Chapter 13 Aspects of Magma Generation and Ascent in Continental Lithosphere

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Overview

The variety of melt-producing reactions and rheological conditions in the crust make it difficult to generalize many aspects of crustal magmatism. Basalt provides both the material and the heat to initiate and sustain magmatism as demonstrated by unequivocal geophysical and geological evidence. Following intrusion of basaltic magma in the deep crust, the quantity of melt produced and its composition are controlled by the presence of hydrous phases; however, it is inappropriate to parameterize the amounts of melt produced by consideration of the modal percentage of the hydrous phases alone. Laboratory and numerical experiments on the heat transfer following hypothetical basaltic underplating indicate that the thermal exchange is largely conductive, in agreement with geological examples. Aggregrate viscosity is the most important physical parameter in controlling melt homogenization and ascent, although there is some ambiguity as to the utility of existing models for magma rheology. Compaction, diapirism, and diking do not appear to be viable ascent mechanisms when considered alone. The observations that midcrustal plutons often occupy crustal scale shear zones and that melt extraction occurs at a variety of scales suggest that models of magmatism require an approach that is not simply cast in terms of simple end member mechanical models. Crustal magmatism might be usefully considered in a geodynamical context where magma segregation is explicitly keyed to tectonic conditions and petrologic diversity is generated and properly understood on a crustal scale.

Notation

		Units
H	Height of underplating model	m
I	Thickness of basalt layer	m
L	Specific latent heat	$J \cdot kg^{-1} \cdot K^{-1}$
Nu	Nusselt number	dimensionless
P	Pressure	$kg \cdot m^{-1} \cdot s^{-2}$
Ste	Stefan number	dimensionless
T	Temperature	K, °C
T_{CR}	Initial country rock temperature	K, °C
T_M	Initial basalt temperature	K, °C
W	Width of underplating model	m
X	Thickness of melt region	m
b	Grain diameter	m, cm
c_p	Specific heat at constant pressure	$J \cdot kg^{-1}$
g	Scalar acceleration of gravity	$m \cdot s^{-2}$
k	Thermal conductivity	$J \cdot s^{-1} \cdot m^{-1} \cdot K^{-1}$
t	Time	s
u	Horizontal component of the velocity	$m \cdot s^{-1}$
w	Vertical component of the velocity	$m \cdot s^{-1}$
x	Horizontal independent variable	m
z	Vertical independent variable	m
β	Thermal expansion coefficient	K^{-1}
γ	Power-law coefficient in viscosity relation	dimensionless
κ	Molecular thermal diffusivity	$m^2 \cdot s^{-1}$
μ	Dynamic viscosity	$kg \cdot m^{-1} \cdot s^{-1}$
μ_r	Reduced dynamic viscosity	dimensionless
ν	Kinematic viscosity	$m^2 \cdot s^{-1}$
ρ_0	Reference density	$kg \cdot m^{-3}$
ϕ	Volume fraction solids	dimensionless
ϕ_M	Limiting volume fraction solids	dimensionless
χ	Porosity	dimensionless

Introduction

A number of models for melt migration and compositional diversity have been constructed for the mantle (McKenzie and Bickle, 1988). Less well

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understood are the processes that drive petrologic diversity in the crust. The differences lie in that the crustal composition is not readily generalized, nor are the changes in chemical potential that drive melt production well understood. The movement of melt is also more complex, due to the extreme changes in temperature and rheological state in the crust. Thus, the chemical and dynamic generalizations that have permitted some degree of progress in the modeling of processes in the mantle are not widely applicable. Despite the accessibility of crustal geologic features, even first-order answers to classical questions such as the origin and meaning of migmatites or the "room problem" in granite magmatism are not yet resolved. The crustal sections that have been described by metamorphic petrologists and structural geologists are often not like those that would seem to be required from the mass balance and transport calculations of igneous petrologists. The goal of this study is to inventory many aspects of crustal magmatism around a central theme of basaltic injection and underplating. The *geological* expression of magmatism and transport is emphasized. The purpose is to present and discuss the kinds of data that are potentially relevant to the construction of transport models.

The face of magmatism can be complex, and temporal and spatial variability is evident: at Arenal volcano, significant chemical variability is manifested on time scales of 10^1 to 10^3 yr (Reagan *et al.*, 1987), which is in contrast to compositionally monotonous systems such as Mt. Hood, and to Katmai where spatially overlapping vents erupt a variety of magma types with diverse plumbing systems and apparently unrelated eruptive styles (Hildreth, 1987). There is also some question as to the characteristic rates of intrusion and the lifetimes of chemical processes in magmatic systems. Halliday *et al.* (1989), for example, estimate that the silicic magma at Glass Mountain may have had a residence time of 0.7 Myr (also see Mahood, 1990), whereas the models of Reagan *et al.* (1991) and Gill and Condomines (1992) posit transit times from 8,000 to 50,000 yr for melt generation, transport, fractionation, and eruption. Intrusive to extrusive ratios vary as well: from 5 : 1 in oceanic settings to 10 : 1 for continents (Crisp, 1984). The physical controls on the variability and evolution of magmatic systems remain poorly characterized: it is not clear what generic feature or index of chemical and physical change is appropriate when trying to compare magmatic systems or generalize the conditions of petrologic diversity. Clearly, broad patterns emerge when the style of intrusion and eruption is indexed to tectonic regime, but within individual centers a diversity of processes is evident.

Petrologic diversity is usually considered to involve crustal melting with variable removal of refractory phases, or crystal fractionation of a mantle-derived melt that may mix with crustally derived components. The importance of crustal sources and the potential contribution of the partial retention of restite in the variations in granitoids are indicated by the presence of complexly zoned zircons (Miller *et al.*, 1992; Paterson *et al.*, 1992), the geochemistry of some enclaves (Chen *et al.*, 1990), and rather remarkable agreement between calculated magma temperatures and plagioclase compositions (Burnham, 1992). These observations indicate that the restite unmixing relations proposed by Chappell and White (1992) and Chappell *et al.* (1987) have a place in petrogenetic schemes. However, major and trace element systematics often require a more complex interpretation for the origin of many plutons and volcanic suites, involving the presence of a mantle component that yields petrologic diversity by fractionation and by providing a mixing component for anatectic melts (DePaolo *et al.*, 1992). The geochemical evidence for a mantle component is discussed in more detail below.

Both scenarios of magma generation require a change in thermal conditions in the crust. One potentially unifying element in the characterization of magmatism comes from the suggestion of Hildreth (1981) that crustal magmatism is fundamentally basaltic. In keeping with the broad interpretation suggested by Hildreth (1981), we do not mean simply a Bowen-style fractional crystallization relation. Instead basalt acts as a heat source for crustal melting and as a chemical reservoir for magma mingling/mixing. It also provides the necessary changes in rheological state to permit extension and large tectonic strains. We explore some elements of the premise that much of the petrologic diversity in magmas originates in the mid- to deep crust where the thermal conditions enhance the likelihood of melt generation and mingling between the basalt and crustal melts.

The term "underplating" has been used to describe the accumulation of basaltic magma in the mid- to deep crust. According to the underplating scenario magma is, initially at least, overlain by upward propagating regions of partial melt and also subject to continued intrusion from below. The Moho is often suggested as the site of basaltic magma accumulation and indeed may be defined by such a process. Underplating has also been used to describe the tectonic accretion of material from a subducting slab, or even the attachment of the slab itself onto the bottom of the overriding plate. Admittedly, this "underplating" nomenclature is potentially confusing; in the rest of our discussion we use underplating only to refer to the interaction of basaltic magma and the crust.

The geological and geophysical evidence for the presence of basaltic magma is as follows:

(*a*) The petrologic diversity of crustal igneous rocks requires basaltic magma as a material source, and just as importantly, as a means of providing the thermal input to generate partial melting and subsequent hybridization of the adjacent crust. In this model basalt is not simply a parent magma, but rather just one of several possible geochemical reservoirs. Numerous petrologic studies of both intrusive and extrusive rocks (Davidson *et al.*, 1990; Feeley and Grunder, 1991; Hildreth *et al.*, 1986; Manduca *et al.*, 1992) have documented the presence of basalt and other geochemical reservoirs that have interacted with the basalt. Repeated intrusion of basalt may lead to partial melting of previous underplates, which would comprise a young and mafic lower crust. Repeated melting of this material could give rise to evolved suites that would be difficult to geochemically distinguish from simple basalt fractionation (Kay *et al.*, 1990; Tepper *et al.*, 1993) if the lower crust is isotopically young. Among the most dramatic examples of the type of interaction of basaltic magma and crust is the basalt–rhyolite association at the Yellowstone Plateau volcanic field (6000 km^3 rhyolite, 100 km^3 basalt). The isotopic character of the eruptive rocks and the absence of intermediate compositions require a large-scale, deep-crustal hybridization between basalt and Archaean crust (Hildreth *et al.*, 1991). The origin of the rhyolites of the Snake River Plain–Yellowstone Plateau province is also consistent with an origin by deep-crustal melting following the intrusion of basalt (Leeman, 1982). Observations of this type provide the key elements of the MASH hypothesis of Hildreth and Moorbath (1988) (melting–assimilation–storage–homogenization), which holds that the "base level" chemical signature in some magmatic systems may be due to intrusion, partial melting, and hybridization of the deep crust by mantle-derived melts.

(*b*) Taken together, the geophysical data such as the high heat flow, uplift, the seismic reflectivity of the lower crust and a well-defined Moho are consistent with an origin by the accumulation of basaltic magma in the mid- and lower crust. The rapid rise time of the very high heat flow in the Basin and Range can be attributed only to a combination of tectonic extension and intrusion of basalt (Lachenbruch and Sass, 1978; Mareschal and Bergantz, 1990). The presence of well-defined subhorizontal reflections and downward-increasing seismic velocities, from 6.8 to 7.8 $km \cdot s^{-1}$, in the lower crust of the Basin and Range is consistent with magmatically underplated rocks (Klemperer *et al.*, 1986; Valasek *et al.*, 1987). Seismic velocity models for the deep-crustal structure in southern Alaska suggest that magmatically emplaced rocks may form part of the "crustal" root (Fuis and Plafker, 1991). Variations in seismically determined crustal thickness in Australia can be attributed to crustal underplating by mantle-derived magmas (Drummond and Collins, 1986). A number of other deep-reflection profiles have features that would permit an interpretation involving basaltic magmatism in the deep crust (Mereu *et al.*, 1989). In one sense, the Moho is a "Stefan" boundary in that its spatial variation and definition represents a degree of freedom in the response of the crust to thermal throughput on a planetary scale (Nelson, 1991).

(*c*) Exposures of the deep crust provide direct examples of magmatic additions to the lower crust with the subsequent generation of zones of partial melting and granulite formation. The best studied of these is the Ivrea–Verbano zone of northern Italy where a crustal column containing mantle peridotite, overlain by layered and homogeneous gabbro and amphibolite facies metasediments, is exposed (Handy and Zingg, 1991; Voshage *et al.*, 1990; Zingg, 1990). The isotopic similarity between the magmatic assemblage and the overlying granitoids, and the positive Eu anomaly in the mafic portions point to a genetic relationship between the two and may be an example of the lower crustal regions of partial melting and mixing.

Quick *et al.* (1992) have mapped pervasive deformation features in the Ivrea–Verbano zone that must have formed under near solidus conditions, yielding fabrics that indicate substantial lateral mass transfer while the melt was in a rheologically "mushy" state, much like textures observed in ophiolite complexes. The compositional variety and volumes of mafic magma suggest that the Ivrea–Verbano zone may represent a very thermally mature system with sustained thermal input. The other end member, perhaps representing the incipient generation of a MASH zone, where the melting of surroundings is minor, may be the Fiambala gabbronorite in Argentina (Grissom *et al.*, 1991). This sill-like body was emplaced at a depth of 21–24 km, synchronous with the local development of granulite facies conditions. The amount of crustal partial melting as a result of the thermal perturbation during intrusion is much less than that observed in the Ivrea–Verbano zone, and like that estimated by Bergantz (1989b) for a single episode of underplating.

(*d*) The occurrence of counterclockwise P–T paths and the coincidence of peak metamorphism and peak temperatures in granulite terrains are consistent with a metamorphic history where crustal thickening occurs at least partly as a result of the addition of magma to the lower crust (Bohlen and Mezger, 1989). An in-depth inventory of the granulite "controversy" is beyond the scope of this paper (the interested reader is directed to the comprehensive volume edited by Vielzeuf and Vidal, 1990). We note however that some granulites clearly have mineral phases consistent with an origin by the generation and removal of partial melt, although there are few examples where the evidence is unequivocal on a regional basis. Incomplete melt removal (Rudnick, 1992) and the uncertainty in the assumptions of trace element partitioning during m ing (Bea, 1991; Sawyer, 1991) may render the dard geochemical arguments of little use in discriminating between a restitic and nonrestitic granulite. Nonetheless, the abundance of mafic xenoliths recording pressures of 0.5–1.0 GPa and magmatic or near magmatic temperatures and textures suggest that one component of crustal growth is the ponding of basic magmas at or near the Moho (Cull, 1990; Rudnick, 1990; Rudnick and Taylor, 1987). The isotopic and trace element character of lower crustal xenoliths requires an origin by basaltic intrusion of the lower crust, followed by simultaneous partial melting of overlying crust, which may include the residue of previous episodes of underplating, fractionation of the mafic melt, and the production of melts representing varying degrees of hybridization (Kempton and Harmon, 1992; Rudnick, 1992). Continued magmatic underplating leads to the cratonization of the crust, yielding a crust that has grown from below, perhaps with episodes of delamination of hybridized and cumulate regions into the mantle (Arndt and Goldstein, 1989; Kay and Kay, 1991).

Melt Generation

The primary factors that control melting of the lower crust are (*a*) bulk composition and mineral mode, (*b*) temperature, (*c*) pressure, (*d*) volatiles (primarily H_2O and CO_2), and (*e*) the physical dynamics of the melting process. The first four factors control the chemical potential of the melt phase, and thus directly influence the extent of melting and the melt composition. The fifth factor includes kinetic factors and transport mechanisms that may overlap with the melt segregation process. This section outlines recent progress in understanding the phase equilibria that are likely to be involved in lower crustal melting. We suggest that what is currently known about lower crustal melting is consistent with the following: (*a*) the generation of many common magmatic suites is initiated by intrusion of the crust by mantle-derived mafic magma; (*b*) heat and volatiles from the mafic magmas partially melt proximal lower crust, yielding a major component of resulting magmatic suites.

A good experimental understanding of the haplogranite system is fairly well in hand, but the experiments are only indirectly applicable to melting of natural rocks besides acid granites. As stated by Tuttle and Bowen (1958), lower-crustal melting is likely to be water-undersaturated and involve mica and amphibole in the phase equilibria that control the melting reactions, the water activities in the melts, and melt productivity. It is now generally accepted that crustal melts that mobilize to form voluminous granitoids are water-undersaturated (Burnham, 1967; Clemens, 1984; Clemens and

Vielzeuf. 1987; Powell, 1983; Thompson, 1983). This has led to a focus on so-called "damp melting," in which the only water in the system is that contained in hydrous metamorphic minerals present before melting began.

Free water is unlikely to be present in pore spaces at >15–20 km depths (Yardley, 1986). It may be possible for water-rich fluids to enter the lower crust by exsolution from crystallizing, mantle-derived magmas (Lange and Carmichael, 1990; Wickham and Peters, 1992). Similarly, CO_2 or mixed H_2O–CO_2 fluids released from mafic magmas may play a role in lower-crustal melting although the source of the CO_2-dominated fluids remains to be resolved (Peterson and Newton, 1990). Although appinites and lamprophyres, common in calc-alkaline intrusions, suggest a possible role for H_2O or H_2O–CO_2 fluid-saturated mafic magmas in the generation of calc-alkaline granitoids, little research has been devoted to this topic. The production of plausible granitoid compositions in fluid-absent, lower-crustal melting experiments, and the discovery by Beard and Lofgren (1989) that water-saturated melts of amphibolite, generated at midcrustal pressures, are unlike natural granitoids, are oblique evidence in support of fluid-absent melting as the dominant process in the lower crust.

Recent chemographic and experimental work on lower-crustal melting has focused on four types of systems: model systems (i.e., specific reactions involving a few components and phases), amphibolites (of more or less basaltic composition), common granitoids (e.g., tonalites), and metapelites. Theoretical phase equilibria of model systems provide a basis for understanding the melting behavior of rocks containing hydrous minerals (Brown and Fyfe, 1970; Burnham, 1979; Thompson, 1982; Thompson and Algor, 1977; Thompson and Tracy, 1979; Wyllie, 1977). Reactions in natural rocks generally have greater thermodynamic variance than those in model systems. Melting is likely to take place through both continuous and discontinuous reactions (Thompson, 1988), making it difficult to generalize melt production.

Melting experiments that used natural rocks as starting materials, evaluated melt fraction through the melting interval, and applied pressures equivalent to the middle to deep crust, have been performed by Vielzeuf and Holloway (1988), Rutter and Wyllie (1988), Wolf and Wyllie (1989), Patiño-Douce and Johnston (1991), Beard and Lofgren (1991), and Rushmer (1991). The following points derive from these studies. (*a*) Closed-system melting of metapelites produces peraluminous melt compositions similiar to S-type granites (Patiño-Douce and Johnston, 1991; Vielzeuf and Holloway, 1988). The main melt producing reaction in both studies is biotite + plagioclase + aluminosilicate + quartz = garnet + melt. At higher temperatures garnet is also consumed to form melt, with spinel and aluminosilicate the most refractory phases. (*b*) Amphibolite melting at lower pressures (7–10 kbar) occurs by "damp" melting mainly via breakdown of hornblende, producing clinopyroxene ± orthopyroxene (Rushmer, 1991; Wolf and Wyllie, 1989). Melts range from granite or trondhjemite to tonalite with increasing temperature and cross the boundary from slightly peraluminous to metaluminous. (*c*) The importance of hydrous minerals in controlling melt productivity and composition is consistently highlighted in all of these studies; the linked parameters mineral mode and bulk rock composition exert a strong control on the amount of melt produced in fluid-absent melting.

This third point, on the relation between mineral mode and melt production, is illustrated by the very different melt fraction vs temperature results for melting of metapelite obtained by Vielzeuf and Holloway (1988) and Patiño-Douce and Johnston (1991). This difference, and the more potassic and felsic nature of the melts produced in the experiments of Vielzeuf and Holloway, is ascribed primarily to the different bulk compositions and mineral modes of the two metapelites, and how well or poorly the initial mineral mode corresponds to the stoichiometry of the major melt-producing reaction, which in this case involves biotite dehydration (Patiño-Douce and Johnston, 1991).

The results of a variety of partial melting experiments (Fig. 1) suggest that melt production vs temperature of natural, hydrous mineral-bearing metamorphic rocks covers a spectrum. At one end of the spectrum are rocks in which the relation between modes of hydrous minerals, quartz, and feldspar, and bulk minimum-melt components, fits well enough with reaction stoichiometry that most melting occurs at the reaction that consumes the hydrous mineral and releases H_2O fluid to dis-

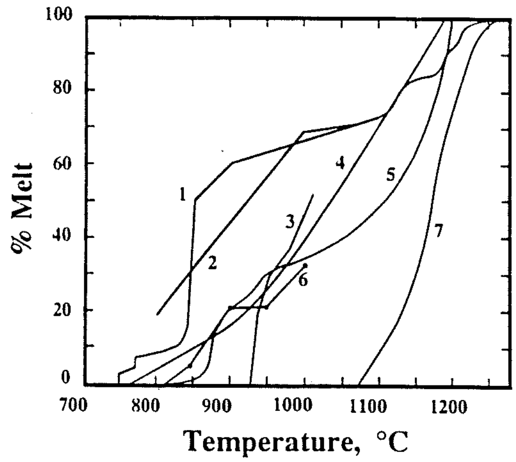

Figure 1 Percentage melt as a function of temperature for a number of lithologies. Pressures range between 7 and 10 kbar. Compositions are metapelites (1,2), silicic plutons (4,5), and amphibolite (3,6). All are damp melting: no free water is present at the start of the experiment. Note rapid and nonmonotonic changes in percentage melt as temperature increases. Curve 7 is the crystallization of a basalt. References for individual curves: (1) Vielzeuf and Holloway (1988); (2) Patiño-Douce and Johnston (1991); (3) Rushmer (1991); (4) Wyllie (1977); (5) Rutter and Wyllie (1988); (6) Beard and Lofgren (1989); (7) Marsh (1981).

solve in the melt. Such rocks will produce melt in step-like fashion, with large volumes produced over a narrow temperature interval. At the other end of the spectrum are rocks that do not contain the components in proportions suitable to maintaining quartz- and feldspar-saturated melting via dehydration reactions. These rocks will not produce major volumes of melt over narrow temperature intervals within the melting range. Because the total volume of rock undergoing partial melting will probably vary over some range of heterogeneity, it seems likely that in many real situations the intermediate pattern of the temperature vs melt fraction spectrum holds, with a dampened correlation between breakdown reactions of hydrous minerals and pulses of increased melt production.

In sum, recent work on lower crustal melting suggests a dominant role for fluid-absent, "damp" melting. Dehydration of biotite occurs at 800–900°C, producing melt fractions of >35% with granite or felsic granodiorite composition. Dehydration of hornblende occurs at 900–1000°C, producing melt fractions of >35% with tonalite or intermediate granodiorite composition. The residual, anhydrous mineral assemblages are similar to those found in granulites (Rudnick, 1992). Metapelites and mica-rich quartzofeldspathic rocks (e.g., some types of metagraywackes) produce the largest volume proportions of melt. What rock types are present in the crust is uncertain, but it is not necessarily limited to direct products of mantle-derived magmas. Metasupracrustal rocks may also be emplaced in the lower crust, particularly in convergent margin orogens (Kempton and Harmon, 1992).

In generic terms, the experimental evidence accords with the geological evidence, in that both are consistent with the initiation of crustal-scale magmatism by the intrusion of basaltic magma at the base of the crust, and the involvement of lower-crustal melting as an important part of the process. Major episodes of lower-crustal melting are associated with temperatures achievable only if basalt intrudes the lower crust, based on the temperatures required for the generation of intermediate magmas. Similarly high temperatures are indicated for many incompatible-element-depleted granulites (Bohlen and Mezger, 1989; Clemens, 1990). The compositions of magmatic rocks that may have been generated in the lower crust are very similar to the experimentally produced melts from vapor-absent runs, with an important exception.

The major point of disagreement between experimentally produced melts and natural magmatic rocks is that the Mg numbers of experimental melts are consistently lower than those of most natural rocks interpreted as probable lower-crustal melts. This is particularly true at intermediate SiO_2 values. Assuming that crustal melting was involved, this discrepancy may be accounted for by partial mingling or mixing of basaltic magmas with lower-crustal melts that were generated by the intrusion of the basaltic magmas, or by restite retention, which assumes that Ca- and Mg-rich refractory minerals from the crustal melting source are entrained in the melt when it leaves the source area. Even if restite unmixing (Chappell and Stephens, 1988) is an important process, intrusion of the lower crust by basalts must still be invoked (Burnham, 1992). The question is then whether magmas are mobilized en masse as melt plus residual solids or whether the melt segregates efficiently from the residual solids, although the answer is undoubtedly somewhere in between and will vary on a case-by-case basis.

Modeling the Physical Interaction of Basalt and Crust

The effectiveness of melt generation, segregation, and homogenization will depend on the rate at which melt is generated, the differences between the viscosity of the crystal–liquid *mixture*, the melt-only viscosity, and the bulk viscosity of the solid matrix. Hence, one of the most important elements in characterizing the dynamics is the rheological state of the partial melt zone (Bergantz, 1990; Wickham, 1987). In addition, there is a growing body of geological evidence that there is a feedback between the stress field in the crust and the style and timing of ascent. Before we consider models of underplating and segregation mechanisms, we will consider the rheological characterization of crystal–melt mixtures.

Rheology of Suspensions

Models for the viscosity of magmatic *liquids* are given by Bottinga and Weill (1972) and Shaw (1972) and a comprehensive summary of the available data on the viscosity of melts of geologic interest can be found in Ryan and Blevins (1987). Obviously, estimates of viscosity based on consideration of only the changing liquid composition have little application to dynamical models of crystal–melt systems; the viscosity can range from nearly infinite under near solidus conditions to 1 Pa · s under near liquidus conditions. Developing robust quantitative models for the influence of increasing crystallinity on the viscosity is difficult: some scaling relations are reviewed by Jomha *et al.* (1991). Even making reliable measurements of the viscosity of suspensions has proven challenging: the same crystal–liquid suspensions yield different viscosities when measured in different viscometers and the measurements are often difficult to reproduce even when repeated evaluations are made on the same viscometer. These results cast doubt on the notion that a single expression can adequately characterize the viscosity of dense suspensions (Cheng, 1984; Frith *et al.*, 1987). The origin of the irreproducibility may reside in variations of the solid packing structure, an organization of the crystals that is an inevitable consequence of flow. This will vary depending on the flow regime and crystal shape and induce a feedback that is particular to a given geometry, shear rate, etc. Despite these difficulties, various empirical relationships can be taken from the engineering literature, although it is important that these results be applied to geologic conditions with caution.

To the viscosity calculated for the crystal-free liquid, a "correction" for the influence of crystals can be added. One such expression often cited is a form of the Krieger–Dougherty expression given by Wildemuth and Williams (1984),

$$\mu_r = \frac{1}{(1 - \phi/\phi_M)^\gamma}, \tag{1}$$

where μ_r is the reduced viscosity, which is the actual viscosity divided by the viscosity of the crystal-free liquid at the same temperature, pressure, and composition; ϕ is the volume particle fraction; and ϕ_M is the volume fraction of particles at which the three-dimensional contact between the particles causes strain to cease at a given shear stress. It should be appreciated that ϕ_M is shear stress dependent and using the absolute maximum packing fraction for ϕ_M may overestimate ϕ_M for many magmatic conditions where strain rates associated with convection may be low. ϕ_M ranges from about 0.5 at low shear stress to a value of 0.75 for infinite shear stress, and γ is a coefficient with experimentally determined values that vary from about 1.3 to almost 3 with a value of 2 being typical of many systems (Barnes *et al.*, 1989; Sengun and Probstein, 1989; Wildemuth and Williams, 1984). Although weakly shear stress dependent, γ is nearly invariant with variations in stress over two orders of magnitude for a wide variety of particle sizes and mixture properties (Wildemuth and Williams, 1984).

The MELTS program (a substantial refinement of the SILMIN algorithm of Ghiorso (1985)) was used to calculate the viscosity as cooling proceeds for a given initial composition and pressure using the method of Shaw (1972). The MELTS program is robust in that it allows for the crystallization of practically all the relevant phases: pyroxenes, common feldspars, quartz, biotite, olivine, the Fe–Ti oxides, and others. In addition, the oxygen fugacity can be controlled. Using MELTS, changes in melt composition and volume percent crystallinity can be calculated as a function of temperature, including the effects of water in the melt.

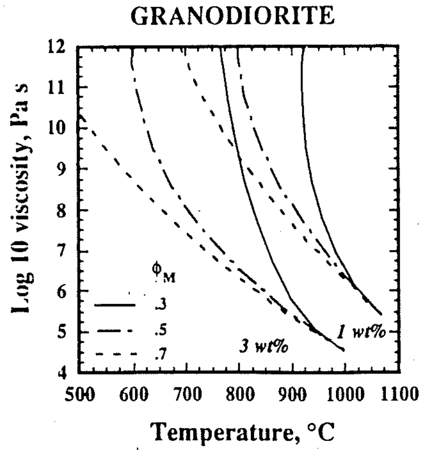

Figure 2 Calculated variation of viscosity with temperature for the Red Lake granodiorite (Noyes *et al.*, 1983) at two different initial water contents, 1 and 3 wt% H_2O. The calculations were done with the MELTS algorithm assuming the oxygen was buffered at Ni–NiO and a total pressure of 2 kbar. Water content will increase as crystallization of anhydrous phases continues. Very high water contents yield low solidus temperatures; given the absence of experimental data under these conditions, viscosity values for temperatures under 600°C should be used with caution.

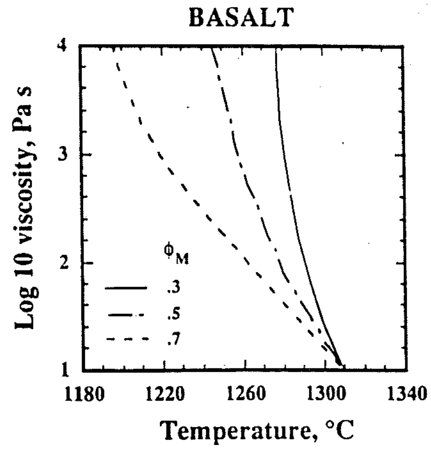

Figure 3 Calculated variation of viscosity with temperature for the mafic inclusions (basaltic composition) in the Burnt lava flow of Grove *et al.* (1988). The calculations were done with the MELTS algorithm assuming the oxygen was buffered at Q–F–M and a total pressure of 8 kbar. These curves are typical of a variety of basalts. Adding small amounts of water to the system (1–2 wt%) had a negligible effect on the calculated viscosity.

Thus at any temperature, the calculated melt composition, water content, and volume fraction crystals can be used in Eq. (1) to yield the viscosity.

Two examples of the calculated viscosity are given here, a granodiorite in Fig. 2 and a basalt in Fig. 3. Three curves are shown in each of the figures; each curve corresponds to a different assumed value of the critical volume fraction, ϕ_M. There is little unambiguous geological evidence of the value of the critical volume fraction as discussed by Bergantz (1990) but a value around 0.5 is suggested from observations on extrusive and plutonic rocks (Marsh, 1981; Miller *et al.*, 1988). For each curve in Figs. 2 and 3, all values of melt viscosity and crystallinity are the same at a given temperature, the only difference among the curves is the value of ϕ_M. The curves for a critical volume fraction of 0.3 and 0.7 delimit an envelope of reasonable values. It is apparent that increasing crystallinity dominates the variation in viscosity. Two initial water contents are given for the granodiorite to demonstrate the effect that increasing water content in the melt will have on the viscosity.

This is important when one considers that the volume fraction of solids as a function of temperature can be very different depending on whether temperature is falling, such as in the case of a magma, or temperature is rising, which is the case in partial melting. Volume fraction curves tend to be smoother for solidification as the nucleation is typically heterogeneous and the undercoolings are demonstrably small (Cashman, 1993). In compositionally similar, but "dirtier" rocks, retrograde assemblages can profoundly affect the melt fraction distribution during partial melting; see Fig. 1. Hence, the disposition of the partial melt may not be simply related to temperature as discussed in Bergantz (1992). This has been demonstrated by Wolf and Wyllie (1991) where it was observed that melt distribution in anisotropic partial melts was not controlled by dihedral angles. This raises the question as to the degree to which equilibrium is maintained during partial melting. The agreement between experimentally produced and calculated major element melt compositions often agrees well with those seen in the rock record (Ashworth and Brown, 1990; Burnham, 1992). Direct natural evidence is uncommon; however, in a remarkable study of the partial melting of a magma chamber's walls that was interrupted by eruption, Bacon (1992) documents that equilibrium was not achieved between plagioclase and melt on time scales of 10^2–10^4 yr. The implica-

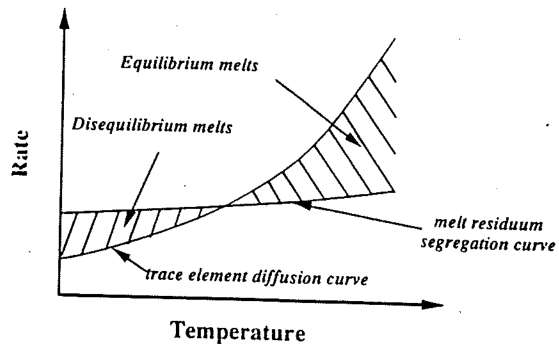

Figure 4 Schematic of relationship between melt segregation and rate of trace element equilibration between melt and residuum. After Sawyer (1991).

tions of disequilibrium melting during melt generation and the geochemical interpretation of migmatites have been addressed by Bea (1991) and Sawyer (1991). The approach to equilibrium can be considered a competition between diffusion and segregation and is shown schematically in Fig. 4. Rubie and Brearley (1990) explore a number of simple models of disequilibrium melting to examine the sensitivity of the melt fraction distribution to the overstepping of the solidus temperature and the effects of variable latent heats of melting, the volume fraction of the melting species, grain sizes, and the diffusion activation energy. For the conditions of magmatic underplating, the initially large thermal overstepping rapidly yields the appropriate melt fraction. However, as the thermal gradient extends outward, the melt fraction may not be simply related to temperature as the *rates* of heating and cooling will be less. Unfortunately, no complete kinetic model for melt-forming reactions exists. This is due in part to the experimental difficulty in establishing the reaction stoichiometry and developing the appropriate kinetic model; the most recent summary of metamorphic kinetics can be found in Kerrick (1991). Hence, estimating the rheological state during partial melting is subject to considerable uncertainty. The available experiments are few and reveal a strong sensitivity of deformation style to melt fraction, particularly near the solidus (Dell'Angelo and Tullis, 1988).

Numerical Model of Basaltic Underplating

We begin our discussion by invoking the simplest imaginable process, as illustrated in Fig. 5. Basaltic magma at the liquidus intrudes some region of the crust that is at ambient temperature T_{CR}. In response to the intrusion, melting begins in the country rock and, subsequently, a region of solid and liquid will form in the region once occupied by the magma. This simple picture has been examined using both numerical and laboratory experiments by a number of workers (Bergantz, 1989a, 1989b; Campbell and Turner, 1987; Fountain *et al.*, 1989; Hodge, 1974; Huppert and

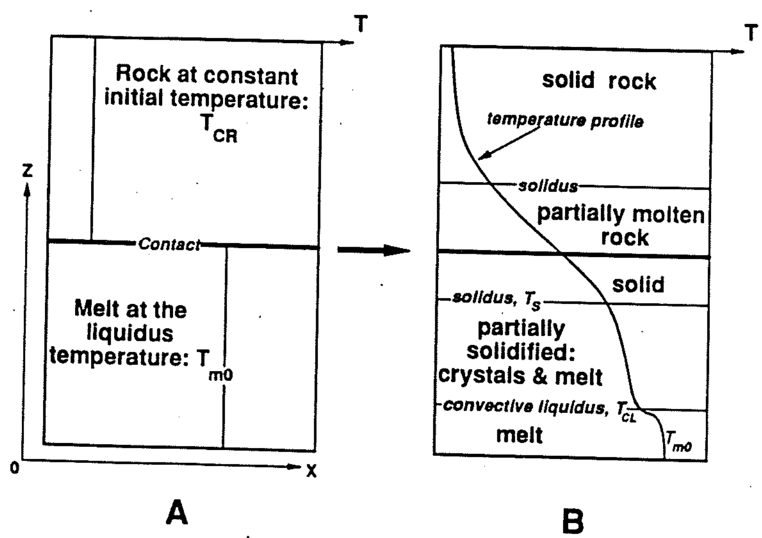

Figure 5 The initial conditions at the time of basaltic underplating, with subsequent evolution. Well-mixed basalt at the liquidus temperature forms a planar interface contact against country rock, which is at some ambient temperature T_{CR}. (A) The temperature profile initially has a step profile that decays, yielding (B) regions of crystal–melt "mush."

Sparks, 1988; Wells, 1980; Yoder, 1990; Younker and Vogel, 1976). Each of these studies has emphasized different physical aspects of basaltic underplating and yielded results that are not all in agreement. The laboratory experiments often demonstrate a wide variety of convective phenomena; however, the solid fraction–enthalpy, and hence transport property, relationships between the model basalt and country rock do not match those found in natural systems. The analytical models usually *assume* that the enthalpy transfer is only by conduction, and so there is always some question as to the role of convection in the magma in enhancing the heat transfer and partial melting of the country rock.

The style and vigor of convection in magma chambers has been subject to uncertainty, and even controversy (Huppert and Sparks, 1991; Marsh, 1991). Much of the uncertainty lies in the difficulty in forming *testable* hypotheses. The simple fluids available for table-top experiments rarely demonstrate the strong rheological contrasts and buoyancy relationships present in magmatic systems. On the other hand, the products of magmatism rarely provide unambiguous evidence of the processes involved in their generation and transport. Buoyancy in magmas is also generated in complex ways, with thermal effects being important near the liquidus and compositional ones important near the solidus. There is no question that magmas convect, although time dependence and the presence of multiple length scales often make the fluid structures and thermal history difficult to generalize with the conventional parameterizations developed for simple fluids. The emphasis of the model presented here is convection and partial melting under one set of arguably geologically relevant conditions. We note that our model assumptions do preclude some aspects of the heat and mass transfer in magmas, such as volatile release and the capacity to model individual crystal growth and settling. However, it is hoped that calculations presented here, which emphasize some of the known rate-limiting features present in geological systems, will encourage a discussion of magmatic convection that directly addresses geological conditions.

One model of the conjugate enthalpy transfer between crystallizing basalt and simultaneous partial melting of the overlying country rock is presented. The important new features of this model are that we are using a set of "best guess" calculated physical property variations: viscosity, density, and all the thermophysical properties vary temporally and spatially as crystallization and concomitant partial melting proceed. In particular, the model attempts to retain the appropriate relationships for enthalpy content and transport properties in the combined basalt–country rock system. Convection in the basalt occurs, and the time dependence, vigor, and form of the convection can be directly related to the crystallization history, which itself is constrained by the phase relations determined under geologically relevant conditions and the progress of melting in the country rock. Thus, the questions of interest are: How might magmatic convection influence the timing of partial melting in the overlying country rock and what is the style of this convection? How does variable viscosity generated by the changing crystallinity and composition of the melt influence the convection? What might the characteristic convective velocities be, and what is the form of the instabilities? What are the overall heat transfer rates from the system relative to those where the transfer is by conduction only?

Oldenburg and Spera (1992) have developed a clever hybrid model for the numerical simulation of crystallizing systems. Their model is a combination of two end-member states: the relative motion condition where interdendritic melt can flow through the crystal pile, and the no relative motion condition where the fluid that is moving through any control volume includes the crystals and equilibrium is assumed. The model considered here uses the no relative motion assumption. In practice this means that at any point in the flow field, the crystallinity, composition, density, viscosity, and all thermophysical properties are determined from the phase constraints directly. This is essentially just application of the phase rule to the control volume over which the local average temperature is determined.

The advantage of this approach is that all the transport property variations can be indexed to temperature. What is required is a means of calculating how density, specific heat, crystallinity, viscosity, and conductivity vary as a function of temperature, including the effects of melt composition as crystallization proceeds. The MELTS algorithm of Ghiorso (discussed previously) provides these quantities. The disadvantage of this approach is that it does not allow for nonequilibrium crystal–liquid segregation. For example, if crystals begin to settle out individually or in the form of a plume, it is possible that interstitial melt could be removed as the plume falls. In our model, any time magma convects to a new region where temperature is different, it is assumed that the crystallinity changes accordingly; temperature and crystal fraction are linked. Thus, convection driven strictly by crystals falling cannot be accommodated. We do not consider this a restrictive condition as it is not clear what role settling crystals may have in driving *bulk* convection.

The combined basalt–country rock system shown in Fig. 5 was represented on a two-dimensional 75 × 75 grid of total width W of 15 m and total height H of 25 m. The depth of the initial basalt layer is l and was 20 m thick. The origin was placed at the lower left corner and Cartesian coordinate axes were introduced: x being the horizontal coordinate axis and z the vertical axis, where z is taken as positive in the upward direction. Fifty-five grid points in the vertical direction were assigned to the fluid layer and twenty to the overlying country rock. Variable grid spacing was used in both the vertical and horizontal directions to ensure that the boundary layers near the intrusive contact could be adequately characterized. Using fewer grid points (40 × 40) yielded the same Nusselt number relationships discussed here, but some details of the thermal structure of the plumes were lost so we elected to use the finer grid.

A comprehensive discussion of the continuum equations describing the heat and mass transfer in systems undergoing crystallization, with possible double-diffusive convection is available (Beckermann and Viskanta, 1988, 1993; Bennon and Incropera, 1987; Ni and Beckermann, 1991; Oldenburg and Spera, 1991, 1992). We will not revisit these derivations, but direct the reader to these sources for clarification of the continuum expression of the model.

The equations required to describe the velocity and temperature fields following underplating, invoking the Boussinesq approximation, are given for

conservation of mass (continuity),

$$\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0, \tag{2}$$

conservation of momentum,

$$\rho_0\left[\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + w\frac{\partial u}{\partial z}\right] = -\frac{\partial P}{\partial x} + \frac{\partial}{\partial x}\left(\mu\frac{\partial u}{\partial x}\right) + \frac{\partial}{\partial z}\left(\mu\frac{\partial u}{\partial z}\right) + \frac{\partial \mu}{\partial x}\frac{\partial u}{\partial x} + \frac{\partial \mu}{\partial z}\frac{\partial w}{\partial x} \tag{3}$$

$$\rho_0\left[\frac{\partial w}{\partial t} + u\frac{\partial w}{\partial x} + w\frac{\partial w}{\partial z}\right] = -\frac{\partial P}{\partial z} + \rho g + \frac{\partial}{\partial x}\left(\mu\frac{\partial w}{\partial x}\right) + \frac{\partial}{\partial z}\left(\mu\frac{\partial w}{\partial z}\right) + \frac{\partial \mu}{\partial x}\frac{\partial u}{\partial z} + \frac{\partial \mu}{\partial z}\frac{\partial w}{\partial z}, \tag{4}$$

and conservation of energy,

$$\rho_0 c_P\left[\frac{\partial T}{\partial t} + u\frac{\partial T}{\partial x} + w\frac{\partial T}{\partial z}\right] = \frac{\partial}{\partial x}\left(k\frac{\partial T}{\partial x}\right) + \frac{\partial}{\partial z}\left(k\frac{\partial T}{\partial z}\right), \tag{5}$$

with initial and boundary conditions on region and boundary temperatures,

$$T(x,z,0) = T_M, \quad 0 \le z \le l; \qquad T(x,z,0) = T_{CR}, \quad l \le z \le T_{CR}, \tag{6}$$

$$\left.\frac{\partial T}{\partial x}\right|_{x=0} = 0, \quad \left.\frac{\partial T}{\partial x}\right|_{x=W} = 0, \quad \left.\frac{\partial T}{\partial z}\right|_{z=0} = 0, \quad T(x,H,t) = T_{CR}, \tag{7}$$

and on mass flux (motion components) at the boundaries,

$$\left.\frac{\partial u}{\partial z}\right|_{x=0} = 0, \quad \left.\frac{\partial u}{\partial z}\right|_{x=W} = 0, \quad u(x,0,t) = 0, \quad u(x,l,t) = 0 \tag{8}$$

$$\left.\frac{\partial w}{\partial x}\right|_{x=0} = 0, \quad \left.\frac{\partial w}{\partial x}\right|_{x=W} = 0, \quad u(x,0,t) = 0, \quad w(x,l,t) = 0 \tag{9}$$

where l is the thickness of the fluid layer. The initial conditions are constant temperature in the country rock T_{CR}, and the basalt is assumed to have intruded at its liquidus, T_M. The horizontal boundary conditions on temperature and velocity are reflecting conditions where it is assumed that the magma layer is infinite in the horizontal dimensions. Thus, effects due to the presence of side walls as sources of drag or heat are not considered. The vertical conditions are that the magma layer is effectively insulated at the bottom. This means that the convection is driven only by cooling from above and is fundamentally different from systems that are heated from below and cooled from above, and buoyancy is generated at *both* boundaries. This seems reasonable as it is difficult to imagine what heat source would be hotter than the basalt itself.

The numerical algorithm used to solve Eqs. (2–5) is a modified version of the TEMPEST code developed at Pacific Northwest Laboratories (Trent and Eyler, 1991). It uses a semi-implicit, finite volume method. We have established code accuracy and reliability by doing benchmark runs on a number of classical natural convection problems, and we have found excellent agreement with published numerical and experimental results, including conditions of variable transport properties.

We consider only one set of calculations here, but one that reveals the important features observed in a variety of simulations. The model basalt composition used was that of the Burnt Lava flow (Grove *et al.*, 1988). The crystallization was assumed to occur under oxygen-buffered conditions of quartz–fayalite–magnetite. It was also assumed that crystallization occurred in the fractionation mode. This does not mean that the crystals were physically removed as temperature dropped, but rather that as crystals grow, they do not back-react and the melt composition evolves accordingly. This is really a statement that thermal equilibrium might be maintained, but the crystals will remain zoned. If resorption is required by advection of crystals to a region of higher temperature, it will initially involve the rim of the crystal. This decision was based on the fact that plagioclase was the dominant near liquidus phase. Very little difference in the convective state or the progress of melting of the country rock was found in simulations where the physical properties were generated assuming complete equilibrium between melt and solid phases.

The MELTS algorithm was used to calculate the required physical property data. The numerical prescription of the enthalpy changes associated with phase changes was done by defining a temperature-dependent effective specific heat capacity that includes changes in both sensible and latent heats. Figure 6 is a plot of the calculated variation in effective specific heat as a function of temperature for the basalt. The nonmonotonic behavior of the effective specific heat as a function of temperature is the result of the magma becoming saturated with new phases as cooling proceeds. Once the melt has become saturated with a phase, with a concomitant "burst" of crystal growth, the effective specific heat will diminish as the increase in crystallinity per degree of temperature drop diminishes. Although the liquidus phase was olivine, the increase in crystallinity with temperature at the liquidus is modest; the two dominant peaks in Fig. 6 are due to the saturation of the magma with plagioclase at about 1275°C and with spinel at 1085°C. Pyroxene appears at about 1260°C. Sensitivity analysis of the model results indicated that geologically reasonable variations in the absolute values, or reasonable variations in the functional form of the effective specific heat function, had little impact on the results. Calculated density variations are shown in Fig. 7. The system density was used in the calculations, in keeping with our no relative motion postulate. This system density reflects both changes in the melt density as cooling and composition of the melt changes during crystallization and increases in density associated with the appearance of the solid phases. The viscosity–crystallinity–temperature relationships for the crystallizing basalt are shown in Fig. 3.

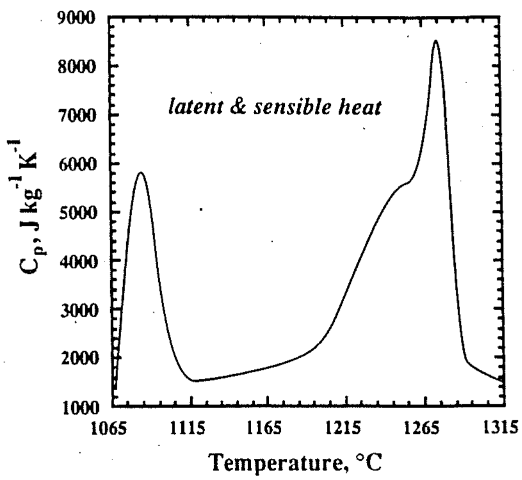

Figure 6 Calculated effective specific heat capacity for the model basalt. This effective specific heat includes both the effects of phase change and an overall drop in system temperature. It also includes the temperature-dependent heat capacities of the minerals and melt. The peak in effective heat capacity at about 1275°C is due to the onset of feldspar crystallization; the peak at about 1085°C is due to the appearance of spinel.

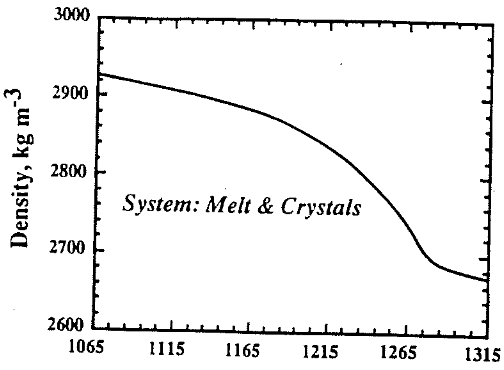

Figure 7 Calculated density of the model basalt. This density includes all the phases present as both crystals and liquid.

The melt fraction–temperature relationship used for the model country rock was that of Rutter and Wyllie (1988) who melted a tonalite under vapor-absent conditions (Fig. 1). They do not report enthalpy changes associated with the melting and so a constant value of 1.5×10^5 J · kg^{-1} was used. This value is consistent with those calculated by Hon and Weill (1982). Temperature-dependent thermal conductivity curves for both the basalt and the tonalite protolith were taken from Touloukian *et al.* (1981).

We initialize the model by assuming a pressure of 8 kbar with an ambient temperature of T_{CR} equal to 450°C and a basalt liquidus, T_M equal to 1315°C. After a short period following intrusion to allow the cold boundary layer at the roof to develop, the initial instability was generated by introducing a random thermal perturbation of 1 to 3°C at each grid point near the basalt–country rock boundary. These rather high perturbation values were required as the presence of variable viscosity rendered the cooling boundary layer very stable. Once convection started, the form and disposition of the instabilities were independent of the location and magnitude of the initial perturbations.

Many of the questions related to the nature of the coupled crystallizing basalt and melting of the country rock can be resolved by considering the time-dependent Nusselt number for the system. In this instance the Nusselt number is defined to be the ratio of the total heat transfer in the presence of convection to that which would occur under conditions of conduction only. This ratio was calculated using the thermal flux determined at the contact between the magma layer and the country rock. The maximum Nusselt number under these conditions is 2 and would occur if the magma were perfectly well mixed by some unspecified stirring agent. To obtain the conduction results for calculation of the Nusselt number, we did a numerical simulation identical to that with convection, except that no fluid motion was permitted. The Nusselt number as a function of the dimensionless conduction time is shown in Fig. 8: four curves are shown and we consider each in turn.

The first is the case of constant viscosity. In this case, crystallization is occurring, but the viscosity is held constant at the value the magma had at the time of intrusion; hence, there is no effect on the viscosity due to increasing crystallinity or changing composition and temperature. Although this is clearly unrealistic for magmas, it provides a useful end-member result. The Nusselt number rises rapidly to a maximum of about 1.77 and then decreases as the magma reservoir temperature decreases and the country rock heats up. The other three curves correspond to cases where the viscosity varied as a function of critical solid fraction, ϕ_M. The curve for a critical volume fraction of 0.7 is queried as there appeared to be abrupt and unreproducible oscillation in the thermal flux.

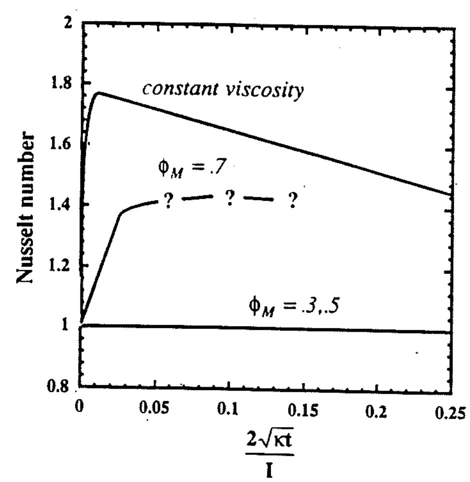

Figure 8 Nusselt number as a function of the dimensionless conductive cooling time. The thickness of the melt layer is l, the thermal diffusivity is κ, and time is t. Four curves are shown: one for the case where viscosity is held constant and three that correspond to the viscosity–temperature relations given in Figure 3. The curves for critical volume fractions, ϕ_M, of 0.3 and 0.5 are practically indistinguishable from each other. See text for a discussion of the query on the curve for ϕ_M equal to 0.7.

perhaps indicating the abrupt destruction of the stagnant layer at the roof (see discussion following); because this was not reproducible, that curve should be interpreted with caution. The cause of this behavior is still under investigation: we will not discuss it further.

The curves for critical volume fractions of 0.3 and 0.5 fall practically on top of each other in Fig. 8. Both have Nusselt number values so close to 1 that it is not apparent from Fig. 8 that there is any difference between them. Under these conditions, even though convection is occurring, the heat transfer to the country rock, and hence the progress of melting, is as if the magma were cooling by conduction only. This is despite the fact that the initial temperature difference between the magma and the country rock was 865°C. For these model runs the region of the magma in which convection is initiated is where the viscosity is varying by no more than an order of magnitude, and more typically by no more than a factor of 2, and where the crystallinity is quite low. This indicates that convection is occurring in a region where the temperature differs by no more than a few (one to five) degrees. Another important observation was that the Nusselt number decayed monotonically; there was no melting back or resorption of the high-viscosity sublayer.

The disposition of the isotherms and plume morphology was similiar to that seen in the study of variable viscosity convection by Olson *et al.* (1988); see their Figs. 17, 18, and 21. For all the cases considered in this study, the convection formed plumes that fell from the leading edge of the crystallizing front. There was no regularity in the positions from which the plumes originated or in the numbers of plumes that might exist in a fully developed state at any time; the plume morphology was broadly like that of classic thermals. The flow field was always irregular but not turbulent; the maximum vertical velocity associated with this convection was of order 10^{-4} m · s^{-1}, and typically much smaller, about 10^{-5} m · s^{-1}. Despite these low velocities the fluid motion was able to keep the interior largely isothermal, probably due to the small temperature contrast associated with convection. Although the flow would on occasion be organized into cells with high aspect ratios (length/width), which did not penetrate to the floor, it was more common for the return flow to be dispersed and irregular.

The partitioning of the fluid into a region with a rigid lid where high viscosity inhibits convection, and an adjacent region where convection can occur, has been noted in other studies of variable-viscosity convection (Chen and Pearlstein, 1988; Chu and Hickox, 1990; Ogawa *et al.*, 1991; Olson *et al.*, 1988; Richter *et al.*, 1983). Of particular relevance are the studies of Jaupart and Parsons (1985) who examine the case where a variable-viscosity fluid layer is cooled only from above, and Brandeis and Jaupart (1986) and Smith (1988) who also add the complication of simultaneous crystallization. Smith (1988) uses linear stability analysis to determine the critical Rayleigh number and demonstrates that, in the near-critical regime, high aspect ratio cells that may not penetrate to the bottom of the chamber form and that the appropriate length scale represents a balance between thermal diffusion and propagation rate of the solidification front. For supercritical conditions where the critical Rayleigh number is substantially exceeded, it is estimated that the convection will still be gentle, as the convective vigor is modulated by variable viscosity and the rate-limiting steps involved with heat transfer out of the roof by conduction.

The results obtained from our modeling validate the suggestions of Smith (1988) and are in agreement with the laboratory experiments of Bergantz (1989a). The convection is never vigorous, although it certainly does occur. The high aspect ratio cells predicted by Smith (1988) were observed, but they appear on an intermittent basis, which is to be expected in the supercritical regime. This type of behavior has been seen in geophysical models of lithospheric convection below the oceanic crust where the mantle has been modeled as a variable-viscosity fluid (Buck and Parmentier, 1986). These results are also consistent with the models of magma chamber convection proposed by Marsh (1989), who addresses some of the petrologic implications and details of the parameterization of convection.

The calculated progress of melting of the country rock due to the presence of a convecting basalt layer follows directly from the low values of the Nusselt number. For the cases where the critical melt fraction ϕ_M was 0.3 and 0.5, the amount of melt generated at the intrusive contact and the way it changed as melting proceeded were indistinguishable from that calculated in the conduction-only models (Bergantz, 1989b). It thus appears that within the framework of the assumptions invoked in this model, calculating the progress of melting following a single episode of underplating can be facilitated by assuming a simple, but still nonlinear, conductive cooling model. These results clearly indicate that multiple intrusions are required to thermally mature a deep crustal section to incite melt generation on a regional scale. However, it should be remembered that most geochemical models of partial melting (see preceding discussion) require melting in the range 10–35%, which is like that estimated here and in Bergantz (1989b). The compositions of melts produced by wholesale melting of the lower crust deviate from naturally occurring compositions (Beard and Lofgren, 1991) and models that would predict melting in excess of 50% should be invoked with caution. Thus, there is no justification for models that suggest that the incipient stages of partial melting involve *bulk* melting of the lower crust.

Our results are also in keeping with observations made during large-scale melting experiments. An artificial magma chamber, or melt layer, 3 × 1.5 m was generated by melting soil at the Oak Ridge National Laboratory (Jacobs *et al.*, 1992). The melting was initiated and sustained by applying electrical power to the ground through four graphite electrodes. The melt pool convected vigorously as it was thermally "pumped" and superheated conditions were attained: melt temperatures reached 1500°C. However, once the power was shut off, convection could be driven only by heat losses to the environment under conditions of thermal decay; convection ceased once the liquidus of the melt was attained (Dunbar, 1993; oral communication). This is an important demonstration of the fact that the rheological conditions associated with the onset of crystallization control the subsequent dynamic evolution of the body.

It should be noted that a number of conditions that could alter the broad conclusions of our model might exist. If the magma chamber experienced reintrusion and subsequent disruption of the upper viscous boundary layer, then a pulse of melting might follow. If the roof were not infinitely rigid, as assumed here, it could buckle under its own weight and delamination of the roof could occur. This would abruptly bring together partially melted country rock with near liquidus temperature magma, yielding a condition that could lead to magma mixing. Any volatile release will aid melt production in the country rock, which we have not considered. Other geometries, such as the multiple dikes depicted by Grove *et al.* (1988), may yield regions of overlapping thermal gradients and enhanced melting. It was also found in the course of a sensitivity analysis that some variable viscosity–temperature relationships *did not* yield a steady-state viscous lid and whole-layer convection occurred. Although these viscosity–temperature relationships are not like those calculated by the MELTS code for use in this example, it is not unreasonable to speculate that there is more to be learned about variable-viscosity convection. Similar results, although on a system that was heated from below, were found by Ogawa *et al.* (1991). A complete parametric treatment of our work on variable-viscosity convection, crystallization, and partial melting will appear elsewhere.

The results of our numerical experiments differ from the laboratory experiments of Huppert and Sparks (1988) and Campbell and Turner (1987). These deservedly influential papers evaluate a number of the fluid dynamical and petrological consequences of basaltic underplating. We agree with many of the geological implications for melt production and the origin of petrologic diversity discussed in these works. However, our continuum models yield estimates of time and length scales of magma generation and cooling that are quite different from theirs. The important differences are that Huppert and Sparks (1988) predict very turbulent conditions in the magma, leading to rapid cooling and concomitant bulk melting of the overlying country rock. The differences can be ascribed to the following: (*a*) we explicitly model strongly variable-viscosity flow, and as importantly, (*b*) the melting and solidification relations between the country rock and the model basalt in our work are like those in natural systems. We preface our discussion by noting that magmas are rarely superheated and that the melt fraction–temperature curves used in models must retain the relationships between melting intervals between basalt and country rock as shown in Fig. 1: the underplated material must have a higher liquidus temperature and a narrower and higher solidus-to-liquidus temperature range than the country rock.

The laboratory experiments of underplating of Huppert and Sparks (1988) use a superheated aqueous solution of $NaNO_3$, underplated beneath a wax (PEG 1000). We were unable to find any reference to the the actual weight percent of $NaNO_3$ for the experiments of Huppert and Sparks (1988); however, the liquidus slope of the $NaNO_3$ solution varies between 0 and -18.1°C (Kirk-Othmer, 1978). The $NaNO_3$ was substantially superheated to 70°C before underplating. The PEG wax used as the country rock has a melting interval between 37 and 40°C and at the start of the experiment was isothermal with a temperature of 20°C. Thus, in no circumstances could the underplated "basalt" solution crystallize in the models, and hence no rheological penalty for crystallization is possible. This is unlike the geological case in many important respects, where basalt crystallizes at temperatures above the ambient temperatures of the country rock. The laboratory roof melting experiments of Campbell and Turner (1987) show more complex behavior as simultaneous crystallization and melting occur in some of their experiments. However, the model basalt in their experiments is also superheated, and hence the same cautions detailed above apply. We hasten to add that the experiments of Huppert and Sparks (1988) and Campbell and Turner (1987) are well motivated and that it should be appreciated that getting the right combination of material in any analog model of underplating is very difficult.

In summary, numerical experiments using geologically relevant thermophysical properties indicate that partial melting from underplated basalt will be largely conductive, *in the early stages of underplating*. This is in agreement with many features from the rock record, including measured amounts of partial melting at plutonic contacts and the compositions of naturally occurring granitoids. We concur with the important conclusions from Huppert and Sparks (1988) regarding the thermal and compositional evolution of regions of underplating: if further underplating occurs, the region will thermally mature and reach a condition where large amounts of melt can be generated by a modest increase in temperature caused by incremental addition of magma to the region of partial melting. Neither the model presented here nor those in Huppert and Sparks (1988) address these conditions in a rigorous way. Without some geological evidence, constructing a model that could be generalized would be difficult.

Melt Collection and Ascent

Once partial melting has been initiated, a gradient in melt fraction will exist away from the underplating interface. Wickham (1987) has considered a number of the controls on the segregation and transport of magma during partial melting; we will focus largely on refinements and new paradigms. A number of processes of homogenization, segregation, and transport can occur during partial melting: convective overturn in the style of porous medium convection; magma mixing involving underlying partially molten basalt; bulk convection of the partially molten region; compaction of the matrix, yielding discrete melt bodies; and, finally, melt (and matrix) transfer by virtue of diking or diapirism.

Melt Convection in a Porous Medium

Once a partly melted region is generated, a variety of dynamical states are possible: simple convection of the melt in a rigid matrix, movement of melt and matrix due to compaction, or bulk convective motion of the region of partial melting. The first is the most simple: it is assumed that the unmelted residuum forms a rigid matrix. This may have application to magma chambers that are composed of regions of largely crystal–liquid mush, such as that now proposed for mid-ocean ridges (Nicolas *et al.*, 1993). This is appropriate for conditions where the melt fraction is below about 50%, although this parameter is very poorly constrained (Bergantz, 1990). Assuming that an interconnected network forms, which can be for as little as 2 wt% melt for a mixture of olivine and basalt powder (Daines and Richter, 1988), the motion of the melt may be considered flow in a porous medium, and the formalism for convection in a porous medium may be applied as noted by Lowell (1982). For the underplating conditions described previously, where the region undergoing partial melting is being heated from below and growing upward, the permeability will be anisotropic and heterogeneous and may well produce local reversals in density due to the presence of water, which will also affect the density. There is no theoretical formalism to predict the onset of convection for these geologically relevant conditions. However, if one assumes that permeability is homogeneous and isotropic, the model relationships of Lowell (1982) can be used. These are summarized in the following expression, which specifies the dependence of the required thickness of the melt region, $X(t)$, on other system properties,

$$X(t) \geq \frac{288\pi^3 \kappa \nu c_p}{\beta g \chi^3 b^2 L \cdot \mathrm{Ste}}, \tag{10}$$

where κ is the thermal diffusivity, ν the kinematic viscosity, c_p the specific heat, β the thermal expansion coefficient, g the scalar acceleration of gravity, b the average grain diameter, L the latent heat of melting, and Ste the Stefan number modified by multiplying the latent heat term by the melt fraction. For typical values of these quantities and a constant melt fraction of 0.3, Eq. (10) is plotted in Fig. 9. Low-viscosity melts may become homogenized over scales of a few meters to tens of meters, which is the length scale over which the model assumptions may be relevant. For more viscous melts, the layer thickness will exceed any reasonable length scale of lithologic homogeneity and convection cannot be simply characterized by an expression like (10).

Expression (10) is based on model assumptions that severely limit its application to crustal melting except as an end-member estimate. A number of additional, potentially relevant complexities have been addressed in the engineering literature: Kaviany (1984) considered the case where the lower boundary temperature is increased linearly, variable porosity was considered by David *et al.* (1991), and variable viscosity of the melt by Blythe and Simpkins (1981). Given the nonlinear nature of these effects, it is not possible to combine the results of these different studies and to assess their importance for convection in partial melts. None of the current studies considers the conditions for convection in a growing melt layer where the permeability and melt viscosity are changing as a function of temperature. In summary, it is unclear whether a distributed melt will be homogenized on anything but small scales. Whether melt homogenization occurs depends most importantly on melt viscosity, which will vary dramatically as a function of volatile content and hence melt progress.

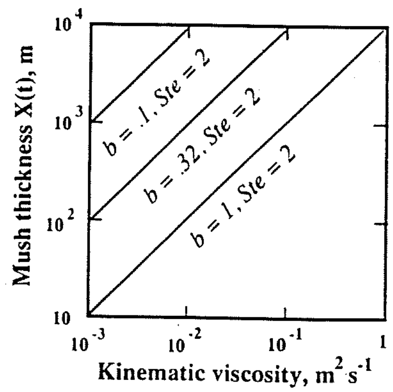

Figure 9 Critical mush thickness for porous media convection to begin as a function of kinematic viscosity. For these curves the grain diameter, b, is in centimeters. Ste is the Stefan number. After Lowell (1982).

Compaction

If the unmelted matrix is not infinitely rigid, compaction and subsequent reorganization of the partial melt region can occur. Compaction refers to the change in melt fraction that occurs as buoyant interstitial fluid moves upward and solid material is displaced downward. The governing equations that describe compaction are developed in the context of mixture theory or averaging techniques and hence the calculated values of the dependent variables represent averages at scales that are large relative to individual grains but small relative to the scale of the gradients in the dependent variables. This is the usual approach in the continuum description of porous media and mixed phase processes and does not limit the utility of the model for most problems of geologic interest. The physical processes and petrologic implications of this form of melt segregation have been considered in the context of melt migration in the mantle (Fowler, 1985, 1990a, 1990b; McKenzie, 1984, 1985; Ribe, 1985; Scott and Stevenson, 1986; Sleep, 1974), in magma chambers (Shirley, 1986, 1987), and in the crust (Fountain *et al.*, 1989; Lowell and Bergantz, 1987; McKenzie, 1985; Wickham, 1987).

A succinct summary of the equations describing compaction is given by Ribe (1987). Of interest here is the case where compaction is occurring in a region of variable partial melt and with gradients where the partial melt fraction goes to zero. Under these conditions, melt can collect into "solitary waves," dubbed solitons or magmons (Scott and Stevenson, 1984; Stevenson and Scott, 1987), where under certain conditions the melt fraction can go to 100% and fully liquid magma chambers can ostensibly form. The model of Fountain *et al.* (1989) demonstrates this in the context of melt migration during crustal anatexis following underplating by mafic magma. Provided the melt fraction at the contact exceeds 25%, it is demonstrated that compaction can yield melt bodies with dimensions on the order of kilometers extending from the contact to a few kilometers above the contact. These length scales are in rough agreement with some geological examples; however, the upward migration is limited to regions that have undergone partial melting. Figure 10 shows the calculated distribution of partial melt. Features not explicitly considered in the model of Fountain *et al.* (1989) are the specifics of thermal or compositional magmatic convection: e.g., the model is spatially one-dimensional. These authors argue that the density will be dominated by the influence of crystals falling off the roof, an ad hoc but not unreasonable assumption once the growing melt layer is itself largely crystal free. Lowell and Bergantz (1987) also consider compaction of a growing partial melt layer and conclude that the compositional effects on density may yield instabilities in the growing melt layer (see discussion on porous media convection above) and hence yield conditions in the region of compaction that would be hard to generalize and would require a multidimensional model to adequately represent the structure of the fluid and solid flow.

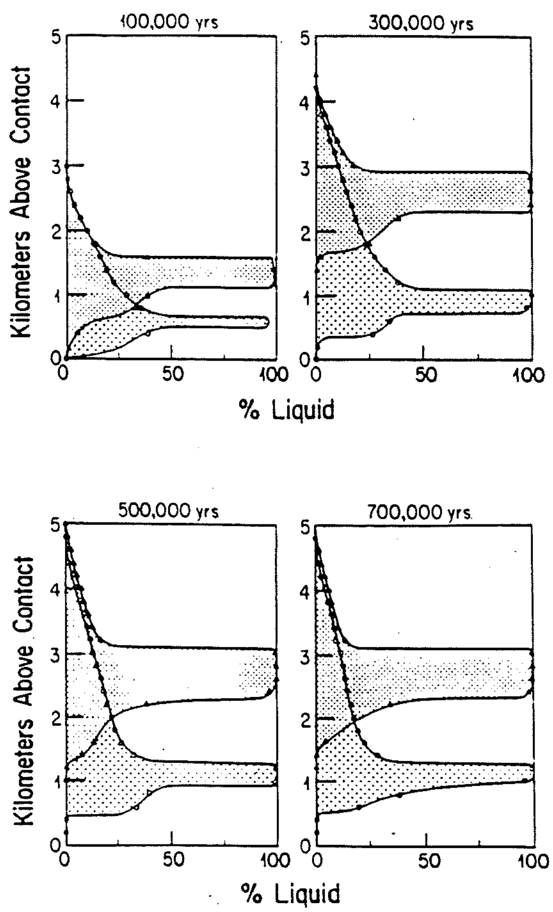

Figure 10 Distribution of granitic melt in the country rock at four different times, 100,000, 300,000, 500,000, and 700,000 yr after intrusion. The stippled area is defined by the curve of percentage liquid versus distance from the intrusion. From Fountain *et al.* (1989).

Diapirism and Diking

The crustal scale transport of magma is often considered in terms of two end members: diapirism and diking. This is somewhat surprising as there is little theoretical or unambiguous geological evidence to suggest that either of these mechanisms alone is dominant in the *crustal* transfer of silicic melts. Diapirism is a type of Rayleigh–Taylor instability and has traditionally been invoked to explain the growth and disposition of salt domes, the spacing of volcanoes (Marsh, 1979), and the ascent of granitic magma in the crust (Miller *et al.*, 1988). The conditions leading to the onset of the instability are developed by Turcotte and Schubert (1982) and more recently by Lister and Kerr (1989) who caution that variable physical properties, geometries, and the presence of a deformable lower layer render many of the more classical estimates of the wavelength and rise time of the fastest growing instability inappropriate. These conditions may be present in partial melt regions that have vertical lithological and rheological variations where a region of greater partial melt is sandwiched between regions with smaller amounts of partial melt and the viscosity contrasts are not large between the layers.

The important rate-limiting element in magmatic diapirism lies in the thermal coupling of the diapir and the surrounding rock. It is thought that heat loss from the magma to the surrounding rock lowers the viscosity of the country rock allowing it to "flow" around the magma body. Implicit in this model is that there is a finite distance that the magma can rise before the enthalpy difference between the diapir and the surroundings diminishes to the point where the density and viscosity contrasts are no longer sufficient to permit continued ascent. This is the basis of what has become known as the hot Stokes model of diapirism. The quantitative description of the model has been developed by Marsh (1982), Morris (1982), Daly and Raefsky (1985), and Mahon *et al.* (1988). It should be noted that when the hot Stokes model of diapirism was first proposed by Marsh and others, the intended application was melt transfer in the mantle.

The hot Stokes model of diapiric transfer of magma contains features that, in principle, should be testable by geologic observation (Bateman, 1984; England, 1990; Paterson *et al.*, 1991). However, the ambiguity in uniquely identifying strain fields that record diapiric ascent as opposed to the fabric generated during (final) emplacement precludes a direct interpretation in many cases. Paterson *et al.* (1991) have considered a wide variety of natural examples and conclude that the width of aureoles and the intensity of wall rock strain around granitoids are generally not as great as those required by laboratory or numerical models for diapirism. England (1990; 1992) argues that the fabrics generated during an emplacement process, such as "ballooning," may make it difficult to recognize the signature of diapirism. This condition is exacerbated if the diapir is compositionally nonuniform, because a number of mixing conditions can arise if there are multiple magmatic elements in the diapir, leading to strain localization and nonuniformity in the thermal and compositional fields. Weinberg (1992) notes that reverse zoning would be the most common pluton geometry if the zoning were controlled only by internal circulation. However, most zoned systems have a normal zonation pattern that progresses from more mafic to felsic, and it appears that an internal circulation driven by internal shear coupling may be of limited importance in generating zonation patterns. Isothermal experiments of diapirism by Cruden (1990) reveal some of the fabrics that can occur within an upward moving melt body by virtue of the viscous coupling to the country rock. The deformation is strongly time dependent and a variety of fabrics can form with penetrative foliations at the margins and a possibly isotropic fabric internally. An element not included in this or other models is the inevitable crystallization that must occur at the boundary as diapirism proceeds. This crystallization will effectively act to move the isotherm associated with internal slip inward, yielding a telescoping system where fractionation can occur by virtue of ascent, and the time history of fabrics will be even more complex than that given by Cruden (1990). These considerations cast some doubt on the notion that kinematic indicators taken from intrusions and the surrounding rock can discriminate between diapirism, as understood in the classic sense, and magma chamber development following initial transport in a dike.

Mechanical and thermal models of magma transport in dikes are considered in this text. It is important to note, however, that some models of melt transfer in dikes suggest that it is a very efficient means of magma transfer. Clemens and Mawer (1992) present a model where it is estimated that a 2000-km^3 batholith can be filled in less than 900 yr. Although this model requires some end-member assumptions, it demonstrates the kinds of time scales that may be possible if magma is transported in dikes.

Tectonic Regimes and Magma Ascent

The previous discussions considered simple systems where the material surrounding the melt was considered to be a mechanical continuum, and the role of tectonism and the rheological partitioning of the crust were not explicitly included. Conflicting and ambiguous geological evidence indicate that this approach, while amenable to modeling, is not sufficient to account for the variety of scales of melt extraction and ascent that are observed in the rock record. For example, on the basis of the compositions of migmatites and partial melts of the midcrust, Sawyer (1991) provides evidence that the rates of melting and melt extraction often are greater than the rates at which chemical equilibrium can be achieved in the melt. Thus, it may be inappropriate to model many important melt-forming reactions with an assumption of thermodynamic equilibrium. There are numerous studies that demonstrate the broadly syntectonic nature of magma ascent. Some studies even infer that a feedback exists between magma generation and tectonism. It is apparent that the material transfer required to accommodate the ascent of magma occurs simultaneously at a variety of scales: therefore, models that attempt to describe the ascent and magma chamber assembly process only on the basis of the near field deformation will necessarily be incomplete.

The possible role of crustal scale structures controlling the ascent and emplacement of magmas is typically determined by the use of kinematic indicators measured in and at the margins of plutons and the disposition of magmatic centers relative to the positions of structural features. A hypothesis linking the ascent of magmas and the development of crustal scale structural features where magmas can possibly follow up and into any zones of weakness created by faulting is attractive in that it appears, on the face of it at least, to provide a solution to the "room problem" in the near field. We are not suggesting that fault bends produce "holes" in the crust that magmas then pour into, but rather that magmas can intrude along shear zones while they are active. With this type of intrusion mechanism the need for a crustal scale ductile halo as required for diapiric rise is obviated. Gravity studies indicate that the form of granitic plutons can be correlated with the tectonic style and occurrence of shear zones: intrusion during extension yields thin plutons composed of a number of subunits and intrusion during transcurrent shear yields a few feeder zones that lie off the shear zone (Vigneresse, 1994).

The structural control of the intrusion of plutons is further suggested by the frequent occurrence of plutons in linear belts as noted by Pitcher and Bussell (1977) in the Andes. This tectonic control on the character of magmatism at convergent margins has been considered in the broadest sense by Glazner (1991) who suggests that plutonism is favored when the tangential component of convergence is large, and volcanism is favored when the normal component of the convergence vector is dominant. However, no specific mechanism for this correlation is proposed. Considering tectonic elements at a similar scale, Tikoff and Teyssier (1992) provide a model where plutonism and batholith assembly occurs in zones of dilatation that accompany the development of en echelon P-shear arrays, which result from a transpressional strain regime created by oblique convergence. The evidence used to support this is the linear distribution of intrusions that lie just off the axis of the strike-slip faults, and the occurrence of syntectonic fabrics whose orientation is coherent on a pluton scale and consistent with an origin by dextral shearing.

Mineral fabrics provide the only direct means for the determination of the kinematic conditions during ascent, emplacement, and subsequent deformation. These fabrics are defined by mineral orientations and their mechanical condition. In general, a plutonic rock can have a fabric that is the result of a sequence of stress fields, from those generated during flow in the magmatic state, to near solidus and subsolidus brittle fabrics due to regional stresses. The use of kinematics to determine the flow in magmatic rocks has been reviewed by Nicolas (1992). The criteria for the determination of those elements of a rock fabric that have an origin by magmatic flow, as opposed to solid-state deformation, are reviewed by Paterson *et al.* (1989). The criteria involve observations at the outcrop and thin section scale: alignment of near liquidus phases indicates flow where crystals are dispersed in a fluid medium. Tiling of crystals can also indicate magmatic flow with a high solid fraction. The presence of plastic strain such as kinking, undulatory extinction, and crystal regrowth indicates deformation in the solid state. When deformation is taking place near the critical melt fraction, the textures can be ambiguous as a transition from magmatic to solid-state flow may occur. Establishing the sense of shear under these conditions, which are of particular interest in mid- to deep-crustal exposures where magma generation and extraction may be occurring, is complex and the interested reader is referred to the study of Blumenfeld and Bouchez (1988) who consider deformation in migmatites and partial melts. The magnetic susceptibility anisotropy of plutonic rocks has also been used to determine flow direction when other means of determining the fabric are inadequate or ambiguous. General agreement is found between the magmatic mineral fabric and the orientation of the magnetic susceptibility anisotropy (Bouchez *et al.*, 1990).

Syntectonic mineral fabrics, that is, mineral fabrics with a component that can be ascribed to both magmatic and regional shear during solidification, provide compelling evidence that magma is, in some instances, intruding *during* tectonic movement. Crustal "openings" and bends, which are the inevitable consequence of shearing, provide access for magmas that reside at deeper crustal levels. Hutton (1988, 1992) and Hutton *et al.* (1990) review a few case histories where magma has intruded along shear zones both vertically and nearly horizontally. Antonellini and Cambray (1992) document magma transport by the process of stepping up along bedding-parallel shear zones in rift systems. Although these models of magma transfer along shear zones appear to provide a resolution to the space problem in the near field, the space must be accommodated elsewhere, ostensibly by thrusting or extension in the far field. Thus, the rate-limiting steps in magmatic ascent will involve some combination of regional stresses and local buoyancy forces related to the magma density contrast.

Among the more provocative corollaries of such models are the suggestions that shearing itself may act as an agent of melt production, or conversely, the presence of melt may act to localize shear. Karlstrom *et al.* (1993) document melt extraction in conjunction with thrusting in a small midcrustal pluton: tectonic movement was instrumental in driving melt segregation within the pluton. Hutton and Reavy (1992) propose that strike-slip transpression thickens the crust, yielding anatectic granitic magmas that will have the usual geochemical signature of crustal melts. This thickening may also yield an undulating Moho, which itself may undergo melting and mixing with mantle diapirs, yielding magmas with geochemical attributes that indicate interaction with mantle-derived materials. Thus, thrusting and transpression provide a means to generate magmas whose geochemical indicators suggest diverse origins, although additional geologic tests need to be articulated to fully test this hypothesis. The presence of melt in shear zones may also act to enhance and/or localize deformation as argued by Davidson *et al.* (1992), who combine a thermal model with kinematic indicators of shearing to demonstrate that large amounts of crustal strain can be accommodated by regions of melt. One key element in this mechanism is that the crust is at near solidus temperatures and hence the melt layer has a long thermal lifetime.

It should be appreciated that kinematic indicators interpreted from the mineral fabric record the last episode of the crystal–melt organization in magma. They record the end stages of the process, whether one of ascent or emplacement, and as noted in the discussion on diapirism, they may record mechanical conditions quite different than those that existed at the time of melt extraction and ascent. Nonetheless, these fabrics are the only structural information directly available.

The tectonic control of the distribution of volcanic vents at both regional and local scales is evident, although complex, perhaps due to the redirection of regional stress fields in the near surface. This is well documented in the study of Bacon (1985) who also notes the correlation between earthquakes and the timing of eruption. More complex relationships between the clustering of cinder cones and the presence of fault zones are recognized in Mexico (Conner, 1990). Vent clustering was found to be pervasive and the azimuths of the clusters appear to be dominantly controlled by the convergence direction, and less so by the orientation of faults. The long-standing conundrum in this regard is the spacing and linear arrangements of volcanic centers in arcs (Jarrard, 1986; Marsh, 1979; Sherrod and Smith, 1990). The positions of these centers often persist for millions of years, erupting a variety of magma types with no clear association with faulting, suggesting instead a control due to processes in the magma supply region. Or it may be that the crust is sufficiently fractured that magmas can at any time exploit an available shear zone, and so the connection between magmatism and shear zones potentially tells us little about the rate-limiting elements in crustal magmatism.

Thus, the distribution and disposition of plutons often indicate some form of structural control, while volcanism reflects structural control in some cases and not in others. One feature that both plutonic and volcanic systems do share is some degree of open system behavior. This is expressed in plutons as repeated intrusion and pluton assembly in an incremental fashion (Harry and Richey, 1963) which can yield a wide variety of zoning patterns (Bergantz, 1990) as well as chemical evidence of magma mixing. In the volcanic record it is expressed by the diversity of magmas erupting from a single center: as individual units often unrelated geochemically in any simple way, or mingling with adjacent magma bodies as demonstrated in the spectacularly exposed Quaternary arc volcano, Tatara–San Pedro in the Chilean Andes (Dungan, 1992).

It appears then that the time-integrated geodynamic and petrologic expression of magmatism might be that of a crustal scale conduit rather than a single chamber (Singer *et al.*, 1989). Implicit in this conduit model is that magmatic systems are assembled from the bottom up, which yields a downward and outward flaring thermal anomaly. The region from which magmatic contributions can be generated by partial melting is potentially much larger at the bottom than at the top and petrologic diversity is generated in this case by crystal–liquid separation by virtue of ascent and open system processes. The style of ascent will vary from processes that are ductile in the deep crust to diking and intrusion along fault-generated bends in the upper crust as rheological conditions will migrate upward with the thermal anomalies. As the thermal anomaly that is driving magmatism in the deep crust migrates upward, it will "consume" the geological evidence of the early stages of magma generation, ascent, and assembly. The end result may be a thermal anomaly on a crustal scale like that interpreted from compressional wave velocity studies at Long Valley, California (Dawson *et al.*, 1990). Geological evidence of the early stages of this process would have to be found in regions where this process has failed, for example, an immature episode of underplating as discussed previously. If melt is generated by anatexis during transpressive orogenesis, a similar style of melt generation and ascent may result. D'Lemos *et al.* (1992) document these transitions in intrusive style with rheological state on a crustal scale where the important physical elements linking regions of melt generation and final pluton formation are "megadikes."

Given the temporal complexity in the spectrum of chemical compositions within single centers, and the presence of intermediate magmas in the midcrust, it is clear that we need to move beyond paradigms of chemical petrology that rely solely on the concept of parental magma chambers as sites for the generation of petrologic diversity. Nor is it clear that simple mechanical models adequately express the many scales at which magma generation, extraction, and ascent occur simultaneously. As discussed previously, basaltic magmatism is a ubiquitous and seemingly necessary element in the generation of crustal melts in some tectonic regimes and represents the first of the likely connections between the mantle and crustal magma chambers. As this "master" perturbation propagates upward in the crust, the conditions and style of magmatic ascent will change as the temperature field changes both spatially and temporally. The success of the methods taken from structural geology in illustrating some of the kinematic aspects of magma ascent and the connection between rheological conditions and temperature suggest that a geodynamic approach to the quantitative description of magmatic processes may provide the missing link in developing scientifically sound models of crustal scale magma genesis.

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References

Antonellini, M. A., and Cambray, F. W. (1992). Relations between sill intrusions and bedding-parallel extensional shear zones in the Mid-continent Rift System of the Lake Superior region, *Tectonophysics* **212**, 331–349.

Arndt, N. T., and Goldstein, S. L. (1989). An open boundary between lower continental crust and mantle: Its role in crust formation and crustal recycling. *Tectonophysics* **161**, 201–212.

Ashworth, J. R., and Brown, M. (1990). An overview of diverse responses to diverse processes at high crustal temperatures. *in* "High-Temperature Metamorphism and Crustal Anatexis" (J. R. Ashworth and M. Brown, eds.), pp. 1–18. Unwin Hyman, London.

Bacon, C. R. (1985). Implications of silicic vent patterns for the presence of large crustal magma chambers. *J. Geophys. Res.* **90**, 11243–11252.

Bacon, C. R. (1992). Partially melted granodiorite and related rocks ejected from Crater Lake caldera, Oregon, *Trans. Roy. Soc. Edinburgh* **83**, 27–47.

Barnes, H. A., Hutton, J. F., and Walters, K. (1989). "An Introduction to Rheology," Elsevier, New York.

Bateman, R. (1984). On the role of diapirism in the segregation, ascent and final emplacement of granitoid magma, *Tectonophysics* **110**, 211–231.

Bea, F. (1991). Geochemical modeling of low melt-fraction anatexis in a peraluminous system: The Pena Negra Complex (central Spain), *Geochem. Cosmochim. Acta* **55**, 1859–1874.

Beard, J. S., and Lofgren, G. E. (1989). Effects of water on the composition of partial melts of greenstone and amphibolite. *Science* **244**, 195–197.

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). Double-diffusive convection during dendritic solidification of a binary mixture. *Physicochem. Hydrodyn.* **10**, 195–213.

Beckermann, C., and Viskanta, R. (1993). Mathematical modeling of transport phenomena during alloy solidification. *Appl. Mech. Rev.* **46**, 1–27.

Bennon, W. D., and Incropera, F. P. (1987). A continuum model for momentum, heat and species transport in binary solid–liquid phase change systems. I. Model formulation. *Int. J. Heat Mass Trans.* **30**, 2161–2170.

Bergantz, G. W. (1989a). Thermal and dynamical state of the crust following underplating: Implications for melt generation and petrologic diversity. *EOS Trans. Am. Geophys. Union* **70**, 1320.

Bergantz, G. W. (1989b). 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* "Modern Methods of Igneous Petrology: Understanding Magmatic Processes" (J. Nicholls and J. K. Russell, eds.), pp. 240–257. Mineralogical Society of America.

Bergantz, G. W. (1992). Conjugate solidification and melting in multicomponent open and closed systems. *Int. J. Heat Mass Trans.* **35**, 533–543.

Blumenfeld, P., and Bouchez, J. L. (1988). Shear criteria in granite and migmatite deformed in the magmatic and solid stages. *J. Struct. Geol.* **10**, 361–371.

Blythe, P. A., and Simpkins, P. G. (1981). Convection in a porous layer for a temperature dependent viscosity. *Int. J. Heat Mass Trans.* **24**, 497–506.

Bohlen, S. R., and Mezger, K. (1989). Origin of granulite terranes and the formation of the lowermost continental crust. *Science* **244**, 326–329.

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

Bouchez, J.-L., Gleizes, G., Djouadi, T., and Rochette, P. (1990). Microstructure and magnetic susceptibility applied to emplacement kinematics of granites: The example of the Foix pluton (French Pyrenees). *Tectonophysics* **184**, 157–171.

Brandeis, G., and Jaupart, C. (1986). On the interaction between convection and crystallization in cooling magma chambers. *Earth Planet. Sci. Lett.* **77**, 345–361.

Brown, G. C., and Fyfe, W. S. (1970). The production of granitic melts during ultrametamorphism. *Contrib. Mineral. Petrol.* **28**, 310–318.

Buck, W. R., and Parmentier, E. M. (1986). Convection beneath young oceanic lithosphere: implications for thermal structure and gravity. *J. Geophys. Res.* **91**, 1961–1974.

Burnham, C. W. (1967). Hydrothermal fluids at the magmatic stage. *in* "Geochemistry of Hydrothermal Ore Deposits" (H. L. Barnes, ed.), pp. 34–76, Holt, Rinehart & Winston, New York.

Burnham, C. W. (1979). Magmas and hydrothermal fluids. *in* "Geochemistry of Hydrothermal Ore Deposits" (H. L. Barnes, ed.), 2nd ed., pp. 71–136. Wiley, New York.

Burnham, C. W. (1992). Calculated melt and restite compositions of some Australian granites. *Trans. Roy. Soc. Edinburgh* **83**, 387–397.

Campbell, I. H., and Turner, J. S. (1987). A laboratory investigation of assimilation at the top of a basaltic magma chamber. *J. Geol.* **95**, 155–172.

Cashman, K. V. (1993). Relationship between plagioclase crystallization and cooling rate in basaltic melts. *Contrib. Mineral. Petrol.* **113**, 126–142.

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

Chappell, B. W., and White, A. J. R. (1992). I- and S-type granites of the Lachlan Fold Belt. *Trans. Roy. Soc. Edinburgh* **83**, 1–26.

Chappell, B. W., White, A. J. R., and Wyborn, D. (1987). The importance of residual source material (restite) in granite petrogenesis. *J. Petrol.* **28**, 1111–1138.

Chen, Y.-M., and Pearlstein, A. J. (1988). Onset of convection in variable viscosity fluids: Assessment of approximate viscosity–temperature relations. *Phys. Fluids* **31**, 1380–1385.

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**, 17757–17785.

Cheng, D. C.-H. (1984). Further observations on the rheological behavior of dense suspensions. *Powder Tech.* **37**, 255–273.

Chu, T. Y., and Hickox, C. E. (1990). Thermal convection with large viscosity variation in an enclosure with localized heating. *J. Heat Trans.* **112**, 388–395.

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

Clemens, J. D. (1990). The granulite–granite connection. *in* "Granulites and Crustal Evolution" (D. Vielzeuf and P. Vidal, eds.), pp. 25–36. Kluwer, Dordrecht.

Clemens, J. D., and Mawer, C. K. (1992). Granitic magma transport by fracture propagation. *Tectonophysics* **204**, 339–360.

Clemens, J. D., and Vielzeuf, D. (1987). Constraints on melting and magma production in the crust. *Earth Planet. Sci. Lett.* **86**, 287–306.

Conner, C. B. (1990). Cinder cone clustering in the Trans-Mexican volcanic belt: Implications for structural and petrologic models. *J. Geophys. Res.* **95**, 19395–19405.

Crisp, J. A. (1984). Rates of magma emplacement and volcanic output. *J. Volc. Geotherm. Res.* **20**, 177–211.

Cruden, A. R. (1990). Flow and fabric development during the diapiric rise of magma. *J. Geol.* **98**, 681–698.

Cull, J. P. (1990). Underplating of the crust and xenolith geotherms in Australia. *Geophys. Res. Lett.* **17**, 1133–1136.

D'Lemos, R. S., Brown, M., and Strachan, R. A. (1992). Granite magma generation, ascent and emplacement within a transpressional orogen. *J. Geol. Soc. London* **149**, 487–490.

Daines, M. J., and Richter, F. M. (1988). An experimental method for directly determining the interconnectivity of melt in a partially molten system, *Geophys. Res. Lett.* **15**, 1459–1462.

Daly, S. F., and Raefsky, A. (1985). On the penetration of a hot diapir through a strongly temperature dependent medium, *Geol. J. Roy. Astron. Soc.* **83**, 657–682.

David, E., Lauriat, G., and Cheng, P. (1991). A numerical solution of variable porosity effects on natural convection in a packed-sphere cavity, *J. Heat Trans.* **113**, 391–399.

Davidson, C., Hollister, L., and Schmid, S. M. (1992). Role of melt in the formation of a deep crustal compressive shear zone: The Maclaren Glacier metamorphic belt, south central Alaska, *Tectonics* **11**, 348–359.

Davidson, J. P., McMillian, N. J., Moorbath, S., Worner, G., Harmon, R. S., and Lopez-Escobar, L. (1990). The Nevados de Payachata volcanic region, II. Evidence for widespread crustal involvement in Andean magmatism. *Contrib. Mineral. Petrol.* **105**, 412–432.

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**, 11021–11050.

Dell'Angelo, L. N., and Tullis, J. (1988). Experimental deformation of partially melted granitic aggregates, *J. Meta. Geol.* **6**, 495–515.

DePaolo, D. J., Perry, F. V., and Baldridge, W. S. (1992). Crustal versus mantle sources of granitic magmas: A two-parameter model based on Nd isotopic studies, *Trans. Roy. Soc. Edinburgh* **83**, 439–446.

Drummond, B. J., and Collins, C. D. N. (1986). Seismic evidence for underplating of the lower continental crust in Australia, *Earth Planet. Sci. Lett.* **79**, 361–372.

Dunbar, N. W., Riciputi, L. R., Jacobs, G. K., and Christie, W. (1993). Generation of rhyolitic melt in an artificial magma: Implications for fractional crystallization processes in natural magmas, *J. Volc. Geotherm. Res.*, **57**, 157–166.

Dungan, M. (1992). The life history of an Andean volcano, *EOS Trans. Am. Geophys. Union* **73**, 406–407.

England, R. W. (1990). The identification of granitic diapirs, *J. Geol. Soc. London* **147**, 931–933.

England, R. W. (1992). The genesis, ascent, and emplacement of the Northern Arran Granite, Scotland: Implications for granite diapirism, *Geol. Soc. Am. Bull.* **104**, 606–614.

Feeley, T. C., and Grunder, A. L. (1991). Mantle contribution to the evolution of middle tertiary silicic magmatism during early stages of extension: The Egan Range volcanic complex, east-central Nevada, *Contrib. Mineral. Petrol.* **106**, 154–169.

Fountain, J. C., Hodge, D. S., and Shaw, R. P. (1989). Melt segregation in anatectic granites: A thermomechanical model, *J. Volc. Geotherm. Res.* **39**, 279–296.

Fowler, A. C. (1985). A mathematical model of magma transport in the asthenosphere, *Geophys. Astrophys. Fluid Dyn.* **33**, 63–96.

Fowler, A. C. (1990a). A compaction model for melt transport in the Earth's asthenosphere. Part I. The basic model, *in* "Magma Transport and Storage" (M. P. Ryan, ed.), pp. 3–14, Wiley, Chichester/Sussex, England.

Fowler, A. C. (1990b). A compaction model for melt transport in the Earth's asthenosphere. Part II. Applications, *in* "Magma Transport and Storage" (M. P. Ryan, ed.), pp. 15–32, Wiley, Chichester/Sussex, England.

Frith, W. J., Mewis, J., and Strivens, T. A. (1987). Rheology of concentrated suspensions: Experimental investigations, *Powder Tech.* **51**, 27–34.

Fuis, G. S., and Plafker, G. (1991). Evolution of deep structure along the trans-Alaska crustal transect, Chugach Mountains and Copper River basin, Southern Alaska, *J. Geophys. Res.* **96**, 4229–4253.

Ghiorso, M. S. (1985). Chemical mass transfer in magmatic systems. I. Thermodynamic relations and numerical algorithms, Contrib. Mineral. Petrol. **90**, 107–120.

Gill, J., and Condomines, M. (1992). Short-lived radioactivity and magma genesis, *Science* **257**, 1368–1376.

Glazner, A. F. (1991). Plutonism, oblique subduction, and continental growth: An example from the Mesozoic of California, *Geology* **19**, 784–786.

Grissom, G. C., DeBari, S. M., Page, S. P., Page, R. F. N., Villar, L. M., Coleman, R. C., and Ramirez, M. V. (1991). The deep crust of an early Paleozoic arc: The Sierra de Fimbala, northwestern Argentina, *in* "Andean Magmatism and Its Tectonic Setting" (R. S. Harmon and C. W. Rapela, eds.), Special Paper 265, pp. 189–200, Geol. Soc. Am., Boulder, CO.

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, northern California: Decoupling of heat and mass transfer, *Contrib. Mineral. Petrol.* **99**, 320–343.

Halliday, A. N., Mahood, G. A., Holden, P., Metz, J. M., Dempster, T. J., and Davidson, J. P. (1989). Evidence for long residence times of rhyolitic magma in the Long Valley magmatic system: The isotope record in precaldera lavas of Glass Mountain, *Earth Planet. Sci. Lett.* **94**, 274–290.

Handy, M. R., and Zingg, A. (1991). The tectonic and rheological evolution of an attenuated section of the continental crust: Ivrea crustal section, southern Alps, northwestern Italy and southern Switzerland, *Geol. Soc. Am. Bull.* **103**, 236–253.

Harry, W. T., and Richey, J. E. (1963). Magmatic pulses in the emplacement of plutons, *Liverpool Manchester Geol. J.* **3**, 254–268.

Hildreth, W. (1981). Gradients in silicic magma chambers: Implications for lithospheric magmatism, *J. Geophys. Res.* **86**, 10153–10192.

Hildreth, W. (1987). 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., 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., Halliday, A. N., and Christiansen, R. L. (1991). Isotopic and chemical evidence concerning the genesis and contamination of basaltic and rhyolitic magma beneath the Yellowstone Plateau volcanic field, *J. Petrol.* **32**, 63–138.

Hildreth, W., and Moorbath, S. (1988). Crustal contributions to arc magmatism in the Andes of Central Chile, *Contrib. Mineral. Petrol.* **98**, 455–489.

Hodge, D. S. (1974). Thermal model for the origin of granitic batholiths, *Nature* **251**, 297–299.

Hon, R., and Weill, D. (1982). Heat balance of basaltic intrusion vs granitic fusion in the lower crust. *EOS Trans. Am. Geophys. Union* **63**, 470.

Huppert, H., and Sparks, R. S. J. (1991). Comments on "On convective style and vigor in sheet-like magma chambers" by Bruce D. Marsh. *J. Petrol.* **32**, 851–854.

Huppert, H. E., and Sparks, R. S. J. (1988). The generation of granitic magmas by intrusion of basalt into continental crust. *J. Petrol.* **29**, 599–624.

Hutton, D. H. W. (1988). Granite emplacement mechanisms and tectonic controls: Inferences from deformation studies. *Trans. Roy. Soc. Edinburgh* **79**, 245–255.

Hutton, D. H. W. (1992). Granite sheeted complexes: Evidence for the dyking ascent mechanism. *Trans. Roy. Soc. Edinburgh* **83**, 377–382.

Hutton, D. H. W., Dempster, T. J., Brown, P. E., and Becker, S. D. (1990). A new mechanism of granite emplacement: Intrusion in active extensional shear zones. *Nature* **343**, 452–455.

Hutton, D. H. W., and Reavy, R. J. (1992). Strike-slip tectonics and granite petrogenesis. *Tectonics* **11**, 960–967.

Jacobs, G. K., Dunbar, N. W., Naney, M. T., and Williams, R. T. (1992). Petrologic and geophysical studies of an artificial magma. *EOS Trans. Am. Geophys. Union* **73**, 411–412.

Jarrard, R. D. (1986). Relations among subduction parameters. *Rev. Geophys.* **24**, 217–284.

Jaupart, C., and Parsons, B. (1985). Convective instabilities in a variable viscosity fluid cooled from above. *Phys. Earth Planet. Inter.* **39**, 14–32.

Jomha, A. I., Merrington, A., Woodcock, L. V., Barnes, H. A., and Lips, A. (1991). Recent developments in dense suspension rheology. *Powder Tech.* **65**, 343–370.

Karlstrom, K. E., Miller, C. F., Kingsbury, J. A., and Wooden, J. L. (1993). Pluton emplacement along an active ductile thrust zone, Piute Mountains, southeastern California: Interaction between deformational and solidification processes. *Geol. Soc. Am. Bull.* **105**, 213–230.

Kaviany, M. (1984). Onset of thermal convection in a saturated porous medium: Experiment and analysis. *Int. J. Heat Mass Trans.* **27**, 2101–2110.

Kay, R. W., and Kay, S. M. (1991). Creation and destruction of lower continental crust, *Geol. Rund.* **80**, 259–278.

Kay, S. M., Kay, R. W., Citron, G. P., and Perfit, M. R. (1990). Calc-alkaline plutonism in the intra-oceanic Aleutian arc, Alaska, *in* "Plutonism from Antarctica to Alaska" (S. M. Kay and C. W. Rapella, eds.), Special Paper 241, pp. 233–255, Geol. Soc. Am., Boulder, CO.

Kempton, P. D., and Harmon, R. S. (1992). Oxygen isotope evidence for large-scale hybridization of the lower crust during magmatic underplating, *Geochem. Cosmochim. Acta* **56**, 971–986.

Kerrick, D. M. (1991). Contact metamorphism. *Rev. Mineral.* **26**, 847.

Kirk-Othmer (1978). "Encyclopedia of Chemical Technology," Vol. 21, 3rd ed., Wiley, New York.

Klemperer, S. L., Hauge, T. A., Hauser, E. C., Oliver, J. E., and Potter, C. J. (1986). The Moho in the northern Basin and Range province, Nevada, along the COCORP 40°N seismic-reflection transect, *Geol. Soc. Am. Bull.* **97**, 603–618.

Lachenbruch, A. H., and Sass, J. H. (1978). Models of an extending lithosphere and heat flow in the Basin and Range province. *Geol. Soc. Am. Bull. Mem.* **152**, 209–250.

Lange, R. A., and Carmichael, I. S. E. (1990). Hydrous basaltic andesites associated with minette and related lavas in western Mexico. *J. Petrol.* **31**, 1225–1259.

Leeman, W. P. (1982). Rhyolites of the Snake River Plain–Yellowstone Plateau province, Idaho and Wyoming: A summary of petrogenetic models, *in* "Cenozic Geology of Idaho" (B. Bonnichsen and R. M. Breckenridge, eds.), pp. 203–212, Idaho Geological Survey.

Lister, J. R., and Kerr, R. C. (1989). The effect of geometry on the gravitational instability of a buoyant region of viscous fluid. *J. Fluid Mech.* **202**, 577–594.

Lowell, R. P. (1982). Thermal convection in magmas generated by hot-plate heating. *Nature* **300**, 253–254.

Lowell, R. P., and Bergantz, G. W. (1987). Melt stability and compaction in a partially molten silicate layer heated from below, *in* "Structure and Dynamics of Partially Solidified Systems" (D. E. Loper, ed.), pp. 383–400, Martinus Nijhoff, Dordrecht.

Mahon, K. I., Harrison, T. M., and Drew, D. A. (1988). Ascent of a granitoid diapir in a temperature varying medium. *J. Geophys. Res.* **93**, 1174–1188.

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. Lett.* **99**, 395–399.

Manduca, C. A., Silver, L. T., and Taylor, H. P. (1992). $^{87}Sr/^{86}Sr$ and $^{18}O/^{16}O$ isotopic systematics and geochemistry of granitoid plutons across a steeply-dipping boundary between contrasting lithospheric blocks in western Idaho. *Contrib. Mineral. Petrol.* **109**, 355–372.

Mareschal, J.-C., and Bergantz, G. B. (1990). Constraints on thermal models of the Basin and Range province. *Tectonophysics* **174**, 137–146.

Marsh, B. D. (1979). Island arc development: Some observations, experiments and speculations. *J. Geol.* **87**, 687–713.

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. (1982). On the mechanics of igneous diapirism, stoping and zone melting. *Am. J. Sci.* **282**, 808–855.

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

Marsh, B. D. (1991). Reply to comments of Huppert and Sparks. *J. Petrol.* **32**, 855–860.

McKenzie, D. (1984). The generation and compaction of partially molten rock. *J. Petrol.* **25**, 713–765.

McKenzie, D. (1985). The extraction of magma from crust and mantle. *Earth Planet. Sci. Lett.* **74**, 81–91.

McKenzie, D., and Bickle, M. J. (1988). The volume and composition of melt generated by extension of the lithosphere. *J. Petrol.* **29**, 625–679.

Mereu, R. F., Mueller, S., and Fountain, D. M. (1989). "Properties and Processes of Earth's Lower Crust," Geophysical Monograph 51, Amer. Geophys. Union.

Miller, C. F., Hanchar, J. M., Wooden, J. L., Bennett, V. C., Harrison, T. M., Wark, D. A., and Foster, D. A. (1992). Source region of a granite batholith: Evidence from lowe crustal xenoliths and inherited accessory minerals. *Trans. Roy. Soc. Edinburgh* **83**, 49–62.

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

Morris, S. (1982). The effects of strongly temperature dependent viscosity on slow flow past a hot sphere. *J. Fluid Mech.* **124**, 1–26.

Nelson, K. D. (1991). A unified view of craton evolution motivated by recent deep seismic reflection and refaction results. *Geophys. J. Int.* **105**, 25–35.

Ni, J., and Beckermann, C. (1991). A volume-averaged twophase model for transport phenomena during solidification. *Metall. Trans. B* **22**, 349–361.

Nicolas, A. (1992). Kinematics in magmatic rocks with special reference to gabbros. *J. Petrol.* **33**, 891–915.

Nicolas, A., Freydier, C., Godard, M., and Vauchez, A. (1993). Magma chambers at oceanic ridges: How large? *Geology* **21**, 53–56.

Noyes, H. J., Wones, D. R., and Frey, F. (1983). A tale of two plutons: Petrographic and mineralogic constraints on the petrogenesis of the Red Lake and Eagle Peak plutons, Central Sierra Nevada, California. *J. Geol.* **91**, 353–379.

Ogawa, M., Schubert, G., and Zebib, A. (1991). Numerical simulations of three-dimensional thermal convection with strongly temperature-dependent viscosity. *J. Fluid Mech.* **233**, 299–328.

Oldenburg, C. M., and Spera, F. J. (1991). Numerical modeling of solidification and convection in a viscous pure binary eutectic system. *Int. J. Heat Mass Trans.* **34**, 2107–2121.

Oldenburg, C. M., and Spera, F. J. (1992). Hybrid model for solidification and convection. *Numer. Heat Trans. B* **21**, 217–229.

Olson, P., Schubert, G., Anderson, C., and Goldman, P. (1988). Plume formation and lithosphere erosion: A comparison of laboratory and numerical experiments. *J. Geophys. Res.* **93**, 15065–15084.

Paterson, B. A., Stephens, W. E., Rogers, G., Williams, I. S., Hinton, R. W., and Herd, D. A. (1992). The nature of zircon inheritance in two granite plutons. *Trans. Roy. Soc. Edinburgh* **83**, 459–472.

Paterson, S. R., Vernon, R. H., and Fowler, T. K. (1991). Aureole tectonics. *in* "Contact Metamorphism" (D. M. Kerrick, ed.), Reviews in Mineralogy, Vol. 26, pp. 673–722.

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 for the origin of peraluminous granitoids and aluminous granulites. *Contrib. Mineral. Petrol.* **107**, 202–218.

Peterson, J. W., and Newton, R. C. (1990). Experimental biotite-quartz melting in KMASH-CO_2 system and the role of CO_2 in the petrogenesis of granites and related rocks. *Am. Min.* **75**, 1029–1042.

Pitcher, W. S., and Bussell, M. A. (1977). Structural control of batholith emplacement in Peru: A review. *J. Geol. Soc. London* **133**, 249–256.

Powell, R. (1983). Processes in granulite-facies metamorphism. *in* "Migmatites, Melting, and Metamorphism" (M. P. Atherton and C. D. Gribbs, eds.), pp. 127–139, Shiva, Cheshire.

Quick, J. E., Sinigoi, S., Negrini, L., Demarchi, G., and Mayer, A. (1992). Synmagmatic deformation in the underplated igneous complex of the Ivrea–Verbano zone. *Geology* **20**, 613–616.

Reagan, M. K., Gill, J. B., Malavassi, E., and Garcia, M. O. (1987). Changes in magma composition at Arenal volcano, Costa Rica, 1968–1985: Real time monitoring of open system differentiation. *Bull. Volc.* **49**, 415–434.

Reagan, M. K., Herrstrom, E. A., and Murrell, M. T. (1991). The time scale of magma generation in Nicaragua from ^{238}U-series nuclide abundances. *Geol. Soc. Am. Abst.* A114.

Ribe, N. M. (1985). The generation and composition of partial melts in the Earth's mantle. *Earth Planet. Sci. Lett.* **73**, 361–376.

Ribe, N. M. (1987). Theory of melt segregation—A review. *J. Volc. Geotherm. Res.* **33**, 241–253.

Richter, F. M., Nataf, H.-C., and Daly, S. F. (1983). Heat transfer and horizontally averaged temperature of convection with large viscosity variation. *J. Fluid Mech.* **129**, 173–192.

Rubie, D. C., and Brearley, A. J. (1990). A model for rates of disequilibrium melting during metamorphism. *in* "High-Temperature Metamorphism and Anatexis" (J. R. Ashworth and M. Brown, eds.), pp. 57–86, Unwin Hyman, London.

Rudnick, R. L. (1990). Growing from below. *Nature* **347**, 711–712.

Rudnick, R. L. (1992). Restites, Eu anomalies, and the lower continental crust. *Geochem. Cosmo.* **56**, 963–970.

Rudnick, R. L., and Taylor, S. R. (1987). The composition and petrogenesis of the lower crust: A xenolith study. *J. Geophys. Res.* **92**, 13981–14005.

Rushmer, T. (1991). Partial melting of two amphibolites: Contrasting experimental results under fluid-absent conditions. *Contrib. Mineral. Petrol.* **107**, 41–59.

Rutter, M. J., and Wyllie, P. J. (1988). Melting of vapor-absent tonalite at 10 kbar to simulate dehydration-melting in the deep crust. *Nature* **331**, 159–160.

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 (10^5 Pascals) Pressure and Higher Pressures," U.S. Geological Survey Bulletin 1764.

Sawyer, E. W. (1991). Disequilibrium melting and the rate of melt-residuum separation during migmatization of mafic rocks from the Grenville Front, Quebec. *J. Petrol.* **32**, 701–738.

Scott, D. R., and Stevenson, D. J. (1984). Magma solitons. *Geophys. Res. Lett.* **11**, 1161–1164.

Scott, D. R., and Stevenson, D. J. (1986). Magma ascent by porous flow. *J. Geophys. Res.* **91**, 2973–2988.

Sengun, M. Z., and Probstein, R. F. (1989). High-shear-limit viscosity and the maximum packing fraction in concentrated monomodal suspensions. *Physicochem. Hydrodyn.* **11**, 229–241.

Shaw, H. R. (1972). Viscosities of magmatic liquids: An empirical method of prediction. *Am. J. Sci.* **272**, 870–893.

Sherrod, D. R., and Smith, J. G. (1990). Quaternary extrusion rates of the Cascade Range, Northwestern United States and Southern British Columbia. *J. Geophys. Res.* **95**, 19465–19474.

Shirley, D. N. (1986). Compaction of igneous cumulates. *J. Geol.* **94**, 795–809.

Shirley, D. N. (1987). Differentiation and compaction in the Palisades Sill, New Jersey. *J. Petrol.* **28**, 835–865.

Singer, B. S., Myers, J. D., Linneman, S. R., and Angevine, C. L. (1989). The thermal history of ascending magma diapirs and the thermal and physical evolution of magmatic conduits. *J. Volc. Geotherm. Res.* **37**, 273–289.

Sleep, N. H. (1974). Segregation of magma from a mostly crystalline mush. *Geol. Soc. Am. Bull.* **85**, 1225–1232.

Smith, M. K. (1988). Thermal convection during the directional solidification of a pure liquid with variable viscosity. *J. Fluid Mech.* **188**, 547–570.

Stevenson, D. J., and Scott, D. R. (1987). Melt migration in deformable media, *in* "Structure and Dynamics of Partially Solidified Systems" (D. E. Loper, ed.), pp. 401–416, Martinus Nijhoff, Dordrecht.

Tepper, J. H., Nelson, B. K., Bergantz, G. W., and Irving, A. J. (1993). Petrology of the Chilliwack Batholith, North Cascades, Washington: Generation of calc-alkaline granitoids by melting of mafic lower crust with variable water fugacity. *Contrib. Mineral. Petrol.* **113**, 333–351.

Thompson, A. B. (1982). Dehydration melting of pelitic rocks and the generation of H_2O undersaturated granitic liquids. *Am. J. Sci.* **282**, 1567–1595.

Thompson, A. B. (1983). Fluid-absent metamorphism. *J. Geol. Soc. London* **140**, 533–547.

Thompson, A. B. (1988). Dehydration melting of crustal rocks. *Rend. Soc. Ital. Mineral. Petrol.* **43**, 41–60.

Thompson, A. B., and Algor, J. R. (1977). Model systems for anatexis of pelitic rocks. I. Theory of melting reactions in the system $KAlO_2$–$NaAlO_2$–SiO_2–H_2O. *Contrib. Mineral. Petrol.* **63**, 247–269.

Thompson, A. B., and Tracy, R. J. (1979). Model systems for the anatexis of pelitic rocks. II. Facies series melting reactions in the system CaO–$KAlO_2$–$NaAlO_2$–Al_2O_3–H_2O. *Contrib. Mineral. Petrol.* **70**, 429–438.

Tikoff, B., and Teyssier, C. (1992). Crustal-scale, en echelon "P-shear" tensional bridges: A possible solution to the batholithic room problem. *Geology* **20**, 927–930.

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

Trent, D. S., and Eyler, E. E. (1991). "TEMPEST: A Computer Program for Three-Dimensional Time-Dependent Hydrothermal Analysis. Vol. 1. Users Manual," Battelle, Pacific Northwest Laboratory publication PNL-4348, Vol. 1, Rev. 3.

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

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. Am. Memoir* **74**.

Valasek, P. A., Hawman, R. B., Johnson, R. A., and Smithson, S. B. (1987). Nature of the lower crust and moho in eastern Nevada from "wide angle" reflection measurements. *Geophys. Res. Lett.* **14**, 1111–1114.

Vielzeuf, D., and Holloway, J. R. (1988). Experimental determination of the fluid-absent melting relations in the pelitic system. *Contrib. Mineral. Petrol.* **98**, 257–276.

Vielzeuf, D., and Vidal, P. (1990). "Granulites and Crustal Evolution," NATO ASI, Series C, Vol. 311, Kluwer, Dordrecht.

Vigneresse, J.-L. (1994). Granite emplacement and the regional deformation field. *Tectonophysics*, in press.

Voshage, H., Hofmann, A. W., Mazzucchelli, M., Rivalenti, G., Sinigoi, S., Raczek, I., and Demarchi, G. (1990). Isotopic evidence from the Ivrea Zone for a hybrid lower crust formed by magmatic underplating. *Nature* **347**, 731–736.

Weinberg, R. F. (1992). Internal circulation in a buoyant two-fluid Newtonian sphere: Implications for composed magmatic diapirs. *Tectonophysics* **110**, 77–94.

Wells, P. R. A. (1980). Thermal models for the magmatic accretion and subsequent metamorphism of continental crust. *Earth Planet. Sci. Lett.* **46**, 253–265.

Wickham, S. M. (1987). The segregation and emplacement of granitic magmas. *J. Geol. Soc. London* **144**, 281–297.

Wickham, S. M., and Peters, M. T. (1992). Oxygen and carbon isotope profiles from Lizzies Basin, East Humboldt Range, Nevada: Constraints on mid-crustal metamorphic and magmatic volatile fluxes. *Contrib. Mineral. Petrol.* **112**, 46–65.

Wildemuth, C. R., and Williams, M. C. (1984). Viscosity of suspensions modeled with a shear-dependent maximum packing fraction. *Rheol. Acta* **23**, 627–635.

Wolf, M. B., and Wyllie, P. J. (1989). The formation of tonalitic liquids during the vapor-absent partial melting of amphibolite at 10 kbar. *EOS Trans. Am. Geophys. Union* **70**, 506.

Wolf, M. B., and Wyllie, P. J. (1991). Dehydration-melting of solid amphibolite at 10 kbar: Textural development, liquid interconnectivity and applications to the segregation of magmas. *Mineral. Petrol.* **44**, 151–179.

Wyllie, P. J. (1977). Crustal anatexis: An experimental review. *Tectonophysics* **43**, 41–71.

Yardley, B. W. D. (1986). Is there water in the deep continental crust? *Nature* **323**, 111.

Yoder, H. S. (1990). Heat transfer during partial melting: An experimental study of a simple binary silicate system. *J. Volc. Geotherm. Res.* **43**, 1–36.

Younker, L. W., and Vogel, T. A. (1976). Plutonism and plate tectonics: The origin of circum-Pacific batholiths. *Can. Mineral.* **14**, 238–244.

Zingg, A. (1990). The Ivrea crustal cross-section (northern Italy and southern Switzerland), *in* "Exposed Cross Sections of the Continental Crust" (M. H. Salisbury and D. M. Fountain, eds.), pp. 1–20. Kluwer, Dordrecht.