km (Guineberteau et al., 1987), and some plutons in New England which have an estimated thickness of about 7.5 km (Hodge et al., 1982). Thinner plutons are also reported: the Lilesville Granite (Waskom and Butler, 1971) the Katahdin pluton (Hodge et al., 1982) and the Flamanville granite (Brun et al., 1990) are modeled as having a depth of about 3 km. Gravity data can also indicate buried or hidden plutons. One that is thought to be 13 km thick has been described near Lumberton, North Carolina (Pratt et al., 1985) and it is thought that a pluton is responsible for the negative gravity anomaly in the Southern North Sea Basin (Donato and Megson, 1990).

The gradient of the gravity anomaly and the value of gravity at the contact can be used to determine if the contact slopes inward or outward and the shape (Bott and Smithson, 1967). Two dimensional modeling of a number of plutons indicates that near vertical, outward dipping contacts are most common, although inward dips have been confirmed. In addition, the method of Bott and Smithson (1967) can also be used to find the contacts if they are not exposed. A three-dimensional shape of a circular or ovoid cylinder or funnel seems to be the most common, with an aspect ratio (diameter divided by depth) that varies from around one for anorogenic plutons to ten for synkinematically emplaced plutons (Vigneresse, 1988). Simple cylindrical or funnel shapes usually give an adequate fit to the gravity data and agrees with the typically limited field information (Courrioux, 1987; Hill, 1988). However, there are also numerous examples of pre-existing structure controlling the shape and disposition of plutons (Pitcher and Bussell, 1977) and producing bodies that are elongate in map view, such as in units I and II of the San Jacinto intrusive complex (Hill, 1988).

The geophysical imaging of volcanic systems may provide some insights into the size and shape of plutons. The contrast in geophysical properties of melts and crystal-melt mushes relative to country rock permits additional geophysical techniques, particularily seismology, to be used to delineate magma bodies in the subsurface. A number of tomographic and gravity studies of active systems have been done (Aucher et al., 1988; Carle, 1988; Dawson et al., 1990; Evans and Zucca, 1988; Iyer et al., 1990; Thorpe et al., 1981) and suggest that the magma chambers under Cascade volcanos to be small: of the order of a few cubic kilometers. A larger thermal anomaly is present under the Long Valley-Mono Craters region of California with a volume of a few hundred cubic kilometers and a vertical extent of up to 20 km. Note however that this anomaly may only represent a few percent melt rather than a contiguous "fluid" chamber (Hill and Bailey, 1990).

There are a number of caveats that need to be invoked when using gravity measurements to determine pluton size and morphology. First, the interpretation of the data is very model-dependent and extrapolating pluton geometry from tectonic regime seems premature given our lack of understanding of pluton morphologies. Second, the shape and volume of the pluton will change as intrusion and magmatism proceed and it is difficult to know what the pluton geometry was at the time when maximum temperatures were recorded in the contact aureole. Finally, if the pluton and contact aureole have undergone any faulting, the reconstruction to a palinspastic base must be done to recover any pluton geometry from the geophysical data.

Open-system behavior

There is little question that magmatic complexes act as open systems at a variety of scales (Hildreth et al., 1986). This means more than just the inevitable loss of heat and volatiles during ascent, cooling and crystallization; open system also refers to the re-intrusion of a single magma chamber by additional magma. This concept of a single open system can be further extended to include magmatic complexes that are amalgamated by repeated intrusion of discrete magma bodies that coexist spatially and temporally, with or without mixing, and yield zoned plutonic suites upon cooling. Both of these types of open system behavior can induce a time dependence in the contact metamorphism at local and regional scales (Barton and Hanson, 1989). Zoned plutonic suites have been described in a number of studies which are considered below; a recent compendium and discussion of zoned plutons in Japan and Southern and Baja California, edited by Shimizu and Gastil (1990), addresses a number of issues related to the origin of chemical and structural variability in plutonic complexes.

One very common characteristic of plutons is the presence of zoning. This zoning is manifested in a number of forms; most importantly in structural zoning and compositional zoning of the major elements, isotopes and trace elements. Practically every imaginable variety of zoning has been documented. The most common is mafic margins to felsic cores (normal zoning) as documented in the Criffell pluton (Stephens and Hallidy, 1980) where the silica content ranges from 64.7% to 72.0% and in most of the plutonic suites of the Sierra Nevada (Bateman et al., 1963). Reverse zoning, with felsic margins to mafic cores, occurs less frequently; examples include the Bottle Lake Complex in Maine (Ayuso, 1984) and the reversely zoned Grizzly Peak cauldron which has a variation from granodiorite to quartz monzonite (Fridrich and Mahood, 1984). Striking vertical zonation occurs as well. It has been demonstrated in the Slinkard pluton and Wolley Creek batholith by Barnes et al. (1990, 1986), where a vertical two tiered dacite over andesite magma chamber is postulated. Vertical zoning is also documented in the Chemehuevi Mountains plutonic suite (John, 1988), where the zoning is defined by a transition from granodiorite at the walls and floor to granite in the interior. This type of zonation has also been mapped in the Cañas and Puscao plutons, Peru (Taylor, 1985). A similar progression is discussed by Hopson et al. (1987) in a remarkable study that documents a continuum of composition and fabric in what is estimated to be up to 17(±5) km of vertical exposure; the pluton displays both horizontal and vertical zoning of diorite to leucogranite. An additional type of zoning is noted by the distribution of mafic enclaves which are most often found at the margins with a decreasing abundance inward. Thus the presence of zoning in what is ostensibly a comagmatic suite provides clues for the physical processes by which plutonic complexes assemble themselves.

Cryptic zoning, as revealed in the isotopic systematics, can be present even with homogeneity in the major elements and mineral chemistry (Roddick and Compston, 1977). One striking example of this is in the San Jacinto complex (Hill and Silver, 1988; Hill et al., 1985) where the variation of the initial strontium ratio does not correlate with any petrological or geochemical parameter other than position. The unzoned Red Lake pluton in the Sierra Nevada also manifests heterogeneity in Sr isotope compositions on the whole rock scale (Hill et al., 1988). I suspect that this kind of zoning is probably very common, the fact that it is not reported more often is more a measure of the expense and difficulty of analyzing an appropriate number of samples that demonstrate the heterogeneity. One of the more exotic manifestations of heterogeneity is the occurrence of the so called ghost stratigraphy where variations in geochemical indexes of the country rock retain continuity along strike, across the contact and into the pluton. The implication for coupled and uncoupled heterogeneity at a variety of scales have been considered by Miller et al. (1988), who explore the length scales implicit in a number of transport processes. Retention of spatial heterogeneity rules out scenarios that call for continuous and vigorous stirring of the magma by some unspecified convective process. In my view, the simplest interpretation is that there is little mixing among the magmatic parcels that assemble themselves into a pluton.

The geochemical zoning described above can be spatially discontinuous with abrupt step changes, or continuous where the variations are gradual. Typically, most of the variation within a mappable unit occurs near the margins and becomes less pronounced inward; sometimes giving way abruptly to a compositionally distinct zone which itself may have gradual changes in geochemical and mineralogical properties. Some plutonic suites will display both discontinuous and gradual zoning between two members, the style varying at different locations along the contact, where relatively smooth changes are abruptly terminated at what appear to be an inner intrusive contact or delimited by zones of material that appears to be mixed (Stephens and Hallidy, 1980). Both styles of internal variation, sharp and gradual, are observed in the well-studied Tuolumne Intrusive Suite (Bateman and Chappell, 1979; Kistler et al., 1986).

The mechanisms proposed to account for the origin of the zoning are usually considered as some combination of two processes: re-intrusion of the system by pulses that may or may not be genetically related, or an in situ process of fractionation. The abrupt changes in mineralogy or other geochemical index with sharp contacts suggests repeated intrusion of chemically and/or rheologically distinct magmatic parcels, sometimes referred to as "pulsing". The parcels combine to form a composite magmatic complex whose constituent pulses need not be related through any simple geochemical scheme. Isotope geochemistry has been particularily useful in demonstrating that the magmas that form a plutonic suite need not be genetically related, despite the fact that major element fractionation schemes provide a permissive interpretation for some kind of genetic link among the pulses. Examples where spatially adjacent pulses are demonstrably not related through a simple closed system separation of phases are the Tuolumne Intrusive Suite (Kistler et al., 1986), the Criffell pluton (Stephens and Hallidy, 1980) and the Zaer pluton (Bouchez and Diot, 1990). The origin of reverse zoning is discussed by Fridrich and Mahood (1984) who propose that the resurgence of a vertically zoned magma chamber can account for reverse zoning. This is consistent with the documented occurrence of reversely zoned intrusions in one of three environments (Fridrich and Mahood, 1984): (1) deeply eroded cauldrons; (2) small, high level bodies; (3) cupolas of a batholith. In situ processes such as sidewall crystallization, as envisioned and modeled by McBirney et al. (1985) and invoked by Sawka (1990) in which a spatially static chamber unmixes by virtue of compositional convection during solidification, yielding a stratified system, have been invoked to account for the generation of zoning. Such models are difficult to constrain. The most buoyant melt is the most viscous and exists in the part of the chamber where the permeability is the lowest. Fluid dynamic modeling of compositional convection with viscous fluids casts doubt that this process can generate the kinds of mass flux required to produce the observed variations (Tait and Jaupart, 1989) in a fixed chamber. One exception might be the volatile rich portion near the top of the magma chamber, but this is not likely to produce the observed chamber-wide zonation. Numerical models of this process that incorporate the strong variations in viscosity coupled to solidification are currently being developed.

The study of Hopson et al. (1987) provides evidence that fractionation and stratification of a mechanically continuous magmatic mass from mafic margins to a more felsic roof and core are possible by virtue of ascent. It is envisioned that in the process of intrusion the near-liquidus phases would be preferentially accreted at the margins and floor, which avoids the problem of having to unmix a homogeneous magma in situ to create the zoning. Local solid-melt segregation modifies the proportions of crystals and is invoked to account for spatial heterogeneity in the major and trace elements in the macroscopically unzoned Meatiq Dome (Sultan et al., 1986). Speer et al. (1989) also invoke solid-melt separation during intrusion as a means of producing local compositional variations. The implication of this is that no one bulk sample of the pluton may ever have represented what was once a liquid.

The origin of the gradients in structure and fabric from the bottom and sides into the interior is consistent with such a picture. This may simply represent the progression from near solidus conditions at the margins where the crystallinity would be high enough such that the magma would behave in a plastic fashion and retain a fabric that would be induced by differential flow in the interior. I do not mean to imply vigorous convection (see below) but rather the auto-intrusion of the magmatic system up through itself. Buoyant upward movement of a magma is most easily accomplished by *internal* slip, thus "telescoping" out of it's lower portions and "flattening" and spreading laterally at the roof. Such a process may produce the deformation that has been dubbed "ballooning" by Sylvester et al. (1978) at the Papoose Flat pluton, by Ramasy (1989) at the Chindamora batholith, in the Cannibal Creek pluton (Bateman, 1985a,b) and in the Flamanville granite (Brun et al., 1990). Patterson (1988) argues persuasively that many of the fabrics ascribed to ballooning are consistent with a model of plutons as piercement diapirs.

In addition to geochemical variations, plutons can be structurally zoned as well, the outer portions often having a pronounced contact-parallel foliation which changes into a fabric free interior (Bateman, 1985b; Compton, 1955; Courrioux, 1987; Ramsay, 1989). In addition, eye-catching features such as schleiren and unusual megacrystic "dikes" and apophyses testify to some kind of fluid motion and subsequent sedimentary style of crystal organization. Elliptical enclaves are often present but determining the strain from these is difficult as one must make assumptions regarding their initial shape; the same problem exists in trying to extract strain information from xenoliths. The most comprehensive discussion to date on the origin and significance of magmatic fabrics can be found in Paterson et al. (1989). The presence of aligned, euhedral minerals, particularily parallel to contacts, is evidence for the origin of foliation by magmatic flow. Conversely, foliations at high angles to the contact and/or continuous with regional cleavage suggest an origin by tectonic processes unrelated to pluton emplacement or internal flow. A provocative study of the internal structures in the South Mountain batholith was done by Abbott (1989) who demonstrated that many of the flow features were folds and represent a laminar flow regime. The "folds" formed during ascent and lateral injection and thus the fluid structures represent the ascent and assembly process rather than the traces of sustained and vigorous convection. The anisotrophy of magnetic susceptibility measurements has also been used to define magmatic flow fabric (Bouchez and Diot, 1990) where mineral fabrics are not present. The classic text by Balk on igneous structures (1937) and the text by Marre (1986) provide good introductions to the practical aspects and terminology in the structural characterization of magmatic fabrics.

One manifestation of open system behavior in volcanic and granitic rocks is the presence of mafic enclaves or inclusions. Although there is no question that magma mingling occurs, as demonstrated in volcanic rocks, the importance of magma mixing on a *chamber wide* scale at epizonal to mesozonal depths is not clear. Geochemistry and dis-equilibrium textures are typically used as evidence for their origin (Hibbard, 1981; Vernon, 1990) although Christiansen and Venchiarutti (1990) caution against using the compositions of mafic enclaves to constrain mixing end members or enclave origin. The presence of enclaves in plutons can been ascribed to three end member processes: (a) the intrusion of mafic magmas, typically high-alumina basalt, into a silicic chamber where the basic magma may be quenched into pillows and subsequently disaggregated and dispersed throughout the chamber (Dorias et al., 1990; Hill, 1988; Holden et al., 1987; Larsen and Smith, 1990; Reid et al., 1983); Frost and Mahood (1987) discuss the kinematic limitations on the ability of magmas to become hybridized; (b) the bringing together of cumulates from the early, near-liquidus crystallization of the pluton forming basic clots (Dodge and Kistler, 1990) and thus the enclaves are interpreted to be "autoliths" in the sense proposed in the seminal work of Pabst (1928); (c) the enclaves are resititic, being the refractory residues of the incompletely fused metaigneous protolith (Chappell and Stephens, 1988; Chappell et al., 1987; Chen et al., 1990; Chen and Williams, 1990). All of these hypotheses have merit and a given pluton may have enclaves with more than one origin (Didier, 1973, 1987).

Whichever interpretation for the origin of enclaves is appropriate could have a large impact on the contact metamorphism. If the enclaves originate by the intrusion of mafic magmas into the chamber, then there is an additional contribution to the enthalpy content of the chamber and the potential for convection to be initiated as a result of the interaction with the hotter mafic magma. An additional issue concerns the timing of the intrusion of the mafic component. Did it occur at the time of intrusion and magma chamber formation, as pictured by Hill (1988)? Does it occur at the time of magma genesis during partial melting in the deep crust, as a corollary to the MASH (melting-assimilation-storage-homogenization) hypothesis of Hildreth and Moorbath (1988)? Or does it occur along the margins during ascent as pictured by Ayrton (1988)? If, however, the enclaves have a restitic or cumulate origin then they do not represent any substantial addition of enthalpy into the chamber, and will also not be involved in driving convection. It is important then to try to determine the additional contribution to the total enthalpy represented by the enclaves.

Assimilation of country rock represents another type of open system behavior. Chemical models of many, if not all, plutons require some assimilation to account for the isotopic diversity. The difficulty in evaluating the impact of assimilation on contact metamorphism is in knowing when and how the assimilation occurred. Assimilation can occur in a number of ways: the thermal and chemical digestion of stoped blocks, the remelting of material previously crystallized at the chamber margins, and the incorporation of country rock partial melt at the chamber walls (Mahood, 1990; McBirney et al., 1987). It can be difficult to identify assimilation if the material being incorporated in the magma has a similar chemical composition, Bacon et al. (1989) use oxygen isotope data to argue for 10-20% assimilation of granitoid country rock in a rhyodacitic chamber; Grunder (1987) uses similar evidence to evaluate the role of assimilation in the Calabozos magmas and argues for 5-30% incorporation of wall and roof rocks. Johnson (1989) and Grove et al. (1988) argue for a decoupling of assimilation and fractionation in space and time. These conclusions are based on the study of volcanic systems where it appears that assimilation occurs dominantly at the tops of the magmatic system. The evidence for assimilation in plutons is more difficult to directly recognize. Agué and Brimhall (1988a;b) use the regional variation of F/OH in biotite from Sierra Nevada granitoids to identify the presence of contamination, ostensibly assimilation, and the composition of the assimilant. One should use this approach cautiously as a complete thermodynamic characterization of biotite remains to be developed; the most recent contribution in this regard is the work of Zhu and Sverjensky (1991).

Perhaps one of the most difficult aspects in evaluating the implications of open-system behavior for contact metamorphism is determining when in the history of a magmatic complex open system behavior occurred and how this influences the heat transfer systematics driving contact metamorphism. Typically, thermal models invoke the instantaneous intrusion of a system that is closed to additional inputs and that cools conductively. Such models often agree reasonably well with the temperatures and extent of contact metamorphism, suggesting that natural systems are insensitive to the thermal activity attendant with open system behavior. This can be understood in light of two end member scenarios: (a) most open system behavior occurs at the time of melt generation and during ascent and magma chamber formation, and before thermal contact metamorphism, or (b) crystallization at the margins effectively insulates the margins from processes in the interior where temperatures are macroscopically buffered by phase changes.

DYNAMIC STATE: COMBINATION OF INTENSIVE AND EXTENSIVE VARIABLES

To fully understand the tempoial history of contact metamorphism requires more than assessing the thermodynamic state by quantifying the intensive and extensive variables. To elucidate the timing and nature of pluton solidification and volatile release requires that a continuum mechanical approach be invoked. This involves identifying the probable heat and mass transfer mechanisms in the magma, guessing the initial and boundary conditions and constructing an analog system, either numerically or in a laboratory setting. Even this doesn't guarantee that a unique and simple picture of the coupling between magmatism and contact metamorphism will emerge; a variety of processes can occur in the magmatic history of a pluton that are difficult to unambiguously identify given only the long-cooled pluton. In addition, constructing the analog system can be costly in terms of computational resources, and laboratory experiments rarely scale adequately to yield results with geologic verisimilitude; see the recent review on magmatic processes by Cashman and Bergantz (1991). It is somewhat reassuring that often the simplest models yield results that are in good agreement with the conditions of contact metamorphism. Nonetheless, the student of contact metamorphism should be aware of the parameterizations for the variety of fluid structures that are possible in magmas and also the variety of pitfalls that surround their application to real systems. As a complete treatment of this subject goes well beyond the scope of this paper, I would direct the reader to some texts that can provide an excellent introduction to the application of continuum principals for heat and mass transfer. For conduction, the classic treatment of Carslaw and Jaeger (1959) remains the standard reference, although I prefer the lucid and comprehensive study of diffusion problems by Ozisik (1980) who addresses the geologically important cases for solving nonlinear equations, composite systems, and phase change problems. An excellent introductory text on natural convection is that of Bejan (1984), a more advanced but very comprehensive work is that of Gebhart et al. (1988). Many of the principals of continuum mechanics are applied to a variety of geological problems by Turcotte and Schubert (1982).

Crystallization

Much of the physical and chemical history of a magma is controlled by the process of solidification/melting. Solidification will partition the magma body into a system with a growing crystal rind and with crystals growing in the interior. How the magma body decides to partition the crystallization process between the margins and interior is not well understood, but whatever the reason it has a profound effect on the ability of the magma to convect and can thermally buffer the contact region from processes occurring inside. Many of the generic elements of magmatic solidification are treated in Marsh (1989a) and Bergantz (1990). Only the most generic and qualitative aspects will be discussed here.

As crystallization proceeds from the margins inward, there will be a transition from cooled material at or below solidus temperatures, into a crystal-liquid mixture above solidus temperatures. This transition is crystallinity is also a transition in rheology, the regions closer to the contact having a higher crystallinity and hence a higher viscosity. Conversely, those regions closer to the center of the chamber will have a higher liquid fraction and a lower viscosity. This suggests that it may be useful to consider in more detail the partitioning of the crystal-melt zone by virtue of these changes in rheology. Figure 6 is a schematic of this transition zone. Based on the work of Marsh (1989a, 1981), (also see discussion in Bergantz (1990)) a 50% crystallinity is chosen as the rheological threshold between material that is effectively solid from that which can have a fluid like behavior. The mush region is where there is a crystallinity between about 25% to 50%; fluid like behavior and porous media flow may occur but strain rates will be much lower that the adjacent suspension region where the crystallinity is less that about 25% and the viscosity only about an order of magnitude greater than that of the free fluid. This partitioning effectively redefines what one might consider the mechanical boundary of the chamber, and also removes that portion of the phase diagram that lies above the 50% crystallinity to participate in driving petrologic diversity. This region is important though as that is where

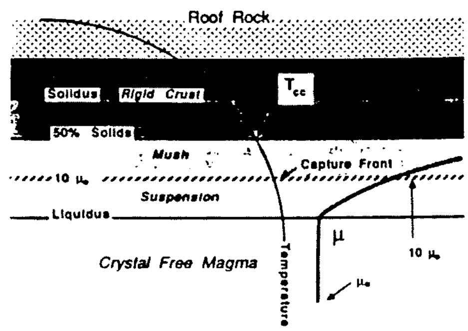

Figure 6. Schematic of solidifying margins of a magma chamber. From Marsh (1989).

volatile saturation may occur which may yield fracturing of the solidified rind in the manner proposed by Burnham (1979a,b). Fracturing may also occur due to thermoelastic stresses, perhaps allowing hydrothermal fluids to move inward (Lister, 1974; Carrigan, 1986), although this would require complete saturation in the country rock, a condition that is probably rarely achieved in nature.

The enthalpy changes that occur during solidification will buffer the heat transfer systematics at the chamber margins, hence any model of magmatic cooling must explicitly account for this. This can be difficult to predict a priori, as discussed in Bergantz (1990) who also discusses practical alternatives to needing to know the amount and composition of every phase and some of the details of the mathematical treatment. One important implication for magmatic processes is the role of crystallization in dynamics. There is a growing appreciation that the application of simple fluid dynamical models to variable viscosity, crystallizing systems requires extreme caution as the simple parameterizations of classic fluid mechanics may have little relation to magmatic systems.

Conduction and convection models

It is not difficult to imagine two end member heat transfer scenarios in a magma chamber: a magma that cools and solidifies by pure conduction, and one that is well mixed such that the internal temperature is everywhere the same and decreases monotonically with time. No doubt magmatic systems lie between these extremes, fluid motion is almost unavoidable given the complex ways in which a buoyancy is generated in a crystallizing magma. The question then is what are the possible fluid structures and how might they influence the heat and mass transfer systematics.

The simplest cooling model for the coupled system pluton-wallrock is conductive cooling of a magma body that is closed to additional inputs. The simplicity lies in that the equations describing the heat transfer are often linear and the usual methods of linear mathematics can be applied directly to their solution. The classic and comprehensive studies that invoke conduction include a number of works by Jaeger (1957, 1968), (only two of which are cited here), and that of Irvine (1970). The requirement of linearity and other simplifying assumptions restricts some of the applications of these models to real systems; enthalpy changes with temperature are typically nonlinear in both the country rock and the magma chamber. Using experimentally determined, nonlinear enthalpy-temperature relations, Bergantz (1989) evaluated the melting history and contact metamorphism associated with basaltic intrusion. Bowers et al. (1990) use a numerical approach to model the cooling history of the Cupsuptic aureole in two and three dimensions; their work is among the first to incorporate geologically meaningful enthalpy-temperature relationships and pluton geometries for the cooling of a granitoid. Their work has additional merit in that the Cupsuptic aureole has been well studied, hence comparison can be made between the model predictions and the actual conditions of contact metamorphism. By comparing the actual position of metamorphic isograds in the aureole with those calculated from the model, they find that the assumption of instantaneous intrusion followed by *conductive* cooling in the magma gives the best fit. Another magmatic system that had a well documented conductive cooling history is that of the Hawaiian lava lakes, a system that is different from the Cupsuptic pluton in practically every respect (Marsh, 1988). This would imply that convection is buffered by the process of solidification, almost regardless of the composition or initial conditions *in closed systems*. The degree of partial melting estimated in some contact aureoles is much like that predicted by Bergantz (1989) who addressed a coupled, conductive model. Clearly, more modeling of the kind done by Bowers et al. (1990) needs to be done in other well studied conjugate magma-country rock systems, particularily where reintrusion and open system behavior can be demonstrated and constrained.

Convection in magmas is transitory and the fluid structures that are suggested by mineral alignment in plutons are not a simple snapshot of a flow field at any given time. It is difficulty to directly identify the evidence of convection based on measurable quantities in the pluton alone, hence it is not surprizing that the role of magmatic convection in the cooling of plutons is a matter of some uncertainty and even controversy. This uncertainty has motivated many geologists to invoke convective schemes that borrowed from the fluid dynamics and engineering literature and that often have little in common with the time dependent and potentially complex styles of fluid motion that can originate in variable viscosity, silicate liquids where crystallization is occurring at the margins and in the interior. Given the seeming confusion over the often invoked parameters such as the Rayleigh number, it is appropriate to briefly review (very briefly) some of the first order elements in simple parameterizations of convective phenomena. For a general exposition on convection, the reader is directed to texts such as Bejan (1984) and Turcotte and Schubert (1982); papers with more immediate applications to magmatic systems are those of Marsh (1988, 1989a,b) and Brandies and Marsh (1989, 1990).

When considering magmatic convection, it is common practice for geologists to invoke the dimensionless group known as the Rayleigh number. The form of the Rayleigh number is:

$$Ra = \frac{\rho g \beta \Delta T L^3}{\alpha \eta} \tag{3}$$

where ρ is density, g the scalar acceleration of gravity, β the thermal expansion coefficient, ΔT, the temperature which exists over a vertical length L, α is the thermal diffusivity and η the dynamic viscosity. One often hears that it is the ratio of the buoyant forces to the viscous forces but that is overly simplified and potentially misleading, see the discussion in Bejan (1984, p. 116-124). This dimensionless group is usually invoked in two contexts: in evaluating the conditions at the onset of convection where the notion of a critical Rayleigh number is invoked, and secondly, in the parameterization of heat transfer in the post-critical or finite amplitude regime. It is thus important to understand the fundamental differences, applications, and limitations of these two ways of parameterizing convection.

To use a concept like the critical Rayleigh number to address the onset of convection, it is important to fully understand the origin of this critical number. The critical Rayleigh number originates in a system where the initial condition is that of a highly simplified fluid layer that is in a quiescent state. This state is infinitesimally perturbed and a stability analysis is done on the perturbed state (see example in Turcotte and Schubert (1982, p. 274-279). The basic question is: will the perturbed system continue to dynamically evolve or will it return to the previous, unperturbed state? The stability analysis yields a critical Rayleigh number which provides a numerical index for determining when the initially quiet system is at the very brink of becoming unstable. As expected, different types of thermal and mechanical boundary conditions, porous media, etc., will yield different values of the critical Rayleigh number. The physical correctness of the theory and application of stability analysis has been verified by decades of numerical and laboratory experiments. The issue surrounding the application to geological systems is: do geological systems resemble those for which a critical Rayleigh number is appropriate? One must be very careful not to apply a critical Rayleigh number to a system with substantially different initial and boundary conditions than those for which the Rayleigh number was derived. For example, in the classic case of convection heated from below and cooled from above with an initially linear temperature variation, known generally as Rayleigh-Benard convection, a critical Rayleigh number can be found whose actual value depends on the form of the mechanical boundary conditions. The length scale appearing in the critical Rayleigh number is the thickness of the layer itself as that is the distance across which the temperature difference driving convection exists. Now the question is, are magma chambers that are being cooled from all sides and that do not have a linear initial temperature difference, the kinds of systems to which the critical Rayleigh number can be applied? The answer is almost always NO, and there is no question that the thickness of the pluton is not, in any geologically meaningful way, related to the onset of convection. To use an expression such as that given above in (3) to estimate the vigor or presence of convection in most magmas is a total misapplication of fluid dynamic principals. As magma chambers are not heated from below in any *regular* way, it appears that in almost all cases the principals of near-critical Rayleigh-Bernard convection have little application to magmatic processes. As discussed above in the section on solidification, magmas are special in that their history is largely buffered by phase changes, and also subject to reintrusion. Thus application of a critical Rayleigh number with a length scale of the chamber thickness is incorrect. Sadly, it is not uncommon in the literature for a petrologist to calculate a critical Rayleigh number, and conclude that the chamber must have been convecting madly, despite lots of field and chemical evidence to the contrary. This is not to say that fluid motion doesn't exist, but rather to affirm that it is highly time dependent (Brandeis and Marsh, 1990) and inexorably linked to the progress of solidification.

What then, are the kinds of fluid instabilities that might exist in vertical layers and how might one parameterize them? Another type of general convective phenomena is known as Rayleigh-Taylor convection. This kind of instability is that of a more dense layer over a less dense layer (rather than a linear variation in density as in Rayleigh-Benard), and it gives rise to "diapiric" type of instabilities. The general form of convection is that of plumes. One notable difference between the Rayleigh-Taylor instabilities and the Rayleigh-Bernard convection is that in the Rayleigh-Taylor case, the buoyant material occupies a discrete layer at the margins. This has direct intuitive appeal for application to magmas where the buoyant material occupies a region near the boundaries where cooling and crystallization are taking place. As the unstable material at the margins forms a "plume" and becomes unstable, it travels through the more dense melt and the return flow is not well organized; and the system may go through a repose time before additional instabilities develop. It may be that the convection is largely driven by the density contrasts attendant with solidification, plumes may rise off the floor due to the removal of heavier phases into the solids as studied by Nielson and Incropera (1991). Many aspects of plume formation have been treated in some detail by Marsh (1988, 1989a,b) who concludes that although convection can occur, it originates as discrete pulses at the margins and it is generally weak, due to the increased thermal resistance associated with a the diminishing capacity of the country rock to carry heat away, and a growing crust at the margins. Even a perfectly well mixed magma will only cool twice as fast as one that is cooling conductively, so the added thermal resistance associated with phase changes and variable viscosity will diminish this number even more. This may be why the isograds calculated from purely conductive cooling agree so well with those observed in nature and why models that call for rapid, whole chamber chamber convection and cooling are inevitably at odds with textural and chemical data in plutons.

The previous discussion was largely for systems that are cooled from above and below. The kinds of fluid motion that can occur at the vertical margins is fundamentally different. There is never a critical Rayleigh number for horizontal gradients in density and the instability has a negligible start up time. The form of convection at the margins is that of boundary layer flows that resemble a snow avalanche. For example, consider a hot fluid next to a cold wall, the cooling will initiate down-flow everywhere along the wall. The down-going fluid will accumulate in the downstream direction and hence the boundary layer will be thicker there and the heat transfer along the margins thus decreases in the direction of flow. If crystallization is occurring at the margins, a porous matrix may occur and most of the marginal flow will be best modeled as flow in porous media. This reduces the mass flux considerably and provides a strong limitation on the efficiency of sidewall processes to drive petrologic diversity. As with convection of all kinds, sidewall processes have been incorrectly invoked to explain a variety of plutonic features that are inconsistent with marginal solidification, see Trial and Spera (1990) for a discussion of the limitations on mass flux rates from sidewall processes.

The recent numerical and laboratory modeling of magmatic processes by a number of groups, e.g., Marsh (1989b), Brandeis and Marsh (1989), Trial and Spera (1990), Bergantz (1991), Tait and Jaupart (1989), Bennon and Incropera (1987), and Nielson and Incropera (1991) reveal that magmatic convection is a complex process where buoyancy is generated in complex ways and invoking dimensionless numbers without careful consideration of the particularities of the system at hand provides meaningless results. The possibility of open system behavior exacerbates this uncertainty. Physically meaningful modeling of magmatic processes is one of the challenges of igneous petrology in the next decade; no one is served by the ad hoc application of convective schemes that do not explicitly include solidification and attempt to quantify the time and length scales of the models thought to best represent magmatic conditions.

In closing, it appears that the best way to proceed in assessing the role of magmatism for the thermal evolution of contact metamorphism would be to start with the simplest model: a solidifying magma body that is cooling internally by conduction. Although this certainly oversimplifies the processes that occurred in the pluton, it may well represent the most thermally realistic conditions of contact metamorphism. This leaves us with a number of questions and possible research directions that *are* accessible given our current understanding of contact metamorphism and technology: what is the effect on the thermal window of reintrusion? Will adding magma initiate a new pulse of contact metamorphism that might be discernable in the rock record? It would be very important to better constrain the timing of volatile release, this would require combining models such as those of Nekvasil (1988) with solidification algorithms such as those of Bergantz (1991) and also require that thermoelastic stresses are calculated. More work needs to be done in establishing the rheological and thermophysical character of crystal-melt systems to better determine the efficiency of melt extraction and internal differentiation. Further field work is also needed to understand the relationship between volcanic and plutonic systems and the plutonic expression of volatile driven processes.

REFERENCES

Abbott, R.N. (1989) Internal structures in part of the South Mountain batholith, Nova Scotia, Canada. Geol. Soc. Amer. Bull. 101, 1493-1506.

Ague, J.J. and Brimhall, G.H. (1988a) Magmatic arc asymmetry and distribution of anomalous plutonic belts in the batholiths of California: Effects of assimilation, crustal thickness, and depth of crystallization. Geol. Soc. Amer. Bull. 100, 912-927.

Ague, J.J. and Brimhall, G.H. (1988b) Regional variations in bulk chemistry, mineralogy, and the compositions of mafic and accessory minerals in the batholiths of California. Geol. Soc. Amer. Bull. 100, 891-911.

Ajakaiye, D.E. (1970) Gravity measurements over the Nigerian Younger granite province. Nature 225, 50-52.

Anderson, A.T., Newman, S., Williams, S.N., Druitt, T.H., Skirius, C. and Stolper, E. (1989) H_2O, CO_2, Cl and gas in Plinian and ash-flow Bishop rhyolite. Geology 17, 221-225.

Anderson, J.L. (1990) The Nature and Origin of Cordilleran Magmatism. Geol. Soc. Amer. Memoir 174, 414 p.

Atherton, M.P. and Tarney, J. (1979) Origin of Granite Batholiths, Geochemical Evidence. Shiva, Cheshire, U.K., 148 p.

Aucher, U., Evans, J.R. and Stauber, D.A. (1988) High-resolution seismic tomography of compressional wave velocity structure at Newberry Volcano, Oregon Cascade Range. J. Geophys. Res. 93, 10,135-10,147.

Ayrton, S. (1988) The zonation of granitic plutons: the "failed ring-dyke" hypothesis. Schweiz. Mineral. Petrogr. Mitt. 68, 1-19.

Ayuso, R.A. (1984) Field relations, crystallization, and petrology of reversely zoned granitic plutons in the Bottle Lake Complex, Maine. U.S.Geol. Surv. Prof. Paper 1320. 58 p.

Bacon, C.R., Adami, L.H. and Lanphere, M.A. (1989) Direct evidence for the origin of low-^{18}O silicic magmas: quenched samples of a magma chamber's partially-fused granitoid walls, Crater Lake, Oregon. Earth Planet. Sci. Letters 96, 199-208.

Bacon, C.R. and Druitt, T.H. (1988) Compositional evolution of the zoned calcalkaline magma chamber of Mount Mazma, Crater Lake, Oregon. Contrib. Mineral. Petrol. 98, 224-256.

Bailey, J.C. (1977) Fluorine in granitic rocks and melts: a review. Chem. Geology 19, 1-42.

Balk, R. (1937) Structural behavior of igneous rocks. Geol. Soc. Amer. Mem. 5, 177.

Barnes, C., Allen, C.M., Hoover, J.D. and Brigham, R.H. (1990) Magmatic components of a tilted plutonic system, Klamath Mountains, California. Geol. Soc. Amer. Memoir 174 331-345.

Barnes, C.G., Allen, C.M. and Saleeby, J.B. (1986) Open and closed system characteristics of a tilted plutonic system, Klamath Mountains, California. J. Geophys. Res. 91, 6073-6090.

Barton, M.D. and Hanson, R.B. (1989) Magmatism and the development of low-pressure metamorphic belts: Implications from the western United States and thermal modeling. Geol. Soc. Amer. Bull. 101, 1051-1065.

Bateman, P.C. and Chappell, B.C. (1979) Crystallization, fractionation, and solidification of the Tuolumne Intrusive Series, Yosemite National Park, California. Geol. Soc. Amer. Bull. 90, 465-482.

Bateman, P.C., Clark, L.D., Huber, N.K., Moore, J.G. and Rinehart, C.D. (1963) The Sierra Nevada batholith, a synthesis of recent work across the central part. U. S. Geol. Surv. Prof. Paper 414-D, 46 p.

Bateman, R. (1985a) Aureole deformation by flattening around a diapir during in situ ballooning: the Cannibal Creek Granite. J. Geol. 93, 293-310.

Bateman, R. (1985b) Progressive crystallization of a granitoid diapir and its relationship to stages of emplacement. J. Geol. 93, 645-662.

Beard, J.S. and Lofgren, G.E. (1991) dehydration melting and water-saturated melting of basaltic and andesitic greenstones and amphibolites at 1, 3, and 6.9 kb. J. Petrol. 32, 365-401.

Beckermann, C. and Viskanta, R. (1988) Natural convection solid/liquid phase change in porous media. Int. J. Heat Mass Trans. 31, 35-46.

Bejan, A. (1984) Convection Heat Transfer. John Wiley & Sons, New York, 477 p.

Bennon, W.D. and Incropera, F.P. (1987) A continuum model for momentum, heat and species transport in binary solid-liquid phase change systems II. Application to solidification in a rectangular cavity. Int. J. Heat Mass Trans. 30, 2171-2187.

Bergantz, G.W. (1989) Underplating and partial melting: Implications for melt generation and extraction. Science 245, 1093-1095.

Bergantz, G.W. (1990) Melt fraction diagrams: the link between chemical and transport models. In: J. Nicholls and J.K. Russell, ed., Modern Methods of Igneous Petrology: Understanding Magmatic Processes. Reviews in Mineralogy 24, 240-257.

Bergantz, G.W. (1991) Conjugate solidification and melting in multicomponent open and closed systems. Int. J. Heat Mass Trans. in press.

Bergantz, G.W. and Lowell, R.P. (1987) The role of conjugate convection in magmatic heat and mass transfer. In: D.E. Loper, ed., Structure and Dynamics of Partially Solidified Systems, NATO ASI Series, Series E, No. 125 367-382.

Blundy, J.D. and Holland, T.J.B. (1990) Calcic amphibole equilibra and a new amphibole-plagioclase geothermometer. Contrib. Mineral. Petrol. 104, 208-224.

Bohlen, S.R. and Lindsley, D.H. (1987) Thermometry and barometry of igneous and metamorphic rocks. Ann. Rev. Earth Planet. Sci. 15, 397-420.

Bott, M.H.P. and Smithson, S.B. (1967) Gravity investigations of subsurface shape and mass distributions of granite batholiths. Geol. Soc. Amer. Bull. 78, 859-878.

Bottinga, Y. and Weill, D.F. (1972) The viscosity of magmatic silicate liquids: A model for calculation. Amer. J. Sci. 272, 438-475.

Bouchez, J.-L. and Diot, H. (1990) Nested granites in question: Contrasted emplacement kinematics of independent magmas in the Zaer pluton, Morocco. Geology 18, 966-969.

Bowers, J.R., Kerrick, D.M. and Furlong, K.P. (1990) Conduction model for the thermal evolution of the Cupsuptic aureole, Maine. Amer. J. Sci. 290, 644-665.

Brandeis, G. and Marsh, B.D. (1989) The convective liquidus in a solidifying magma chamber: a fluid dynamic investigation. Nature 339, 613-616.

Brandeis, G. and Marsh, B.D. (1990) Transient magmatic convection prolonged by solidification. Geophys. Res. Letters 17, 1125-1128.

Brisbin, W.C. and Green, A.G. (1980) Gravity model of the Aulneau batholith, northwestern Ontario. Can. J. Earth Sci. 17, 968-977.

Brun, J.P., Gapais, D., Cogne, J.P., Ledru, P. and Vigneresse, J.L. (1990) The Flamanville Granite (Northwest France): an example of a syntectonically expanding pluton. Geol. J. 25, 271-286.

Burnham, C.W. (1979a) The importance of volatile constituents. In: H. S. Yoder, ed., The Evolution of Igneous Rocks, Fiftieth Anniversary Perspectives, 439-482.

Burnham, C.W. (1979b) Magmas and hydrothermal fluids. In: H.L. Barnes, ed., Geochemistry of Hydrothermal Ore Deposits, 2 edition, 71-136.

Carle, S.F. (1988) Three dimensional gravity modeling of the geologic structure of Long Valley caldera. J. Geophys. Res. 93, 13,237-13,250.

Carrigan, C.R. (1986) A two-phase hydrothermal cooling model for shallow intrusions. J. Volc. Geophys. Res. 28, 175-192.

Carslaw, H.S. and Jaeger, J.C. (1959) Conduction of Heat in Solids. Clarendon Press, Oxford, 510 p.

Cashman, K.V. and Bergantz, G.W. (1991) Magmatic processes, Rev. Geophys., U.S. National Report 1987-1990, Contributions in Volcanology, Geochemistry, and Petrology, 500-509.

Cashman, K.V. (1990) Textural constraints on the kinetics of crystallization of igneous rocks. In: J. Nicholls and J.K. Russell, ed., Modern Methods of Igneous Petrology: Understanding Magmatic Processes. Rev. Mineral. 259-314.

Chappell, B.W. and Stephens, W.E. (1988) Origin of infracrustal (I-type) granite magmas. Trans. Royal Soc. Edinburgh 79, 71-86.

Chappell, B.W., White, A.J.R. and Wyborn, D. (1987) The importance of residual source material (restite) in granite petrogenesis. J. Petrol. 28, 1,111-1,138.

Chen, Y.D., Price, R.C. and White, A.J.R. (1990) Mafic inclusions from the Glenbog and Bluegum granite suites, southeastern Australia. J. Geophys. Res. 95, 17,757-17,785.

Chen, Y.D. and Williams, I.S. (1990) Zircon inheritance in mafic inclusions from Bega batholith granites, southeastern Australia: an ion microprobe study. J. Geophys. Res. 95, 17,787-17,796.

Cheng, P. (1978) Heat transfer in geothermal systems. Adv. Heat Transfer 14, 1-105.

Cheng, P. and Verma, A.K. (1981) The effect of subcooled liquid on film boiling about a vertical heated surface in a porous medium. Int. J. Heat Mass Trans. 24, 1151-1160.

Christiansen, E.H. and Venchiarutti, D.A. (1990) Magmatic inclusions in rhyolites of the Spor Mountation Formation, western Utah: Limitations on compositional inferences from inclusions in granitic rocks. J. Geophys. Res. 95, 17,717-17,728.

Civetta, L., Galati, R. and Santacroce, R. (1991) Magma mixing and convective compositional layering within the Vesuvius magma chamber. Bull. Volc. 53, 287-300.

Clemens, J.D. (1984) Water contents of silicic to intermediate magmas. Lithos 17, 273-287.

Compton, R.R. (1955) Trondhjemite batholith near Bidwell Bar, California. Geol. Soc. Amer. Bull. 66, 9-44.

Courrioux, G. (1987) Oblique diapirism: the Criffel granodiorite/granite zoned pluton (southwest Scotland). J. Struct. Geol. 9, 313-330.

Dawes, R.L. and Evans, B.W. (1991) Mineralogy and geothermobarometry of magmatic epidote-bearing dikes, Front Range, Colorado. Geol. Soc. Amer. Bull. 103, 1017-1031.

Dawson, P.B., Evans, J.R. and Iyer, H.M. (1990) Teleseimic tomography of the compressional wave velocity structure beneath the Long Valley region, California. J. Geophys. Res. 95, 11,021-11,050.

Didier, J. (1973) Granites and Their Enclaves. Elsevier, Amsterdam, 393 p.

Didier, J. (1987) Contribution of enclave studies to the understanding of origin and evolution of granitic magmas. Geol. Rundschau 76, 41-50.

Dodge, F.C.W. and Kistler, R.W. (1990) Some additional observations on inclusions in the granitic rocks of the Sierra Nevada. J. Geophys. Res. 95, 17,841-17,848.

Donato, J.A. and Megson, J.B. (1990) A buried granite batholith beneath the East Midland Shelf of the southern North Sea Basin. J. Geol. 147, 133-140.
Dorias, M.J., Whitney, J.A. and Roden, M.F. (1990) Origin of mafic enclaves in the Dinkey Creek pluton, central Sierra Nevada batholith, California. J. Petrol. 1990, 853-881.
Dunbar, N.W., Hervig, R.L. and Kyle, P.R. (1989) Determination of pre-eruptive H_2O, F, and Cl contents of silicic magmas using melt inclusions: examples from Taupo volcanic center, New Zealand. Bull. Volc. 51, 177-184.
Elkins, L.T. and Grove, T.L. (1990) Ternary feldspar experiments and thermdynamic models. Amer. Mineral. 75, 544-559.
Ernst, W.G. (1988) Metamorphism and Crustal Evolution of the Western United States. Prentice Hall, Englewood Cliffs, NJ, 1152 p.
Evans, B.W. and Vance, J.A. (1987) Epidote phenocrysts in dacitic dikes, Boulder County, Colorado. Contrib. Mineral. Petrol. 96, 178-185.
Evans, J.R. and Zucca, J.J. (1988) Active high resolution seismic tomography of compressional wave velocity and attenuation structure at Medicine Lake Volcano, Northern California Cascade Range. J. Geophys. Res. 93, 15,016-15,036.
Fridrich, C.J. and Mahood, G.A. (1984) Reverse zoning in the resurgent intrusions of the Grizzly Peak cauldron, Sawatch Range, Colorado. Geol. Soc. Amer. Bull. 95, 779-787.
Frost, B.R. and Lindsley, D.H. (1991) Occurrence of iron-titanium oxides in igneous rocks. In: D. Lindsley, ed., Oxide Minerals: Petrologic and Magnetic Significance, Rev. Mineral. 25, 433-468.
Frost, T.P. and Mahood, G.A. (1987) Field, chemical, and physical constraints on mafic-felsic interaction in the Lamarck Granodiorite, Sierra Nevada, California. Geol. Soc. Amer. Bull. 99, 272-291.
Fuhrman, M.L. and Lindsley, D.H. (1988) Ternary feldspar modeling and thermometry. Amer. Mineral. 73, 201-215.
Gebhart, B., Jaluria, Y., Mahajan, R.L. and Sammakia, B. (1988) Buoyancy-Induced Flows and Transport. Hemisphere. Cambridge Univ. Press, Cambridge, 971 p.
Ghiorso, M.S. (1984) Activity/composition relations in the ternary feldspars. Contrib. Mineral. Petrol. 87, 282-296.
Ghiorso, M.S. (1985) Chemical mass transfer in magmatic systems I. Thermodynamic relations and numerical algorithms. Contrib. Mineral. Petrol. 90, 107-120.
Ghiorso, M.S. and Carmichael, I.S.E. (1985) Chemical mass transfer in magmatic processes II. Applications in equilibrium crystallization, fractionation and assimilation. Contrib. Mineral. Petrol. 90, 121-141.
Gibb, R.A. and van Boeckel, J. (1970) Three-dimensional interpretations of the Round Lake batholith, northeastern Ontario. Can. J. Earth Sci. 7, 156-.
Green, N.L. and Udansky, S. . (1986) Ternary feldspar mixing relations and feldspar thermobarometry. Amer. Mineral. 71, 1,100-1,108.
Grove, T.L., Kinzler, R.J., Baker, M.B., Donnelly-Nolan, J.M. and Lesher, C.E. (1988) Assimilation of granite by basaltic magma at Burnt Lava flow, Medicine Lake volcano, nothern California: Decoupling of heat and mass transfer. Contrib. Mineral. Petrol. 99, 320-343.
Grunder, A.L. (1987) Low $\delta^{18}O$ silicic volcanic rocks at the Calabozos caldera complex, southern Andes. Contrib. Mineral. Petrol. 95, 71-81.
Guineberteau, B., Bouchez, J. L. and Vigneresse, J.L. (1987) The Mortagne granite pluton (France) emplaced by pull-apart along a shear zone: Structural and gravimetric arguments and regional implication. Geol. Soc. Amer. Bull. 99, 763-770.
Gupta, V.K. and Sutcliffe, R.H. (1990) Mafic-ultramafic intrusives and their gravity field: Lac des Iles area, northern Canada. Geol. Soc. Amer. Bull. 102, 1471-1483.
Hammarstrom, J.M. and Zen, E-an (1986) Aluminum in hornblende: An empirical igneous geobarometer. Amer. Mineral. 71, 1297-1313.
Harmon, R.S. and Barreiro, B.A. (1984) Andean Magmatism, Chemical and Isotopic Constraints, Shiva, Cheshire, UK, 250 p.
Hervig, R.L., Dunbar, N., Westrich, H.R. and Kyle, P.R. (1989) Pre-eruptive water content of rhyolitic magmas as determined by ion microprobe analyses of melt inclusions in phenocrysts. J. Volc. Geophys. Res. 36, 293-302.
Hibbard, M.J. (1981) The magma mixing origin of mantled feldspars. Contrib. Mineral. Petrol. 76, 158-170.
Hildreth, W. (1987a) New perspectives on the eruption of 1912 in the Valley of Ten Thousand Smokes, Katmai National Park, Alaska. Bull. Volc. 49, 680-693.
Hildreth, W. (1987b) Review of "Magmatism at a Plate Edge". Tectonophysics 140, 335-339.
Hildreth, W., Grove, T. L. and Dungan, M. A. (1986) Introduction to the special section on open magmatic systems. J. Geophys. Res. 91, 5887-5889.

Hildreth, W. and Moorbath, S. (1988) Crustal contributions to arc magmatism in the Andes of Central Chile. Contrib. Mineral. Petrol. 98, 455-489.
Hill, D.P. and Bailey, R.A. (1990) The evolving image of a complex magmatic system beneath Long Valley caldera and the Mono-Inyo volcanic chain, eastern California. EOS, Trans. Amer. Geophys. Union 71, 961.
Hill, M., O'Neil, J.R., Noyes, H., Frey, F. and Wones, D.R. (1988) Sr, Nd and O isotope variations in compositionally zoned and unzoned plutons in the central Sierra Nevada batholith. Amer. J. Sci. 288-A, 213-241.
Hill, R.I. (1988) San Jacinto Intrusive Complex 1. Geology and mineral chemistry, and a model for intermittent recharge of tonalitic magma chambers. J. Geophys. Res. 93, 10,325-10,348.
Hill, R.I. and Silver, L.T. (1988) San Jacinto Intrusive Complex 3. Constraints on crustal magma chamber processes from strontium isotope heterogeneity. J. Geophys. Res. 93, 10,373-10,388.
Hill, R.I., Silver, L.T., Chappell, B.W. and Taylor, H.P. (1985) Solidification and recharge of SiO2-rich plutonic magma chambers. Nature 313, 643-646.
Hodge, D.S., Abbey, D.A., Harbin, M.A., Patterson, J.L., Ring, M.J. and Sweeny, J.F. (1982) Gravity studies of subsurface mass distributions of granitic rocks in Maine and New Hampshire. Amer. J. Sci. 282, 1289-1324.
Hodges, K.V. and McKenna, L.W. (1987) Realistic propagation of uncertainties in geologic thermo-barometry. Amer. Mineral. 72, 671-680.
Holden, P., Halliday, A.N. and Stephens, W.E. (1987) Neodymium and strontium isotope content of microdiorite enclaves points to mantle input to granitoid production. Nature 330, 53-56.
Hollister, L.S., Grissom, G.C., Peters, E.K., Stowell, H.H. and Sisson, V.B. (1987) Confirmation of the empirical correlation of Al in hornblende with pressure of solidification of calc-alkaline plutons. Amer. Mineral. 72, 231-239.
Holloway, J.R. (1976) Fluids in the evolution of granitic magmas: consequences of finite CO_2 solubility. Geol. Soc. Amer. Bull. 87, 1513-1518.
Hopson, C.A., Dellinger, D.A. and Mattinson, J.M. (1987) Crustal, mantle and hybrid components of a 4-dimensionally zoned pluton, North Cascades, Washington. EOS, Trans. Amer. Geophys. Union 68, no. 44, 1513.
Huang, W. and Wyllie, P.J. (1986) Phase relationships of gabbro-tonolite-granite-water at 15 kbar with applications to differentiation and anatexis. Amer. Mineral. 71, 301-316.
Hyndman, D.W. (1981) Controls on source and depth of emplacement of granitic magma. Geology 9, 244-249.
Irvine, T.N. (1970) Heat transfer during solidification of layered intrusions. I. sheets and sills. Can. J. Earth Sci. 7, 1031-1061.
Iyer, H.M., Evans, J.R., Dawson, P.B., Stauber, D.A. and Achauer, U. (1990) Differences in magma storage in different volcanic environments as revealed by seismic tomography: silicic volcanic centers and subduction-related volcanicoes. In: M.P. Ryan, ed., Magma Transport and Storage, 293-316.
Jaeger, J.C. (1957) The temperature in the neighborhood of a cooling intrusive sheet. Amer. J. Sci. 255, 306-318.
Jaeger, J.C. (1968) Cooling and solidification of igneous rocks. In: H. Hess, ed., Basalts, 503-536.
John, B.E. (1988) Structural reconstruction and zonation of a tilted mid-crustal magma chamber: The felsic Chemehuevi Mountains plutonic suite. Geology 16, 613-617.
Johnson, C.M. (1989) Isotopic zonations in silicic magma chambers. Geology 17, 1136-1139.
Johnson, J.W. and Norton, D. (1985) Theoretical prediction of hydrothermal conditions and chemical equilibra during skarn formation in porphyry copper systems. Econ. Geology 80, 1797-1823.
Johnson, M.C. and Rutherford, M.J. (1989a) Experimental calibration of the aluminum-in-hornblende geobarometer with application to Long Valley caldera (California) volcanic rocks. Geology 17, 837-841.
Johnson, M.C. and Rutherford, M.J. (1989b) Experimentally determined conditions in the Fish Canyon Tuff, Colorado, magma chamber. J. Petrol. 30, 711-737.
Kay, S.M. and Rapela, C.W. (1990) Plutonism from Antarctica to Alaska, Special Paper 241, Geol. Soc. Ameica. Geol. Soc. America, Boulder, CO, 263 p.
Kistler, R.W., Chappell, B.W., Peck, D.L. and Bateman, P.C. (1986) Isotopic variation in the Tuolumne Intrusive Suite, central Sierra Nevada, California. Contrib. Mineral. Petrol. 94, 205-220.
Lange, R.L. and Carmichael, I.S.E. (1990) Thermodynamic propreties of silicate liquids with emphasis on density, thermal expansion and compressibility. In: J. Nicholls and J.K. Russell, ed., Modern Methods of Igneous Petrol.: Understanding Magmatic Processes. Rev. Mineral. 25-64.
Larsen, L.L. and Smith, E.I. (1990) Mafic enclaves in the Wilson Ridge pluton, northwestern Arizona: Implications for the generation of a calc-alkaline intermediate pluton in an extensional environment. J. Geophys. Res. 95, 17,693-17,716.

Lee, M.K. (1986) A new gravity survey of the Lake District and three dimensional model of the granite batholith. J. Geol. 143, 425-435.

Lipman, P.W. (1988) Evolution of silicic magma in the upper crust: the mid-Tertiary Latir volcanic firld and its cogenetic granitic batholith, northern New Mexico, U.S.A. Trans. Royal Soc. Edinburgh 79, 265-288.

Lister, C.R.B. (1974) On the penetration of water into hot rock. Geophys. J. Royal Astron. Soc. 39, 465-509.

Locke, C.A. and Brown, G.C. (1978) Geophysical constraints on structure and emplacement of Shap granite. Nature 272, 526-528.

Luth, W.C. (1976) Granitic rocks. In: D.K. Bailey and R. MacDonald, ed., The Evolution of the Crystalline Rocks, 335-418.

Maaloe, S. and Wyllie, P.J. (1975) Water content of a granite magma deduced from the sequence of crystallization determined experimentally with water-undersaturated conditons. Contrib. Mineral. Petrol. 52, 175-191.

Mahood, G.A. (1990) Second reply to comment of R.S.J. Sparks, H.E. Huppert and C.J.N. Wilson on "Evidence for long residence times of rhyolitic magma in the Long Valley magmatic system: the isotopic record in the precaldera lavas of Glass Mountain". Earth Planet. Sci. Letters 99, 395-399.

Manning, D.A.C. (1981) The effect of fluorine on liquidus phase relationships in the system qz-ab-or with excess water at 1 kb. Contrib. Mineral. Petrol. 76, 206-215.

Marre, J. (1986) The Structural Analysis of Granitic Rocks. North Oxford Academic Publishers Ltd., London, 123 p.

Marsh, B.D. (1981) On the crystallinity, probability of occurrence, and rheology of lava and magma. Contrib. Mineral. Petrol. 78, 85-98.

Marsh, B.D. (1984) Mechanics and energetics of magma formation and ascent. In: F.R. Boyd, ed., Studies in Geophysics, Explosive Volcanism: Inception, Evolution and Hazards, 67-83.

Marsh, B.D. (1988) Crystal capture and retention in convecting magma. Geol. Soc. Amer. Bull. 100, 1720-1737.

Marsh, B.D. (1989a) Magma chambers. Ann. Rev. Earth Planet. Sciences 439-474.

Marsh, B.D. (1989b) On convective style and vigor in sheet-like magma bodies. J. Petrol. 30, 479-530.

McBirney, A.R., Baker, B.H. and Nilson, R.H. (1985) Liquid Fractionation. Part I: Basic principles and experimental simulations. J. Volc. Geophys. Res. 24, 1-24.

McBirney, A.R. and Murase, T. (1984) Rheological properties of magmas. Ann. Rev. Earth Planet. Sci. 12, 337-357.

McBirney, A.R., Taylor, H.P. and Armstrong, R.L. (1987) Paricutin re-examined: a classic example of crustal assimilation in calc-alkaline magma. Contrib. Mineral. Petrol. 95, 4-20.

McMillan, P.F. and Holloway, J.R. (1987) Water solubility in aluminosilicate melts. Contrib. Mineral. Petrol. 97, 320-332.

Mereu, R.F., Mueller, S. and Fountain, D.M. (1989) Properties and Processes of Earth's Lower Crust. 338 p.

Metzner, A.B. (1985) Rheology of suspensions in polymeric liquids. J. Rheology 29, 739-775.

Miller, C.F., Watson, E.B. and Harrison, T.M. (1988) Perspectives on the source, segregation and transport of granitoid magmas. Trans. Royal Soc. Edinburgh 79, 135-156.

Morse, S.A. (1988) Motion of crystals, solute, and heat in layered intrusions. Can. Mineralogist 26, 209-224.

Naney, M.T. (1983) Phase equilibria of rock forming ferromagnesian silicates in granitic systems. Amer. J. Sci. 283, 993-1033.

Nielson, D.G. and Incropera, F.P. (1991) Unidirectional solidification of a binary alloy and the effects of induced fluid motion. Int. J. Heat Mass Trans. 34, 1717-1732.

Nekvasil, H. (1988a) The calculated effect of anorthite on the crystallization paths of H_2O undersaturated haplogranitic melts. Amer. Mineral. 73, 966-981.

Nekvasil, H. (1988b) The calculation of equilibrium crystallization paths of compositionally simple hydrous felsic melts. Amer. Mineral. 73, 956-965.

Nielsen, R.L. (1990) Simulation of igneous differentiation processes. In: J. Nicholls and J.K. Russell, ed., Modern Methods of Igneous Petrology: Understanding Magmatic Processes. Rev. Mineral. 24, 65-106.

Norton, D. and Taylor, H.P.J. (1979) Quantitative simulation of the hydrothermal systems of crystallizing magmas on the basis of transport theory and oxygen isotope data: an analysis of the Skaergaard intrusion. J. Petrol. 20, 421-486.

Oldenburg, C.M. and Spera, F.J. (1991) Simulation of phase change and convection in magma bodies. Trans. ASME in press.

Oliver, H.W. (1977) Gravity and magnetic investigations of the Sierra Nevada batholith, California. Geol. Soc. Amer. Bull. 88, 445-461.

Ozisik, M.N. (1980) Heat Conduction. John Wiley & Sons, New York, 687 p.
Papst, A. (1928) Observations on inclusions in the granitic rocks of the Sierra Nevada. Univ. Calif. Publ. Bull. Dep. Geol. Sci. 17, 326-368.
Parmentier, E.M. (1979) Two-phase natural convection to a vertical heated surface in a permeable medium. Int. J. Heat Mass Trans. 22, 849-855.
Parmentier, E.M. and Schedl, A. (1981) Thermal aureoles of igneous intrusions: some possible indication of hydrothermal convective cooling. J. Geol. 89, 1-22.
Paterson, S.R. (1988) Cannibal Creek Granite: Post-tectonic balooning pluton or pre-tectonic piercement diapir? J. Geol. 96, 730-736.
Paterson, S.R., Vernon, R.H. and Tobisch, O.T. (1989) A review of the criteria for the identification of magmatic and tectonic foliations in granitoids. J. Struct. Geol. 11, 349-363.
Patiño-Douce, A.E. and Johnston, A.D. (1991) Phase equilibria and melt productivity in the pelitic system: implications fro the origin of peraluminous granitoids and aluminous granulites. Contrib. Mineral. Petrol. 107, 202-218.
Peikert, E.W. (1962) Three dimensional specific gravity variation in the Glen Alpine stock, Sierra Nevada, California. Geol. Soc. Amer. Bull. 73, 1437-1441.
Peterson, J.W. and Newton, R.C. (1990) Experimental biotite-quartz melting in the KMASH-CO_2 system and the role of CO_2 in the petrogenesis of granites and related rocks. Amer. Mineral. 75, 1029-1042.
Pichavant, M. (1987) Effects of B and H_2O on liquidus phase relations in the haplogranite system at 1 kbar. Amer. Mineral. 72, 1056-1070.
Pichavant, M. and Manning, D. (1984) Petrogenesis of tourmaline granites and topaz granites; the contribution of experimental data. Phys. Earth Planet. Interior 35, 31-50.
Pitcher, W.S. (1978) The anatomy of a batholith. J. Geol. Soc. London 135, 157-182.
Pitcher, W.S. (1979) The nature, ascent and emplacement of granitic magmas. J. Geol. Soc. London 136, 627-662.
Pitcher, W.S. (1987) Granites and yet more granites forty years on. Geol. Rundschau 76, 51-79.
Pitcher, W.S., Atherton, M.P., Cobbing, E.J. and Beckinsale, R.D. (1985) Magmatism at a Plate Edge, Halsted Press, 328 p.
Pitcher, W.S. and Bussell, M.A. (1977) Structural control of batholith emplacement in Peru: a review. J. Geol. Soc. London 133, 249-256.
Pratt, T.L., Costain, J.K., Coruh, C., Glover, L. and Robinson, E.S. (1985) Geophysical evidence for an allochthonous Alleghanian granitoid beneath the basement surface of the Coastal Plain near Lumberton, North Carolina. Geol. Soc. Amer. Bull. 96, 1070-1076.
Presnall, D.C. and Bateman, P.C. (1973) Fusion relationships in the $NaAlSi_3O_8$-$CaAl_2Si_2O_8$-SiO_2-H_2O and the generation of granitic magmas in the Sierra Nevada batholith. Geol. Soc. Amer. Bull. 84, 3,181-3,202.
Ramsay, J.G. (1989) Emplacement kinematics of a granite diapir: the Chindamora batholith, Zimbabwe. J. Struct. Geol. 11, 191-209.
Reid, J.B., Evans, O.C. and Fates, D.G. (1983) Magma mixing in granitic rocks of the central Sierra Nevada, California. Earth Planet. Sci. Letters 66, 243-261.
Roddick, J.A. (1983) Circum-Pacific Plutonic Terranes, Memoir 159, Geol. Soc. America. Geol. Soc. Amer., Boulder, CO, 316 p.
Roddick, J.C. and Compston, W. (1977) Strontium isotopic equilibration: a solution to a paradox. Earth Planet. Sci. Letters 34, 238-246.
Rushmer, T. (1991) Partial melting of two amphibolites: contrasting experimental results under fluid-absent conditions. Contrib. Mineral. Petrol. 107, 41-59.
Rutter, M.J., Van der Laan, S.R. and Wyllie, P.J. (1989) Experimental data for a proposed empirical igneous geobarometer: Aluminum in hornblende at 10 kbar pressure. Geology 17, 897-900.
Ryan, M.P. and Blevins, J.Y.K. (1987) The viscosity of synthetic and natural silicate melts and glasses at high temperatures and 1 bar (105 Pascals) pressure and at higher pressures. U.S. Geol. Surv. Bull. 1764 563.
Ryerson, F.J., Weed, H.C. and Piwinskii, A.J. (1988) Rheology of subliquidus magmas 1. Picritic compositions. J. Geophys. Res. 93, 3421-3436.
Sawka, W.N., Chappell, B.W. and Kistler, R.W. (1990) Granitoid compositional zoning by side-wall boundary layer differentiation: evidence from the Palisade Crest intrusive suite, central Sierra Nevada, California. J. Petrol. 31, 519-554.
Shimizu, M. and Gastil, G. (1990) Recent advances in concepts concerning zoned plutons in Japan and Southern and Baja California, University Museum, University of Tokyo, Nature and Culture, no.2, 244 p.
Silver, L.A., Ihinger, P.D. and Stolper, E. (1990) The influence of bulk composition on the speciation of water in silicate glasses. Contrib. Mineral. Petrol. 104, 142-162.

Sommer, M.A. and Schramm, L.S. (1983) An analysis of the water concentrations in silicate melt inclusions in quartz phenocrysts from the Bandelier Tuff, Jemez mountains New Mexico. Bull. Volc. 46, 312-317.

Speer, J.A., Naeem, A. and Almohandis, A.A. (1989) Small-scale variations and subtle zoning in granitoid plutons: the Liberty Hill pluton, South Carolina, U.S.A. Chemical Geol. 75, 153-181.

Stephens, W.E. and Hallidy, A.N. (1980) Discontinuties in the composition surface of a zoned pluton, Criffell, Scotland. Geol. Soc. Amer. Bull. 91, 165-170.

Stephenson, R. and Thomas, M.D. (1979) Three-dimensional gravity analysis of the Kiglapait layered intrusion, Labrador. Can. J. Earth Sci. 16, 24-37.

Stern, C.R. and Wyllie, P.J. (1981a) Phase relationships of I-type granite with H_2O to 35 kilobars: The Dinky Lakes biotite-granite from the Sierra Nevada Batholith. J. Geophys. Res. 86, 10,412-10,422.

Stern, C.R. and Wyllie, P.J. (1981b) Phase relationships of S-type granite with H_2O to 35 kilobars: Muscovite granite from Harney Peak, South Dakota. J. Geophys. Res. 86, 10,515-10,529.

Stolper, E., Fine, G., Johnson, T. and Newman, S. (1987) Solubility of carbon dioxide in albitic melt. Amer. Mineral. 72, 1071-1085.

Sultan, M., Batiza, R. and Sturchio, N.C. (1986) The origin of small-scale geochemical and mineralogic variations in a granite intrusion. Contrib. Mineral. Petrol. 93, 513-523.

Swanson, S.E., Naney, M.T., Westrich, H.R. and Eichelberger, J.C. (1989) Crystallization history of Obsidian Dome, Inyo Domes, California. Bull. Volc. 51, 161-176.

Sykes, M.L. and Holloway, J.L. (1987) Evolution of granitic magmas during ascent: A phase equilibrium model. In: B.O. Mysen, ed., Magmatic Processes: Physicochemical Principles. Special Publications of the Geochem. Soc. 1, 447-462.

Sylvester, A.G., Oertel, G., Nelson, C.A. and Christie, J.M. (1978) Papoose Flat Pluton: A granitic blister in the Inyo Mountains, California. Geol. Soc. Amer. Bull. 89, 1205-1219.

Tait, S. and Jaupart, C. (1989) Compositional convection in viscous melts. Nature 338, 571-574.

Taylor, W.P. (1985) Three-dimensional variation within granite plutons: a model for the crystallization of the Cañas and Puscao plutons. In: W.S. Pitcher, M.P. Atherton, E.J. Cobbibg and R.D. Beckinsale, ed., Magmatism at a Plate Edge, 228-234.

Thorpe, R.S., Locke, C.A., Brown, G.C., Francis, P.W. and Randal, M. (1981) Magma chamber beneath Poas volcano, Costa Rica. J. Geol.SL 138, 367-373.

Touloukian, Y.S., Judd, W.R. and Roy, R.F. (1981) Physical Properties of Rocks and Minerals, McGraw-Hill Book Co., New York.

Trial, A.F. and Spera, F.J. (1990) Mechanisms for the generation of compositional heterogenetics in magma chambers, Geol. Soc. Amer. Bull. 102, 353-367.

Turcotte, D.L. and Schubert, G. (1982) Geodynamics: Applications of Continuum Physics to Geologic Problems. John Wiley, New York, 450 p.

Tuttle, O F. and Bowen, N.L. (1958) Origin of granite in the light of experimental studies in the system $NaAlSi_3O_8$-$KAlSi_3O_8$-SiO_2-H_2O. Geol. Soc. Amer. Memoir 74. 156 p.

Vernon, R.H. (1990) Crystallization and hybridism in microgranitoid enclave magmas: Microstructural evidence. J. Geophys. Res. 95, 17,849-17,859.

Vielzeuf, D. and Holloway, J.R. (1988) Experimental determination of the fluid-absent melting relations in the pelitic system. Con. Min. Pet. 98, 257-276.

Vielzeuf, D. and Vidal, P. (1990) Granulites and Crustal Evolution, NATO ASI, Series C. Kluwer, Dordrecht, 585 p.

Vigneresse, J.-L. (1988) Forme et volume des plutons granitiques. Bull. Geol. Soc. France 8, 897-906

Vigneresse, J.-L. (1990) Use and misuse of geophysical data to determine the shape and depth of granitic intrusions. Geol. J.25, 249-260.

Vogel, T.A., Eichelberger, J.C., Younker, L.W., Schuraytz, B.C., Horkwitz, J.P., Stockman, H.W. and Westrich, H.W. (1989) Petrology and emplacement dynamics of intrusive and extrusive rhyolites of Obsidian Dome, Inyo Craters volcanic chain, eastern California. J. Geophys. Res. 94, 17,937-17,956.

Vyhnal, C.R., McSween, H.Y. and Speer, J.A. (1991) Hornblende chemistry in southern Appalachian granitoids: Implications for aluminum hornblende thermobarometry and magmatic epidote stability. Amer. Mineral. 76, 176-188.

Waskom, J.D. and Butler, J.R. (1971) Geology and gravity of the Lilesville Granite batholith, North Carolina. Geol. Soc. Amer. Bull. 82, 2827-2844.

Westrich, H.R., Stockman, H.W. and Eichelberger, J.C. (1988) Degassing of rhyolite magma during ascent and emplacement. J. Geophys. Res. 93, 6503-6511.

Whitney, J.A. (1975) The effects of pressure, temperature and XH_2O on phase assemblage in four synthetic rock compositions. J. Geol. 83, 1-27.

Whitney, J.A. (1988) The origin of granite: The role and source of water in the evolution of granitic magmas. Geol. Soc. Amer. Bull. 100, 1886-1897.

Whitney, J.A. and Stormer, J.C. (1976) Geothermometry and geobarometry in epizonal granitic intrusions: A comparision of iron-titanium and coexisting feldspars. Amer. Mineral. 61, 751-761.
Whitney, J.A. and Stormer, J.C. (1977) Two-feldspar geothermometry, geobarometry in mesozonal granitic intrusions: Three examples from the Piedmont of Georgia. Contrib. Mineral. Petrol. 63, 51-64.
Wolf, M.B. and Wyllie, P.J. (1989) The formation of tonalitic liquids during thge vapor-absent partial melting of amphibolite at 10 kbar. EOS Trans. Amer. Geophys. Union 70, 506.
Wyllie, P.J. (1977) Crustal anatexis: An experimental review. Tectonophysics 43, 41-71.
Wyllie, P.J. (1988) Magma genesis, plate tectonics and chemical differentiation of the earth. Rev. of Geophysics 26, 370-404.
Wyllie, P.J., Huang, W.-L., Stern, C.R. and Malloe, S. (1976) Granitic magmas: possible and impossible sources, water contents, and crystallization sequences. Can. J. Earth Sciences 13, 1007-1019.
Zen, E-an. (1988) Phase relations of peraluminous granitic rocks and their petrogenetic implications. Ann. Rev. Earth Planet. Sci. 16, 21-51.
Zen, E-an. (1989) Plumbing the depths of batholiths. Amer. J. Sci. 289, 1137-1157.
Zen, E-an. and Hammarstrom, J.M. (1984) Magmatic epidote and its petrologic significance. Geology 12, 515-518.
Zhu, C. and Sverjensky, D.A. (1991) Partitioning of F-Cl-OH between minerals and hydrothermal fluids. Geochimica 55, 1837-1858.

ACKNOWLEDGMENTS

The support of NSF grants EAR-9019217 and OCE-9000913 is acknowledged. I also thank Ralph Dawes, Cliff Todd, and Professors Alberto Patiño-Douce and Stu McCallum for partial reviews and comments.