

Dynamics of differentiation in magma reservoirs

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Abstract. In large magma chambers, gradients of temperature and composition develop due to cooling and to fractional crystallization. Unstable density differences lead to differential motions between melt and crystals, and a major goal is to explain how this might result in chemical differentiation of magma. Arriving at a full description of the physics of crystallizing magma chambers is a challenge because of the large number of processes potentially involved, the many coupled variables, and the different geometrical shapes. Furthermore, perturbations are caused by the reinjection of melt from a deep source, eruption to the Earth's surface, and the assimilation of country rock. Physical models of increasing complexity have been developed with emphasis on three fundamental approaches. One is, given that large gradients in temperature and composition may occur, to specify how to apply thermodynamic constraints so that coexisting liquid and solid compositions may be calculated. The second is to leave the differentiation trend as the solution to be found, i.e., to specify how cooling occurs and to predict the evolution of the composition of the residual liquid and of the solid forming. The third is to simplify the physics so that the effects of coupled heat and mass transfer may be studied with a reduced set of variables. The complex shapes of magma chambers imply that boundary layers develop with density gradients at various angles to gravity, leading to various convective flows and profiles of liquid stratification. Early studies were mainly concerned with describing fluid flow in the liquid interior of large reservoirs, due to gradients developed at the margins. More recent work has focused on the internal structure and flow field of boundary layers and in particular on the gradients of solid fraction and interstitial melt composition which develop within them. Crystal settling may occur in a surprisingly diverse range of regimes and may lead to intermittent deposition events even with small crystal concentrations. Incorporating thermodynamic constraints in the study of the dynamics of settling has only just begun. Many dynamical phenomena have been found using theoretical arguments, laboratory experiments on analog systems, and numerical calculations on simplified chemical systems. However, they have seldom been applied to natural silicate melts whose phase diagrams and important physical properties such as thermal conductivity and chemical diffusion coefficients remain poorly known. There is a gap between model predictions and observations, as many models are designed to explain large-scale features and many observations deal with the local texture and mineral assemblages of the rocks. This review stresses the relevance to the geological problem of the work carried out in parallel in other disciplines, such as physics, fluid dynamics, and metallurgy.

Introduction

The study of the physics of the processes that may take place in magma chambers has been a strikingly active research field during the last 20 years or so. The main purpose of this review is to put the contribution of this work into some historical context. The issues of how heat transfer and crystallization take place have been discussed qualitatively, since the early days of research on igneous differentiation [Bowen, 1921, 1928; Hess, 1960; Jackson, 1961; Wager, 1963; Wager and Brown, 1968]. The macroscopic physical framework necessary to treat such difficult physical problems was not available then. In a sense, many of the recent physical studies have tackled quantitatively problems that had been outlined qualitatively many years ago. However, we shall also see that the focus has shifted away from those early questions, as it has become clear that partially crystallized systems are exceptionally rich in new dynamic phenomena.

The geological observations require the physical processes of cooling and of separation of crystals and residual liquids. The tendency has been, however, to use the rocks to deduce how a reservoir operates in a physical sense. This approach avoids the physical problem which has its own logic. This review concerns what causes the system to evolve in such a way as to produce the observed differentiation trend, i.e., how cooling and crystallization are likely to develop from given starting conditions. We will not attempt an exhaustive summary of all recent work on magma dynamics because it covers a very wide range of phenomena but will concentrate on two central questions. First, how do crystallization and chemical evolution of magmas take place, and second, what determines the rate of cooling of a magma body? These questions are strongly related because, on the one hand, chemical evolution of the magma depends on where crystallisation occurs and how separation of crystals and residual melt takes place, while on the other hand, the way in which heat loss takes place determines how crystallization proceeds. This coupling of cooling and chemical evolution of the liquid, the phase change, and the macroscopic transport phenomena makes the physical problem a difficult one to treat from first principles.

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The difficulty involved may be illustrated by the following example. One early model for igneous layering was double-diffusive convective phenomena [McBirney and Noyes, 1979; Turner, 1980]. In fluids in which two or more components contribute to the density and diffuse at different rates, a kind of "staircase" stratification can develop, in which layers convect on vertical scales that are much less than the overall thickness of the liquid body and are separated by thin diffusive interfaces. Various suggestions have been made for how such layering in the liquid state might be translated in the final rock [e.g., Kerr and Turner, 1982; Irvine, 1980]. This idea has not proved as fruitful as might originally have been thought. The physical reason, first pointed out by McBirney [1985] and Morse [1986], is that in a crystallizing system temperature and composition cannot be varied independently because one is constrained by the liquidus relationship. In all the experiments, the liquids were superheated, and layering occurs because a significant amount of thermal buoyancy is available [e.g., Chen and Turner, 1981]. This is unlikely in a system constrained to lie close to its liquidus.

Research on the dynamics of crystallization has been very active because it has applications in many different fields including metallurgy and the crystal growth industry. Thus several collective books and reviews have been published in the recent past [Loper, 1987; Huppert, 1990; Davis et al., 1992]. To set the stage for the present discussion, we first summarize the basic framework adopted by petrological studies. This framework has been gradually adapted to accommodate new facts, but its basic tenets have been kept. We also include a discussion of the respective roles of physical, geochemical and petrological models. This is important because the three approaches do not deal with the same pieces of evidence and differ in their predictive abilities. For example, there are few physical models of igneous textures, and yet these constitute one of the key observations used by petrologists. Physical models are only able to address gross features of igneous intrusions, and hence we shall include a description of these. For each mechanism, we explain its implications for petrological analysis and discuss the evidence for and against it in the geological record.

Petrogenetic and Geochemical Models

In this section, we review a series of models for magmatic differentiation which do not deal explicitly with the physics of the differential motions between magma and crystals. Most of these models have been developed within a chemical and thermodynamic framework and have not addressed directly the relationship between the thermal regime of magma chambers and their differentiation behavior.

Cumulus Theory

For most of this century, principally since the influential work of Bowen [1928], the main emphasis has been on the investigation of the thermodynamics of the chemical equilibria between crystals and liquids in various igneous systems. In the small capsules in which this type of experiment is carried out, large thermal gradients can generally be avoided, and the charge is essentially homogeneous. This work has focused on the liquid compositions and mineral assemblages generated as a function of temperature for a given starting composition. Perhaps as a result, magma chambers became viewed in the same light, that is, as a body of liquid with essentially one composition at any given

time, in equilibrium with a precise mineral assemblage. The main physical idea invoked to explain magmatic differentiation was the gravitational settling of crystals from less dense liquid.

The qualitative physical picture of a magma chamber was founded on two principle pieces of evidence: that of the overall stratigraphic relations of mineralogical and chemical variations and that of the cumulate textures of the rocks. The textures were interpreted as indicating that the crystals were deposited as a sediment. The physical history of the formation of a rock was subdivided into a phase of deposition, one of enlargement of the deposited crystals when close to or at the top of the cumulus pile (so-called adcumulus growth) and one of crystallization of the interstitial trapped liquid [Wager et al., 1960]. The main physical agent of mass transport during adcumulus growth was thought to be chemical diffusion. The fact that the cumulate rocks became progressively more evolved as a function of stratigraphic height was taken to show that the whole volume of liquid in the chamber was evolving chemically because of crystallization [Morse, 1979a,b]. It was thus supposed that crystal settling was efficient in the sense that it occurred at a much faster rate than crystallization of the body of liquid.

One additional consequence of these ideas is the requirement that the partially crystallized layer at the floor of a reservoir must be thin, because the diffusive chemical flux is not large and cannot supply high temperature chemical components for adcumulus growth over large volumes [Wager, 1963]. This led to important conclusions regarding the heat budget of crystallization at the floor. Taking into account the fact that the bottom contact is initially cold, as shown by the common occurrence of basal chill zones, the temperature gradient through a thick layer of adcumulate rocks must be directed downward and small. Thus, if heat is removed by conduction through the bottom contact, the melt region whose temperature is close to the liquidus, where adcumulus growth occurs, is thick. This is inconsistent with the chemical budget considerations outlined above. Thus Wager [1963] proposed that heat must be transported into the overlying melt layer, which implies that the interior of the reservoir is undercooled. Morse [1986] has recently elaborated on this idea.

Bottom Crystallization With Chemical Diffusion

Cawthorn and McCarthy [1980, 1985] have argued that at the floor of an intrusion, crystallization occurs in situ, and have stressed the continuous variations of trace elements over some zones of the Bushveld complex. For example, chromium concentration varies in a monotonic fashion over a height of 8 m in magnetite layers of the Upper Zone of this complex. Cawthorn and McCarthy [1980, 1985] considered an essentially flat solidification front moving at a constant velocity with a diffusion boundary layer developing in the melt ahead of it. In the case of a compatible trace element, the initial solid takes a large amount of this element, as dictated by the partition coefficient, and a diffusive boundary layer develops which is depleted in this element. This leads to a decrease of the concentration in the solid rock until a steady state is reached such that the solid has the same trace element concentration as the melt. The observations are inconsistent with known values for the diffusion coefficient of the species studied, and hence some form of convection is called for [McCarthy et al., 1985]. This model relies on the assumption that the interface between solid and melt can be treated as a flat boundary, such that there is little gradient of interstitial melt fraction below it.

Reinjection

One striking feature of large igneous intrusions is the repetition of sequences of layers, in which the proportions of constituent minerals vary, on a wide range of length scales [Parsons, 1987]. Large-scale cyclic layering occurs on length scale of 50 m. in ultramafic zones at the base of intrusions. For this type of layering, Jackson [1961] developed a sophisticated conceptual model involving an interaction in a closed system between convection and crystallization. He proposed that the build up of crystals in a sublayer in the lower part of the magma chamber might isolate it from the main body of liquid. This sublayer then crystallized separately, until eventually the concentration of crystals became sufficient to suppress convection and an abrupt sedimentation event occurred, depositing one cyclic layer. The explanation that has been much more favored in the recent literature is that of reinjection of the chamber with more primitive melt that interrupts the crystallization sequence. Original geological examples were the Rhum Intrusion [Brown, 1956] and the Muskox Intrusion [Irvine and Smith, 1967], but this interpretation has now been extended to others.

Assimilation

With the advent of trace element and isotope geochemistry, it has become clear that wall rock assimilation may play an important role [Allègre and Minster, 1978; DePaolo, 1981]. Changing Sr and Nd isotopic ratios in cumulate rocks provide unambiguous evidence for the mixing of melts coming from sources with different compositions and ages. Present interpretations have fractional crystallization and assimilation occurring simultaneously, with the magma chamber melting its own host rocks [DePaolo, 1985].

Geochemical Models of Igneous Intrusions

The present conceptual framework for the chemical evolution of magma reservoirs includes fractional crystallization, assimilation, reinjection, and venting. Detailed modeling of the paths in composition space that are followed by liquids has been carried out, generally concentrating on trace elements and isotopes [O'Hara, 1977; Palacz, 1984; Stewart and DePaolo, 1990]. Major element trends can be deduced from phase diagrams, when they are available [Ghiorso, 1985; Nielsen, 1988, 1989; Ghiorso and Sack, 1994]. In these models, parameters include the amounts of fractionation and assimilation, which may be allowed to vary as a function of time (and hence stratigraphic height in the cumulate layer). In the following, for the sake of simplicity, we shall refer to such models as "parameterized."

Discussion: Differentiation in Large Magma Chambers

Differentiation of a magma chamber involves large-scale relative motions of solid and liquid, which must be studied with specific methods. Experiments carried out with natural magmas in petrological capsules cannot be used to study the importance of such processes in a physically realistic manner. The characteristic length scales inherent in such phenomena determine, for example, the magnitudes of the buoyancy and viscous forces acting in a convective flow and hence the dynamic regime. Whereas laboratory experiments with natural magmatic liquids establish equilibrium thermodynamic constraints where length scales do not have to be specified, they give no information on dynamic phenomena.

A large number of studies have dealt with the diffusive transport of a given chemical species due to gradients of other intensive thermodynamic variables. One such variable is temperature, which gives rise to the Soret effect [Hildreth, 1981; Walker et al., 1981; Lesher, 1986]. Other variables are the concentrations of the other chemical species involved [Baker, 1990, 1992; Kress and Ghiorso, 1993; Trial and Spera, 1994; Liang et al., 1994; Kress and Ghiorso, 1995]. The effects depend on the magnitude of gradients which are present, which cannot be specified a priori. In realistic conditions with thermal boundary layers evolving due to heat conduction or convective breakdown, the mass fluxes due to Soret diffusion and cross diffusion are small [Carrigan and Cygan, 1986; Lesher and Walker, 1988]. Such effects, however, are important on a small scale and may be responsible for some of the features of igneous textures.

In many "parameterized" differentiation models, a key assumption is that of a homogeneous system, in which each part of the system simultaneously undergoes the same evolution. This is the major premise that more recent work on convection and the dynamics of boundary layers has called into question. The emphasis has now shifted to studies of the physics of partially crystallized layers, as anticipated by Hess [1972].

Drilling through the partially solidified crust of basaltic lava lakes at Kilauea volcano, Hawaii, has provided us with an opportunity to study directly how a large body of melt differentiates. A series of studies have given us access to profiles of solid fraction, interstitial melt composition and temperature as a function of time and have indicated that several processes are simultaneously active [Richter and Moore, 1966; Wright et al., 1976; Helz, 1980; Helz and Thornber, 1987; Helz, 1987; Helz et al., 1989]. For our present purposes, it is only useful to mention three. One is that crystal settling is essentially limited to the coarse phenocrysts which were present in the lava when it was erupted. The major crystal phases which formed when the lake cooled, olivine, augite and plagioclase, did not settle to any appreciable extent. A second result is that low-density melt from the floor cumulate pile rises through the lake interior and mixes with the melt at the base of the growing upper crust. A third result is that the boundary layers contain mineral assemblages and coexisting interstitial liquids spanning the whole crystallization interval between the solidus and the liquidus.

Importance of Physical Models

Purposes of Physical Models for Igneous Differentiation

Chemical evidence on igneous intrusions is rapidly accumulating, which causes a dilemma. On the one hand, building a physical model from first principles in order to reproduce such vast amounts of data is at present impossible. On the other hand, a complex sequence of events for any given magma chamber implies a large number of adjustable parameters, which virtually ensures a satisfactory fit to the data. Thus it has been argued that it is pointless to try and understand the physics of each process individually. Setting aside the obvious points that the advance of scientific knowledge requires such fundamental studies and that extrapolating knowledge from known intrusions to other situations requires an understanding of the physics involved, there are two main reasons why physical models are needed.

Parameterized chemical models may not be unique, and some aspects of them may be criticized on physical grounds. In other words, such models guarantee consistency with the data, as they are basically set up as inverse problems, but there is no guarantee

of compatibility with physical laws. For example, consider a case of adcumulus growth, such that interstitial melt between touching crystals evolves chemically to produce a totally crystallized rock essentially made of a single mineral phase. This requires efficient chemical exchange between the cumulate pile and the overlying reservoir of liquid. The point that chemical diffusion would be ineffective over a large thickness of partially solidified material was appreciated by the proponents of the cumulus theory, who concluded that adcumulates were evidence of thin partially-solidified layers. A physical model is required to demonstrate that this is a correct inference. Other key elements of petrogenetic models have not been studied from a physical viewpoint. For example, despite the enormous weight attached to the idea of crystal settling, not much consideration was given to how this might actually take place in a convecting magma. One thing made clear by the physical studies is that in a reservoir of large dimensions, many processes which might seem plausible at first sight are found to be unstable.

Another reason to develop physical models is because parameterized models are seldom able to specify the actual mechanisms at work. For example, the amount of assimilation is often found to vary as a function of time in an evolving chamber [DePaolo, 1985; Stewart and DePaolo, 1990]. The melting of country rock may occur during or soon after emplacement, with a large volume of magma slowly getting incorporated in the primary liquid. Alternatively, it may occur continuously whilst the reservoir is differentiating. It is the efficiency of mixing which varies in the first case, while it is the efficiency of melting which varies in the second one.

Inconsistencies With the Cumulus Theory

Some features of layered igneous rocks appear inconsistent with the cumulus theory. In the Skaergaard, there is a sharp discontinuity between the rocks that are supposed to have formed by sedimentation onto the floor of the intrusion and the Upper Border Group and Marginal Border Group that are thought to have formed by growth of crystals sticking onto the walls. Olivine crystals may be found in the latter even though they are much denser than the melt from which they grew. A major cumulus phase of mafic rocks is plagioclase which seems likely to be less dense than many mafic magmas [McBirney and

Noyes, 1979; Campbell *et al.*, 1978]. This is clearly inconsistent with crystal settling.

Evidence From Lava Series

Another problem has been the comparison between intrusive and extrusive rocks that gives direct insights into the composition of melt in the reservoir both as a function of depth and as a function of time. Many volcanic deposits give evidence that magma chambers are chemically stratified [e.g., Larsen and Thorarinsson, 1977; Hildreth, 1981; Bogaard and Schmincke, 1985; Huijsmans and Barton, 1989; Druitt *et al.*, 1989]. This violates the assumptions behind most geochemical models.

General Features of Igneous Differentiation

We show here that the characteristics of igneous intrusions vary systematically as a function of a magma chamber thickness. Also, the same internal features are repeated in different intrusions of similar overall bulk composition and size, indicating that the physical processes of differentiation are reproducible. Such regularities in the rock record are best assessed from a physical point of view.

The degree of internal differentiation of igneous intrusions is largest in basaltic systems, and is related to the size of the reservoir. Figure 1 shows the compositions of plagioclase feldspar, a ubiquitous phase, in several mafic intrusions of similar bulk compositions which span almost 3 orders of magnitude in vertical thickness. A general effect of differentiation is that the anorthite content of plagioclase decreases. Such a decrease is very pronounced as one goes up the stratigraphic sequence in the Bushveld and Skaergaard intrusions, which are approximately 9 km and 3 km thick, respectively. In the 300-m-thick Palisades sill, differentiation of the melt was much weaker. Finally, in the thin (40 m) Dippin sill, there is no significant variation of plagioclase composition. These features are apparent on a large scale but are made more complex on a smaller scale. Another important point shown by Figure 1 is that in cases in which both floor and roof sequences are preserved, both show the same trend of differentiation; however, the floor sequence is typically 6-7 times thicker than the roof sequence.

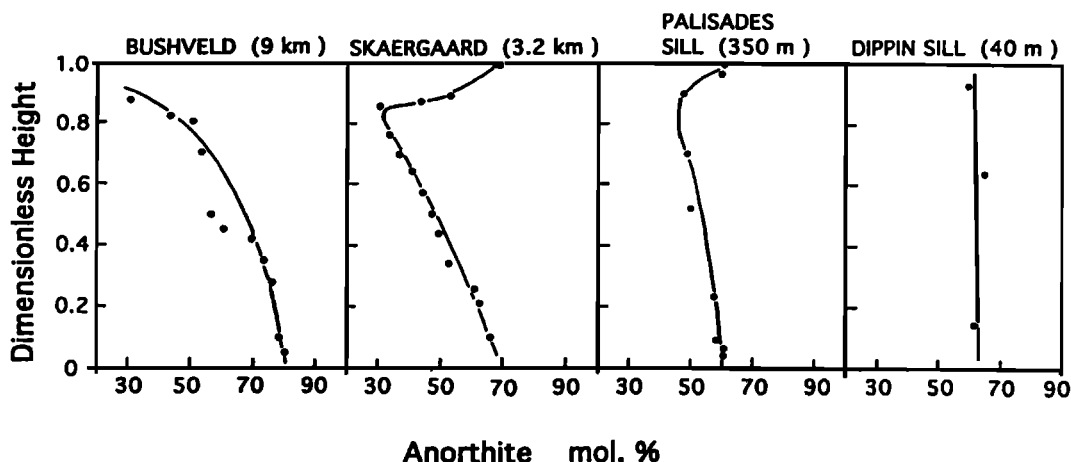


Figure 1. The stratigraphic variations of plagioclase composition for intrusions spanning a large size range. Differentiation becomes more pronounced the larger the intrusion. Both roof and floor sequences of the Palisades Sill and the Skaergaard Intrusion are preserved and show that the floor sequence is thicker than the roof sequence, typically by a factor of 6. References are Wager and Brown [1968], McBirney [1989], Shirley [1987], and Gibb and Henderson [1978].

In Figure 2, we compare the vertical distribution of incompatible trace elements in three of the same intrusions. Starting from the floor, concentrations remain essentially constant over a large vertical extent and increase markedly toward the top. This indicates that as crystallization and differentiation proceeded, liquid/crystal separation was very efficient and residual liquid became progressively enriched in incompatible trace elements. The final concentration buildup occurs when the melt was sufficiently enriched to precipitate specific host minerals.

Mafic intrusions exhibit similar mineral assemblages and overall stratigraphic relations. This point was emphasized by *Wager and Brown* [1968] and *Jackson* [1970] and implies that as far as the main differentiation trend is concerned, essentially the same sequence of processes was repeated. For example, Figure 3 shows the stratigraphy of the Bushveld and Stillwater intrusions and demonstrates that major changes of mineral compositions occur at similar heights in the sequence. Furthermore, major precious metal-bearing sulphide horizons are found at almost the same positions. A very similar sequence seems to exist in the much younger Sept-Iles mafic complex, Quebec, which has similar dimensions [*Loncarevic et al.*, 1990]. In comparison, the much thinner Palisades sill was never able to produce dunites and extreme concentrations of precious metals. Furthermore, the onset of mineralogical layering occurs at similar distances from the lower contact in intrusions of different thicknesses. For example, a layer of increased olivine content is observed at 11 m from the bottom in the 600-m-thick Lambertville sill [*Jacobein*, 1949], at 12 m in the 350-m Palisades sill [*Walker*, 1940] and at 9 m in the 43-m Dippin sill [*Gibb and Henderson*, 1978]. Another example is that in large intrusions, rocks highly enriched in olivine are first found at distances between 100 and 300 m from the floor contact in several different complexes: Stillwater [*Jackson*, 1961], Bushveld [*Vermaak*, 1976], and Muskox [*Irvine*, 1970]. These observations indicate that differentiation starts at some specific distance from the contact, independently of the size of the intrusions.

Structure of Boundary Layers

Model Thermodynamic Systems

Physical models cannot realistically hope to incorporate the same level of chemical and thermodynamic detail that exists in natural silicate systems, and it is legitimate to ask what is the

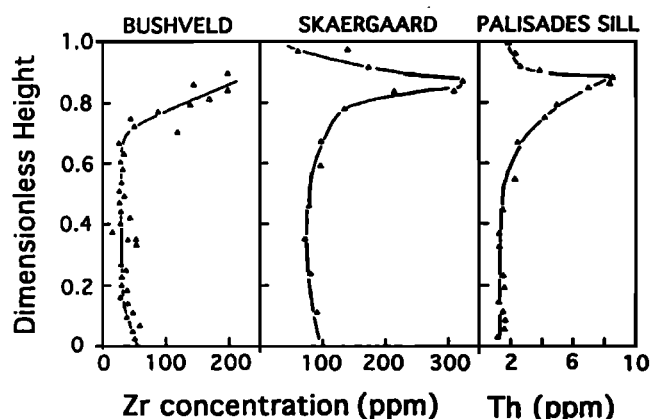


Figure 2. Stratigraphic variations of incompatible trace elements concentrations for three of the same intrusions as in Figure 1. References are *Cawthorn* [1983], *McBirney* [1989] and *Shirley* [1987].

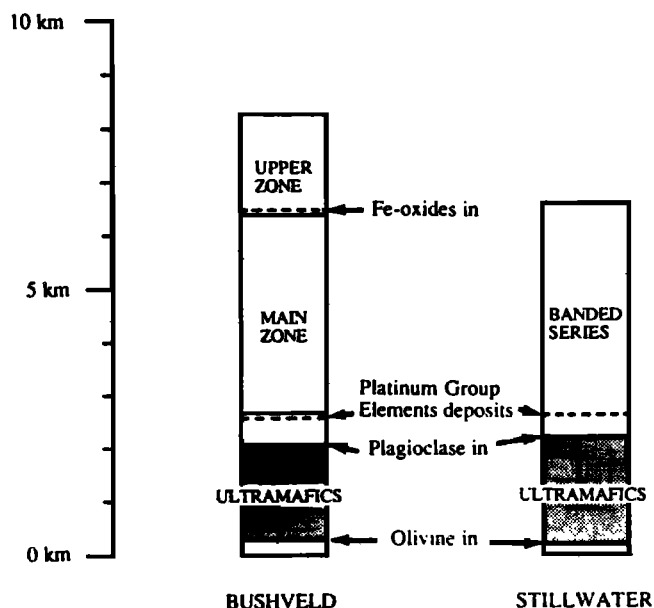


Figure 3. Comparative stratigraphy of the Bushveld and Stillwater intrusions emphasizing some large-scale features of the mineralogic sequences from [*Naldrett et al.*, 1987].

best kind of simplified analogue. This depends on the problem which is being posed.

Systems involving both binary eutectics without solid solution and a single solid solution loop have been used. Natural magmas exhibit a mixture of these features in the sense that several different solid phases form, while each phase exhibits some extent of solid solution. Magmatic differentiation trends are recorded both by the varying amounts of minerals phases present in the rocks and by the so-called "cryptic" variation of the compositions of the various solid solutions. In binary systems with just a solid solution loop, all nucleation takes place at the initial temperature and further crystallization is achieved by growth at different compositions. The solid composition may adjust by reequilibration with liquids if solidification conditions permit. In binary eutectic systems without solid solution, each one of the two solid phases has a fixed composition and the chemical evolution of the liquid can be directly predicted as a function of fraction crystallized. Enrichment or depletion of the final solid with respect to the initial liquid can be measured for either component of the binary.

The origin of modal layering, that is, the formation of solids with variable proportions of different mineral phases, and the conditions for adcumulus growth can be studied with a simple eutectic system. It will certainly be desirable in the future to have some solid solution in order to see how this other degree of freedom affects crystallization conditions. One may reasonably expect that as far as the fluid dynamics of crystallization are concerned, this will not change the basic principles at work and will only affect the specific details of the chemical evolution. We discuss below observations in the boundary layers of lava lakes which shed some light on this issue.

Quiescent Case

Before evaluating the effects of differential motion between magma and crystals, which entails both heat and mass transfer, it is useful to consider static conditions. This illustrates the initial conditions when motions start and allows a simple discussion of several features of cooling and crystallization in a large chamber.

After its emplacement, magma loses heat to the surrounding country rock which is initially effected through the growth of cold thermal boundary layers at the margins of the body (Figure 4).

In most circumstances, i.e., for normal "cold" country rock, the country rock/magma interface is below the solidus temperature: thus the boundary layer spans the whole crystallization interval. This is probably the most important physical fact. Thermodynamic constraints dictate that mineral assemblages differ in bulk composition with the melt and hence that gradients of melt composition are set up when crystallization occurs. In principle, one may envisage the growth of a totally solid layer, with chemical diffusion in the melt supplying the required elements to the solid. However, this seldom occurs in practice and a two-phase system develops with the solid phase as crystals disseminated through the undercooled layer. For binary systems with and without solid solution, this has been demonstrated theoretically and experimentally when the cooling rate is above a certain critical value [Mullins and Sekerka, 1964; Loper and Roberts, 1978, 1981; Hills *et al.*, 1983; Caroli *et al.*, 1985; Worster, 1986]. For silicate melts, this is shown by direct observation [Helz, 1987]. Models for such a mixed system have been derived using various approximations.

By definition, there is a large gradient of temperature across a thermal boundary layer and, to a lesser extent, a gradient of pressure. If the boundary layer is much thicker than the typical

crystal size, as is usually the case, thermodynamic equilibrium may be achieved by diffusion between a crystal and the melt surrounding it but cannot be achieved over the whole boundary layer. The simplest approximation is that at every position, melt and crystals are in thermodynamic equilibrium at the local conditions. Thus the temperature gradient implies one of chemical composition. For a given bulk starting composition, the local value of liquid composition lies on the liquidus and is determined by the local temperature. The crystal content and solid composition are also known using the phase diagram. Hills *et al.* [1983], Roberts and Loper [1983], and Bennon and Incropera [1987] have developed a so-called "mixture" model along these lines, where the system of melt plus crystals can be studied as a single material with effective properties depending on temperature and crystal content. Experiments on the conductive cooling of simple binary systems, such as aqueous solutions of simple salts, have shown the validity of the theory and of the assumption of local thermodynamic equilibrium [Huppert and Worster, 1985; Tait and Jaupart, 1992a]. In these calculations and experiments, it may be seen that the local solid fraction increases continuously from zero at the liquidus to one at the eutectic temperature.

In magma chambers, therefore, a thermal boundary layer should contain the whole liquid line of descent of the initial magma. Such characteristics have been observed in basaltic lava lakes. The summary of Helz [1987] states that the order of appearance of the phases is simply related to the temperature at the local position in the boundary layer. The extent to which the phase compositions reflect local equilibrium varies according to the phase. Moving from the hottest to coldest parts of the boundary layer, as the interstitial liquid becomes more evolved, the olivine grains reequilibrate and have progressively more iron-rich compositions. Diffusion is less efficient in plagioclase grains which become zoned in consequence, equilibrium being maintained only between the surfaces of the grains and the interstitial liquid. Pyroxenes show intermediate behavior and reequilibrate partially. Thus, in this natural boundary layer, local equilibrium for the phases, the mineral compositions, and the interstitial liquid compositions is nearly attained. Some of the interstitial liquids are rhyolitic in composition, even at an early stage of evolution.

Departures from local thermodynamic equilibrium may be expected but are very difficult to treat theoretically. Attempts have been limited to the definition of the smallest number of variables which allow a tractable set of equations [Loper, 1987; Hills and Roberts, 1993] and to simple parameterizations of local diffusion between crystals and surrounding melt [Toramaru, 1991]. Using expressions for the kinetic rates of nucleation and crystal growth, Brandeis and Jaupart [1987a,b] and Spohn *et al.* [1988] have calculated the distribution of crystal sizes. This is useful because crystal size not only dictates the efficiency of crystal settling but is also an independent variable which may be measured. At any given time in the partially crystallized boundary layer, crystal size increases toward the contact with country rock, while crystal content simultaneously increases. As cooling proceeds, the mean crystal size increases and the final result in fully solidified rock will be an increase of crystal size with increasing distance from the margin. Model predictions have been compared successfully with grain size data in several sills and dikes [Brandeis and Jaupart, 1987a,b; Spohn *et al.*, 1988; Hort and Spohn, 1991]. During solidification at the floor of a magma reservoir, for example, the largest crystals are lying in a high crystallinity region close to the margin: those crystals

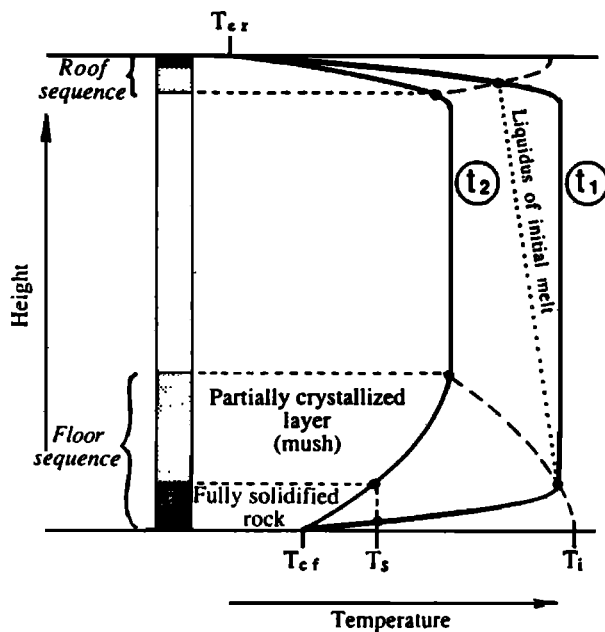


Figure 4. Shows qualitatively the thermal structure of boundary layers that are formed at the contacts with the country rock colder than the solidus temperature. The coldest part of the boundary layers contain solid rock, while the remainder spans the temperature interval between solidus and liquidus. Thermal convection leads to a higher heat flow through the roof than through the floor and hence to an asymmetry in the thicknesses of the two thermal boundary layers. This can lead to a thicker rock sequence at the base than at the roof. Two complicating factors are illustrated. One is the variation of liquidus temperature with pressure, and hence height in the chamber, and the other is the possibility that the contact temperatures at the floor and roof are different.

with the largest settling velocities are already at the floor and have nowhere to settle to.

"Slurry" and "Mush" Models

When relative motions between melt and crystals are allowed, there are two main types of theoretical models. Both are identical in the motionless state and only differ in dynamic situations. In "mush" models [Roberts and Loper, 1983; Hills *et al.*, 1983; Drew, 1983; Worster, 1991, 1992], the crystals stay fixed with respect to each other, and melt is allowed to move with respect to them in response to buoyancy forces. The variables for such models are solid (or liquid) fraction, temperature, and melt composition. This type of model applies, for example, when crystallization takes the form of dendrites forming an interconnected network of solids attached to the cooling boundary. This is appropriate to many situations in metallurgy and to many aqueous solutions of simple salts which have been used in laboratory experiments. In the geological case, it appears valid for the cooling of spinifex-textured komatiite flows [Turner *et al.*, 1986]. It is also valid for the coldest part of the boundary layer where crystals start forming an interconnected network. This certainly occurs at solid fractions higher than 60%, which is the packing limit for a random spatial distribution of equal spheres but may, in fact, occur at lower solid fractions as crystals may be arranged in chain structures. In "slurry" models [Loper and Roberts, 1987], crystals and liquid may move, as a whole or independently of each other. These models require the additional specification of crystal sizes. A slurry may be involved in bulk convective motions involving both crystals and melt. Interestingly, one of the earliest slurry models is that of Malkus [1973] who applied it to the cooling of Earth's core and included crystal settling.

Liquid Density

Density varies as a function of both temperature and composition, which may lead to two different kinds of convective motions, driven by unstable thermal gradients and unstable compositional gradients. Thus the crucial issue is how the densities of naturally occurring melts tend to evolve as they cool and differentiate. A concept that has proved useful is that of fractionation density [Sparks and Huppert, 1984]. This is the density of the chemical components that are incorporated into a mineral phase but considered as a melt. This is an elegant way of avoiding the problem of the jump in density that occurs because of the change of phase. One can see directly that if the fractionation density of a mineral is superior to the density of the melt from which it is crystallizing, the residual melt must be less dense, and vice-versa. An example in which residual melt is less dense is given in Figure 5.

In natural magmas, the predominant effect of fractional crystallization is to increase in the SiO_2 content of the melt, which tends to reduce density. In basaltic melts, density can initially increase with differentiation because of an early trend of iron enrichment from primitive to ferrobasalts. The key issue concerns the existence and position of the point at which Fe-Ti oxides start to precipitate, which is controlled by the oxygen fugacity of the melt as well as its bulk composition [Juster *et al.*, 1989; Snyder *et al.*, 1993]. Most basic and ultrabasic intrusions have Fe-Ti oxide minerals. After Fe-Ti oxides saturate in such tholeiitic systems, the Fe/Mg ratio of the melt typically continues to increase whereas the absolute Fe content of the melt and its density decrease. The initial volatile content of the melt is also

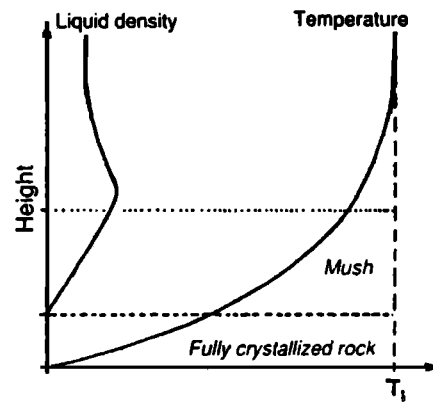


Figure 5. Schematic section through a boundary layer formed when the lower contact is below the solidus temperature. A steep temperature gradient is created by thermal diffusion. The partially solidified region between solidus and liquidus contains the whole liquid line of descent. In this case the stabilizing thermal effect on density is overwhelmed by the destabilizing effect of composition, and the bulk profile of density in the interstitial liquid is unstable. There is also a thin, unstable boundary layer just above the mush created by chemical diffusion.

very important. Figure 6 from [Snyder *et al.*, 1993] shows that for a particular basaltic liquid, the density maximum is suppressed for an initial water content larger than approximately 0.5 wt %. Basalts in subduction-related settings seem generally to contain more than this amount [Sisson and Layne, 1993]. Even mid-ocean ridge basalts which are among the driest mafic magmas quite commonly have close to this amount of water [Byers *et al.*, 1984; Dixon *et al.*, 1988] as well as other volatiles such as CO_2 and halogens.

We conclude that crystallization the boundary layer of magma chambers will in most circumstances release light residual melt, because the compositional effect that tends to reduce the density dominates the opposing thermal effect. Hence the interstitial liquid will be convectively unstable in boundary layers at the floor and sidewalls.

Boundary Layer Instabilities

Calculations show that unstable density gradients are generated both within boundary layers and at their edges due to

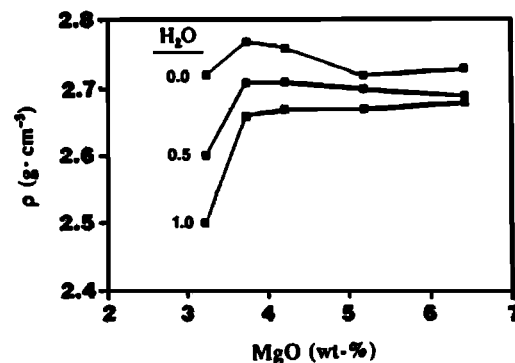


Figure 6. The density of residual melt as a function of chemical differentiation for a tholeiitic mafic liquid after Snyder *et al.* [1993]. The density maximum that is present under dry conditions is eliminated if the initial water content is greater than approximately 0.5 wt %.

diffusion. Convection takes different forms depending on the location of the boundary layer. Along horizontal margins, at the roof and floor, say, convection will occur as a set of plumes of small dimensions originating from the boundary layer. Along a vertical or sloping margin, convection will take the form of a sidewall current which may be either turbulent or laminar depending on the dimensions and viscosities involved. Research has looked into these phenomena independently, whereas all must be expected to occur in a real magma chamber. However, fortunately, these boundary layers contribute in different ways to the thermal and compositional evolution of the reservoir. In terms of heat removal, the most efficient interface is that which lies at the roof. Thus, for the purposes of determining the cooling rate of a magma reservoir, accurate understanding of this boundary layer must be obtained. On the other hand, for the purposes of looking at the cumulate sequence which grows at the floor of a sheetlike intrusion, attention must be paid mostly to in situ crystallization in the boundary layer [Jaupart and Brandeis, 1986], with additional effects of crystal settling from the reservoir interior. To understand how the chamber liquid interior may become chemically stratified, one must focus attention on the sidewalls, although boundary layer instabilities at the roof and floor may contribute somewhat to the process.

Thermal Convection

Compositional convection weakly affects the thermal regime of crystallization [Tait and Jaupart, 1989, 1992a]. Thus how the interior of a magma chamber cools and crystallizes depends critically on the existence of thermal convection. The overall cooling time of a reservoir will also depend on heat transport in the roof rocks, whether or not they melt, whether or not hydrothermal convection occurs, their thermal conductivity, etc. The thermal regime of the magma also bears on the importance of crystal settling. In the absence of thermal convection, crystals nucleate and grow only in boundary layers, and the interior of the reservoir away from the boundary layers remains at its initial temperature. With thermal convection, on the contrary, the whole magma reservoir is being cooled by the motion of cold plumes, and crystals may nucleate and grow over the whole magma volume (Figures 7 and 8). Thus, in fact, although this

intimate relation is not always appreciated, the "cumulus" theory rests on the existence of thermal convection! It may therefore not come as a surprise that the existence or absence of thermal convection has been hotly debated [Carrigan, 1987; Marsh, 1989; Huppert and Turner, 1991; Marsh, 1991; Davaille and Jaupart, 1993b].

Temperature Difference Driving Convection

The Rayleigh number defines the potential for thermal convection in a layer of liquid subjected to a temperature difference ΔT :

$$Ra = \frac{g \alpha \Delta T d^3}{\kappa \nu} \quad (1)$$

where α is the coefficient of thermal expansion, κ is thermal diffusivity, d is the thickness of the liquid layer, and ν is kinematic viscosity. The Rayleigh number may be calculated once appropriate scales have been specified for the variables. In magma chambers, temperature differences imply changes of melt viscosity and rheological changes due to the presence of crystals. Thus there are large variations of physical properties, which makes it difficult to select the proper scales. For example, it is clear that the overall temperature difference between the magma and its country rock, which may take values larger than 500°C, is not relevant as magma is chilled against the colder country rock. In the definition of the Rayleigh number, the thickness of the liquid layer is raised to the power three, and hence the size of the system is the most important parameter concerning the existence or absence of convection. In a large reservoir, therefore, some form of natural convection seems inevitable. At high values of Ra , when convection is active, the important length scale is the thickness of the unstable boundary layer, and it is sufficient for many purposes to study the local dynamics of boundary layers.

Most knowledge of thermal convection pertains to constant viscosity systems and can only be extrapolated to the magmatic case with caution. In general terms, one expects the coldest part of the boundary layer, which is very viscous, to remain stable, and the key question is to define which part of the overall temperature difference is available to drive convection. On the basis of this general principle, Turner *et al.* [1986], Brandeis and



Figure 7a. A set of thermal plumes generated by sudden cooling at the upper boundary of a layer of silicone oil. Note the small dimensions of the plumes in comparison with the overall thickness of the layer of liquid.

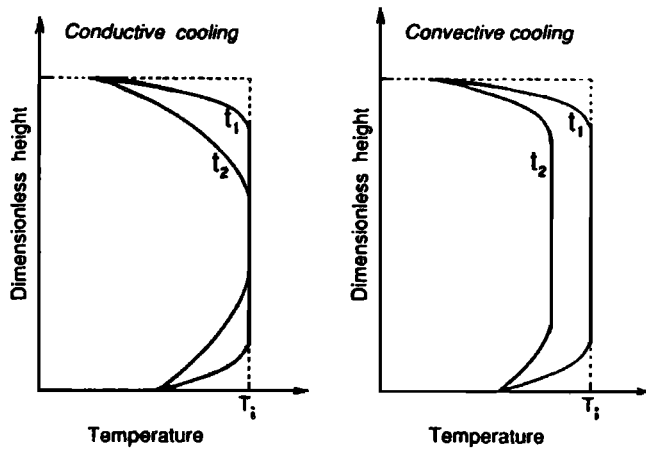


Figure 7b. This contrasts qualitatively some aspects of the thermal histories of bodies of liquid that are cooled by thermal conduction and convection. The left-hand panel shows conduction in which the boundary layers slowly thicken with time and the center of the liquid layer stays at the initial temperature for a significant portion of the cooling history. The right-hand panel shows convection, in which the stirring effect of convection keeps the boundary layers thin and rapidly transmits cooling to the center of the liquid.

Marsh [1989], Worster *et al.* [1990], and Kerr *et al.* [1990] have relied on the assumption that convection is driven by the amount of kinetic undercooling, i.e., the temperature difference ahead of the partially crystallized layer, which takes a typical value of a few degrees. This assumption was supported by laboratory experiments on various systems with dendritic mushy layers. However, in such conditions, convective breakdown cannot affect the partially crystallized layer simply because the crystals are interconnected. In most cases of petrological interest, thick thermal boundary layers develop with crystals suspended in fluid.

A series of recent papers have addressed the general variable viscosity problem, and scaling laws for the unstable temperature difference and for the convective heat flux are available [Smith, 1988; Ogawa *et al.*, 1992; Davaille and Jaupart, 1993a].

Assuming that partially crystallized magma behaves like a slurry, it is possible to determine the characteristics of convection at the roof of a cooling chamber [Davaille and Jaupart, 1993b]. The viscosity contrast across the actively convecting melt takes a typical value of about 10, and the temperature scale for convection is determined by the viscosity function

$$\Delta T_v = \frac{\mu}{\frac{d\mu}{dT}} \quad (2)$$

where T is temperature and μ is the magma viscosity and where all variables are evaluated at the temperature of the interior.

Evidence for Thermal Convection

The occurrence of thermal convection in cooling magma bodies is supported by observations in the molten region of Hawaiian lava lakes [Wright and Okamura, 1977]. Cooling occurs first by conduction, and after a finite time, it is observed that temperature begins to fluctuate. The magnitude of these fluctuations varies systematically as a function of depth in the thermal boundary layer. This magnitude is zero in the coldest part of the thermal boundary layer where the crystal content is highest and increases with depth to a maximum value of about 20°C (Figure 9). These observations are similar to what happens in variable viscosity convection. The unstable boundary layer involves melt with crystal contents going from about 10%, the value in the lake interior, to about 22% [Davaille and Jaupart, 1993b]. In this case, the slurry approximation, such that crystals do not form an interconnected network, is justified. Later in the evolution of the magma body, the crystal content is high everywhere, and the slurry approximation is probably not valid. Other evidence for the existence of thermal convection is that in several large sills and intrusions, rock sequences which have crystallized from the roof are much thinner than those which formed at the floor (Figure 1). Thermal convection acts to keep the upper thermal boundary layer thin (Figure 4) [Worster *et al.*, 1990].

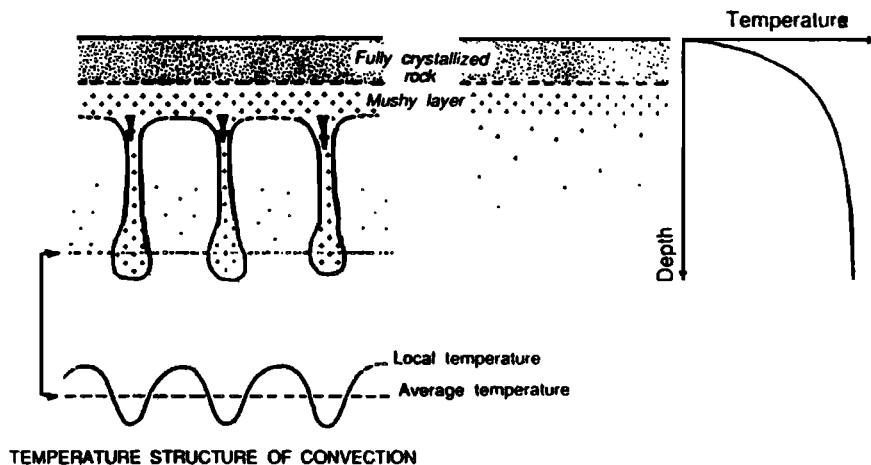


Figure 8. In the coldest part of the partially solidified boundary layer, the crystals may form a rigid network, or mush, whereas in the outer part a slurry may exist. In a purely conductive state, crystals that sediment into the main body of melt would be expected to remelt as fall into the hotter interior (right-hand panel). In the convective state, thermal plumes fall from the slurry part of the boundary layer and hence will contain crystals. This may lead to nucleation and growth of crystals in the main body of melt.

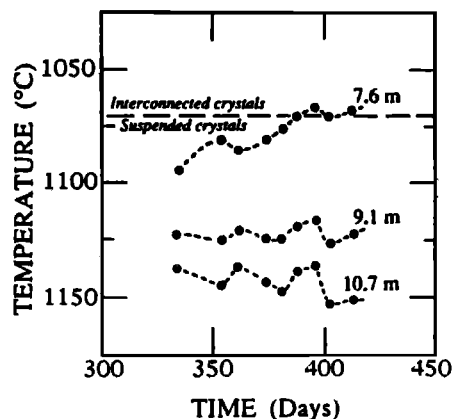


Figure 9. The structure of the upper boundary layer of a Hawaiian lava lake after Wright and Okamura [1977], showing the magnitude of temperature fluctuations that were observed that are associated with the onset of thermal convection in the melt.

Implications for Magma Chambers

The implications of the probable existence of thermal convection in magma chambers are essentially twofold. One is that the heat flux out of the convecting magma takes large values and may be enough to melt the overlying country rock [Aherm *et al.*, 1981; Huppert and Sparks, 1988]. Thus assimilation may indeed be achieved as the reservoir cools down and its amount is estimated. One obvious feature of this process is that the rate of melting decreases with time, as the reservoir becomes colder and more differentiated. The other implication pertains to where crystallization occurs. Thermal convection operates through the motion of plumes, which carry cooled magma downward, and hence also small crystals. In such a situation, the reservoir interior is well-mixed in the thermal sense, but the flow field is made up of plumelike upwellings and downwellings which essentially extend over the whole thickness of liquid. Thus crystals nucleated at the roof may end up at the floor, where they may mix with crystals grown in situ in different conditions [Brandeis and Jaupart, 1986]. Another important fact is that plume motions are intrinsically intermittent, which implies that the record of crystallization at the floor may be quite complex.

Compositional Convection

When cooling is from below at the floor of a magma chamber, an unstable density gradient can result if the residual liquid produced by crystallization is sufficiently depleted in the heavy component. The essence of the physics may be illustrated with a binary system, for which one can write the liquid density (ρ) as

$$\rho = \rho_0 [1 - \alpha(T - T_0) + \beta(C - C_0)] \quad (3)$$

where T is the temperature and C is the concentration of the heavier component; α , the thermal expansion coefficient, and β , its compositional analogue, are positive constants, and ρ_0 is a reference density at the temperature T_0 and composition C_0 . It is common that the compositional term dominates the thermal term. In several simple binary systems, the liquidus can be approximated as a straight line of constant slope Γ , as is often the case for aqueous solutions used in laboratory experiments:

$$T - T_s = \Gamma(C - C_s) \quad (4)$$

where T_s is the melting temperature of pure solid (composition C_s). Equation shows that, in the mushy layer, there is only one independent thermodynamic variable, and hence double-diffusive convection is not active. The density distribution in a mush cooled from below and in which the residual liquid is (compositionally) less dense than the starting liquid is shown in Figure 5.

Onset of Compositional Convection

One consequence of the formation of a mush, is that two independent convective instabilities are possible [Worster, 1992; Chen and Chen, 1991; Tait and Jaupart, 1992b]: that of the thin chemical boundary layer just ahead of the mush and that of the interstitial liquid (Figure 5). A local Rayleigh number criterion for the former instability can be written

$$Ra = \frac{g \Delta \rho_1 \delta^3}{D \mu} = Ra(\text{crit}) \quad (5)$$

where δ is the thickness of the boundary layer, $\Delta \rho_1$ is the difference in fluid density across the boundary layer, D is the chemical diffusivity, and μ is the melt viscosity; δ is proportional to the mush thickness before instability occurs, and hence this local criterion may be rewritten in terms of the mush thickness h [Tait and Jaupart, 1989]. A similar critical Rayleigh number criterion for the onset of instability of the interstitial fluid is

$$Ra_p = \frac{g \Delta \rho_2 \Pi_0 h}{\kappa \mu} = Ra_p(\text{crit}) \quad (6)$$

where h is the mush thickness, κ is the thermal diffusivity, $\Delta \rho_2$ is the density difference across the mushy layer, and Π_0 is a representative permeability. In the case of magmas, the chemical boundary layer becomes unstable first. Thus one expects a first phase of weak compositional convection driven by the small composition difference in the chemical boundary layer lying at the top of the mush, followed by a much stronger one involving interstitial melt in the whole mush. Once this second instability has taken place, the thin boundary layer at the top of the mush is destroyed and only the mushy layer mode of convection remains.

Flow Pattern of Compositional Convection

The key idea for understanding convection in the mush is that of porosity fluctuations. If chemical diffusion is neglected in the transport of solute, compositional convection will create porosity fluctuations because the effect on porosity of advection depends on the sign of the vertical velocity (W) [Flemings, 1974; Fowler, 1985; Tait and Jaupart, 1992a]. Dissolution and precipitation are the means by which the solute budget is closed. As fluid rises through the temperature gradient, it heats up, and downwelling fluid is cooled. The condition of local equilibrium requires the composition of the fluid to adjust by melting crystals in upwellings and crystallizing in downwellings. Near the edge of an upwelling, W is small, cooling dominates, and porosity tends to decrease, albeit less rapidly than in downwellings. At the centers of upwellings, W is largest and here advection can dominate and porosity can increase (Figure 10). Thus dissolution is most efficient at the center of an upwelling and increasingly less so towards the edge. Horizontal gradients of permeability develop, which lead to different flow rates in downwellings and upwellings.

Fully Developed Compositional Convection

We now present a synthesis of laboratory experiments and theoretical studies to document the development of convection

[Copley *et al.*, 1970; Sarazin and Hellawell, 1988; Beckermann and Viskanta, 1988; Chen and Chen, 1991; Neilson and Incropera, 1991; Tait and Jaupart, 1992a,b; Tait *et al.*, 1992; Amberg and Homsy, 1993]. As convection develops within the mush, porosity fluctuations grow and lead to flow focusing. The initial planform is an array of hexagonal cells with linear upwellings along the edges [Tait *et al.*, 1992]. The pattern changes by the edges of the hexagons getting choked with crystals (Figure 10). This process does not necessarily go to completion, but when it does, the ultimate form is that of a set of vertical channels or "chimneys" devoid of crystals at the nodes of the hexagons, where residual melt flows rapidly upward. The process of chimney formation has been observed in several metallic alloys and aqueous solutions and depends strongly on the microscopic structure of the mushy layer [Huppert and Hallworth, 1993].

Compositional convection causes the bulk liquid in the reservoir to evolve chemically. This may be estimated with various approximations of how the compositional budget is closed [Turner *et al.*, 1986; Woods and Huppert, 1989]. Detailed numerical calculations are not yet available over long time intervals because of the necessity to resolve both small-scale gradients of porosity and permeability, as well as large-scale chemical evolution [Neilson and Incropera, 1991]. As we shall

see, numerical calculations have been more extensive for sidewall cooling because they are more amenable to the nature and length scale of the flow.

A crucial effect of compositional convection is that flow in upwellings is considerably faster than that in downwellings. Thus thermodynamic conditions may be far from equilibrium in the upwellings. In the extreme case of chimneys, which are devoid of crystals, the melt rises without reacting with the cumulate layer (Figure 11). The compositional flux that is carried by convection causes the body of overlying melt to differentiate and hence its liquidus temperature to decrease. The effect of this, all else being equal, is to make the mush thicken more slowly (Figure 12a). Compositional convection carries a relatively low heat flux and so vertical profiles of temperature do not differ greatly from those that would be observed due to conduction alone. In consequence, the greater the compositional flux, the thinner is the mush for given thermal conditions (Figure 12b). The residual liquid flowing out of the mush is not on the liquidus because it has been heated up whilst rising through high-permeability channels. This fluid is mixed into the body of overlying liquid and hence will tend to generate a path of evolution in temperature/composition space that is different from the liquidus. That is to say that in the absence of other effects, this type of convection will cause the overlying liquid to become

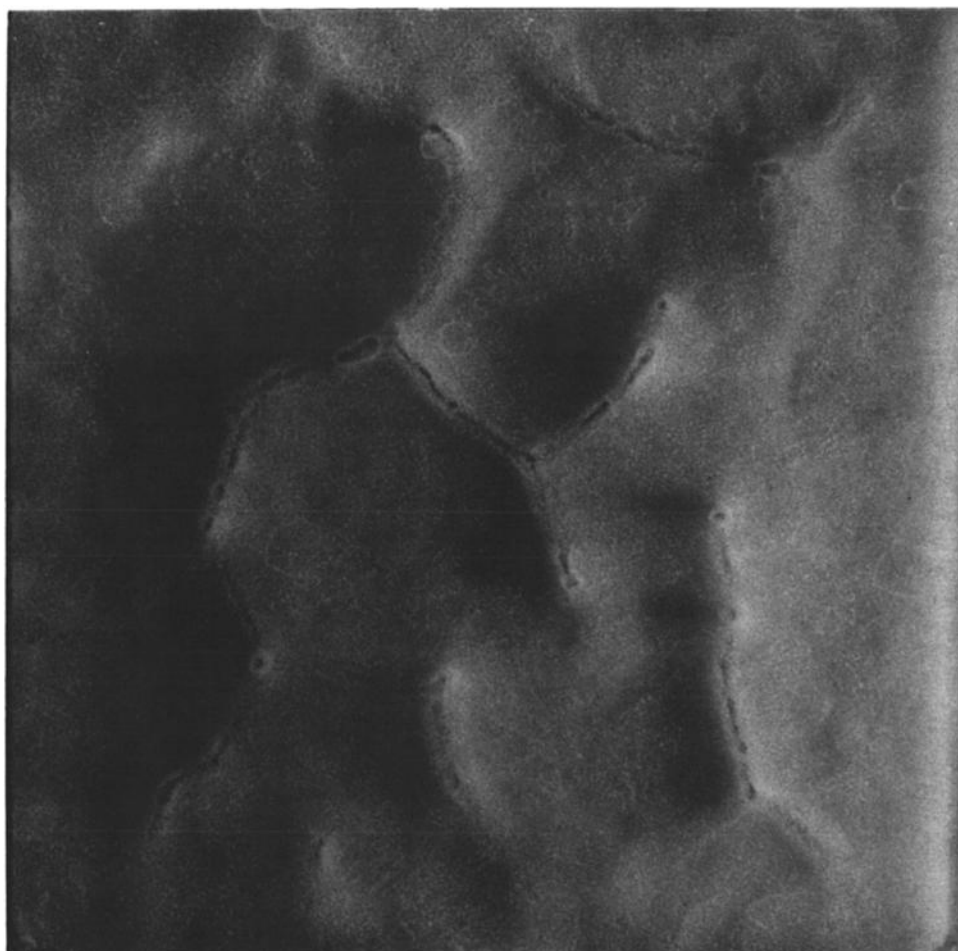


Figure 10. Photographs showing the development of porosity fluctuations in the mush formed in an experiment with aqueous NH_4Cl solution. A cellular pattern develops initially, with upwelling occurring along the margins of hexagonal cells. This ultimately becomes focused, and only cylindrical chimneys remain that form at the nodes of the hexagonal cells.



Figure 10. (continued)

superheated (Figure 13) (see also *Morse* [1986]). This is contrary to the assumption of many petrological models that the system is homogeneous at all times, with crystals distributed throughout the liquid.

Evidence From Intrusions

The onset of convective differentiation occurs at some finite time, when unstable density gradients have built up over a critical thickness. Thus one expects to see a transition in the rock sequence from early formed units to later units which formed when compositional convection was active. Studies of layered intrusions have usually focused on the cyclic units, without much attention being paid to their relationship to underlying basal zones. The first event discussed in any detail is the formation of the first olivine cumulate layer as a result of reinjection [*Irvine and Smith*, 1967; *Wilson*, 1982; *Raedeke and McCallum*, 1984]. This model requires the first crystal on the liquidus, olivine, to settle directly onto the floor. However, crystallization requires cooling and hence some boundary layer growth prior to reinjection and settling. Thus olivine should settle onto a cumulate layer containing minerals formed at temperature below the liquidus. Cooling from the bottom is demonstrated by the existence of chill zones and hornfels, showing that the magma lost heat to the underlying rock.

We now look in more detail at the internal stratigraphy of basal zones taking the Stillwater intrusion as an example [*Page*, 1979]. Adjacent to the basal contact with country rock, the rocks are fine-grained. Moving up the stratigraphy, they become coarser-grained, indicating that the cooling rate was decreasing, and pass through a sequence of plagioclase/pyroxene bearing rocks to those containing mainly pyroxene (Figure 14). The top

of the basal zone is taken to be where olivine becomes the main constituent of the rock [*Zientek et al.*, 1985]. Figure 14 shows an enlarged portion of a simplified phase diagram for such a mafic system and also the structure of the conductive boundary layer prior to the onset of compositional convection. Basal zones should record the structure of this boundary layer. Right adjacent to the contact, the initial liquid can be fully crystallized in situ before the onset of convection. If negligible mass transport occurs, the same final rock composition should be found at all levels. As we move up the stratigraphic sequence, however, the rocks become progressively enriched in pyroxene and olivine, which contain the high melting point components of the melt (Figure 14). This progression may be interpreted as recording the advection of fresh melt from the uncrystallized interior of the reservoir, which provides the required flux of high melting point components. Such a convective circulation is a prime candidate as a mechanism of adcumulus growth [*Kerr and Tait*, 1986; *Tait and Jaupart*, 1992b].

By continuity, the downward flow of melt from the overlying reservoir requires an upward flow of the residual melt. A related question is to find field evidence for this upward flow. An important observation is that adcumulate rocks appear to be homogeneous over large distances. Thus, if they are due to downward flow from the reservoir interior, the return flow must have been localized in narrow zones. The focused flow structures of mushy layer convection have the required properties. Both the Bushveld and Stillwater intrusions show vertical or subvertical structures which crosscut the horizontally layered rocks, such as iron-rich pegmatites derived from very evolved melts [*Viljoen and Scoon*, 1985; *Scoon and Mitchell*, 1994]. The rocks in these structures crystallized from more differentiated melts than did

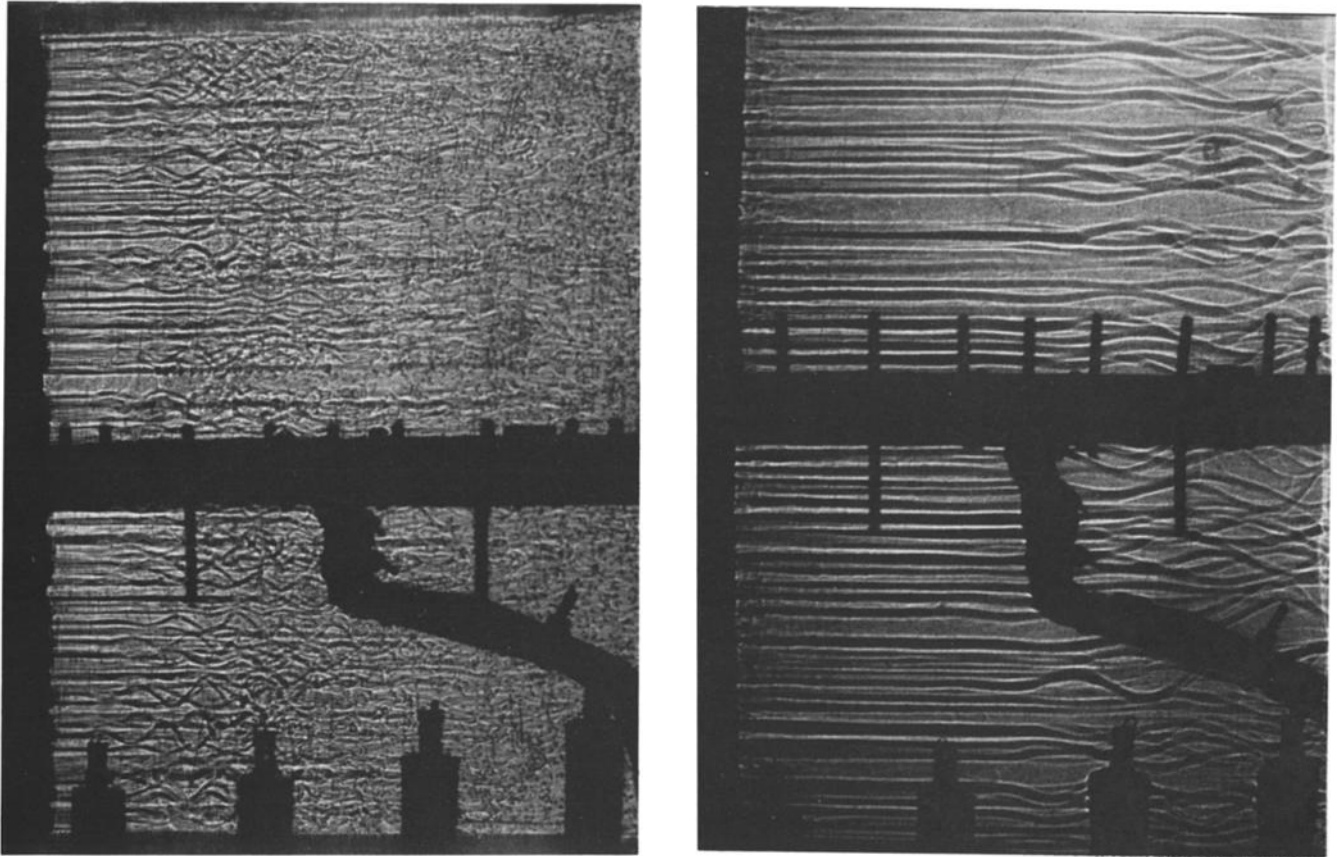


Figure 11. Fully developed compositional convection in which cylindrical plumes are rising from chimneys in the mushy region. The lower boundary of the tank is cooled, while the others are insulated. The flow pattern may be contrasted with that of thermal convection in Figure 7.

the host rocks which they cut, and furthermore, the higher up in the overall stratigraphy they are found, the more differentiated are their compositions.

The stratigraphic sequence of an intrusion records a temporal sequence of events in the chamber. The liquid part of the reservoir evolves chemically at a small rate because of its large size, which leads to the gradual variation of composition documented above (Figure 1). At the floor, the rate of thickening of the crystal pile diminishes with time, in part because the cooling rate is decreasing, and in part because the liquidus of the melt is decreasing with time. This implies that stratigraphic gradients in the rock composition cannot be simply interpreted in terms of rates of processes such as differentiation or assimilation in the liquid interior. For example, a steep gradient of isotopic composition, such as that found in the Bushveld complex [Sharpe, 1985] may not necessarily imply a rapid change, such as the arrival of a new batch of magma. It may also be interpreted as recording progressive assimilation at a time when the mush was growing at a very slow rate (Figure 12).

Another point concerns volatile species. The presence of very evolved (rhyolitic) melt compositions in the cold parts of the boundary layers of basaltic lava lakes implies that in the case where the pressure is high enough to have larger initial dissolved volatile contents in the melt, volatile species will be considerably enriched deep in a magmatic mushy layer. Thus fluids may be generated very early in the history of crystallization and are not necessarily to be associated only with the latest stages of differentiation of the whole magma body. Such fluids may not be

expected to remain in situ because of their large buoyancy and should pervade the rocks. Fluids are indeed present throughout the cumulate layers of the Stillwater and Bushveld complexes [Boudreau *et al.*, 1986; Mathez *et al.*, 1989].

A final implication is that the liquid part of the chamber can differentiate with crystallization occurring principally at the margins. Thus the liquid may evolve chemically with few if any crystals in suspension. This possibly explains why very evolved aphyric magmas can be erupted, with successive erupted liquids following a differentiation trend. The melt will have undergone crystallization not only of the phase or phases that are on the liquidus of the main body of melt but also of all of the phases that are crystallizing deep in the mush. A parameterized model by Langmuir [1989] shows some of the consequences of this type of remixing for the chemical evolution of a melt reservoir.

Convection Generated at Sidewalls

A vertical or sloping boundary differs from the case of a horizontal boundary in that there is no stable conductive state and hence no stability criterion analogous to the critical Rayleigh number. The situation is inherently unstable. Although the fundamental physical principles are the same, these flows have a different logic because of the different orientation of the density gradients with respect to gravity. We briefly discuss the flows that occur when light residual fluid is being produced by cooling at a vertical wall. The major characteristic of these flows is to generate a thin boundary layer of fluid that moves parallel to the

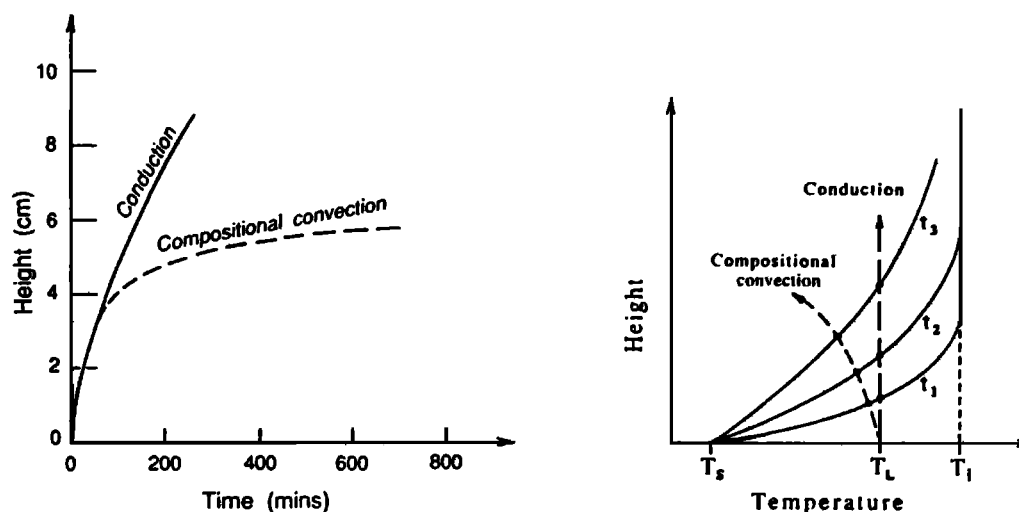


Figure 12. (a). The thickness of the mushy layer formed by cooling a liquid from below. The solid curve shows the results in purely conductive conditions as the viscosity of the solution was too high to allow convection to take place. The growth is proportional to the square root of time. The dashed curve shows the growth of the mush for the same experiment as that shown in Figure 10. After the initial conductive phase, compositional convection considerably slows the growth of the mush because it is causing the liquidus of the overlying solution to decrease with time. Imposed thermal conditions were the same for both experiments. (b) The solid lines show temperature profiles in the lower boundary layer for three different times. The vertical dashed line shows how the height of the mush will increase under conductive conditions (T_L is the liquidus temperature of the initial liquid). The decrease of the liquidus temperature that occurs under convective conditions leads to a thinner mush for a given temperature profile.

wall. Upon reaching the upper or lower boundary of the system, this fluid spreads out horizontally and then the slow vertical return flow takes place outside the boundary layer over the whole of the remaining surface area of the system. The major consequence of this is that vertical thermal and/or chemical stratification builds up in the reservoir of fluid.

Early analyses delimited the regimes that are possible for the purely diffusive case, i.e., without mushy layers forming [Nilson, 1985; Spera *et al.*, 1989; Jarvis, 1991]. Different regimes are possible because thermal diffusivity is very much greater than chemical diffusivity and because the temperature and composition are here taken to have opposing effects on the density of the fluid [Nilson *et al.*, 1985]. A thin, buoyant chemical boundary layer is embedded in a thick, dense thermal boundary layer. Depending on the numerical values of the diffusivities and on the relative sizes of the buoyancy forces that result from temperature and composition, one of three possible regimes will result. These are a thermally dominated regime in which all of the flow is downward because even the depleted buoyant fluid is dragged down by viscous coupling with the coldest fluid in the thermal boundary layer. A compositionally dominated regime is also possible in which the buoyant fluid closest to the wall drags upward the denser cold fluid. Finally, there is a so-called counterflow regime in which the compositionally light part of the boundary layer flows upward while the thermally heavy part flows downward. Figure 15 shows an example of such a flow and the induced chemical stratification in the liquid part of the reservoir.

The formation of a mush at a sidewall modifies the quantitative aspects of such flows, as it leads to vertical percolation of light residual fluid in the permeable part of the wall. Thermal downflow can only occur outside the mushy region. Experimental work has been largely qualitative [Huppert

et al., 1986a; Thompson and Szekely, 1988; Leitch, 1987] and has shown the rapid buildup of compositional stratification (Figure 16). Numerous numerical studies have also been carried out and are able to reproduce the main features of both the flows and their chemical effects [Szekely and Jassal, 1978; Maples and Poirier, 1984; Voller *et al.*, 1989; Oldenburg and Spera, 1991]. Numerical studies are made possible because these flows have larger characteristic length scales than in the case of horizontal crystallizing mushy layers.

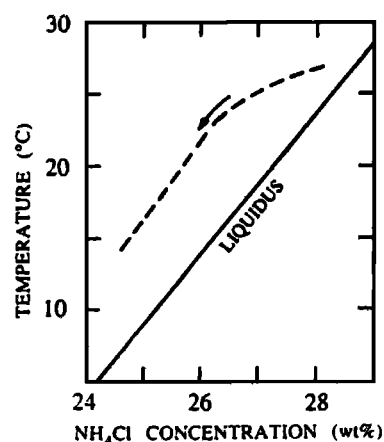


Figure 13. The evolution of the overlying reservoir of liquid in the phase diagram for aqueous NH_4Cl solutions for the same experiment featured in Figure 10. Note that although crystallization is taking place in the lower boundary layer, the overlying solution becomes increasingly superheated because as residual fluid rises up through the thermal boundary layer in the chimneys, it is heated above its liquidus by horizontal conduction of heat.

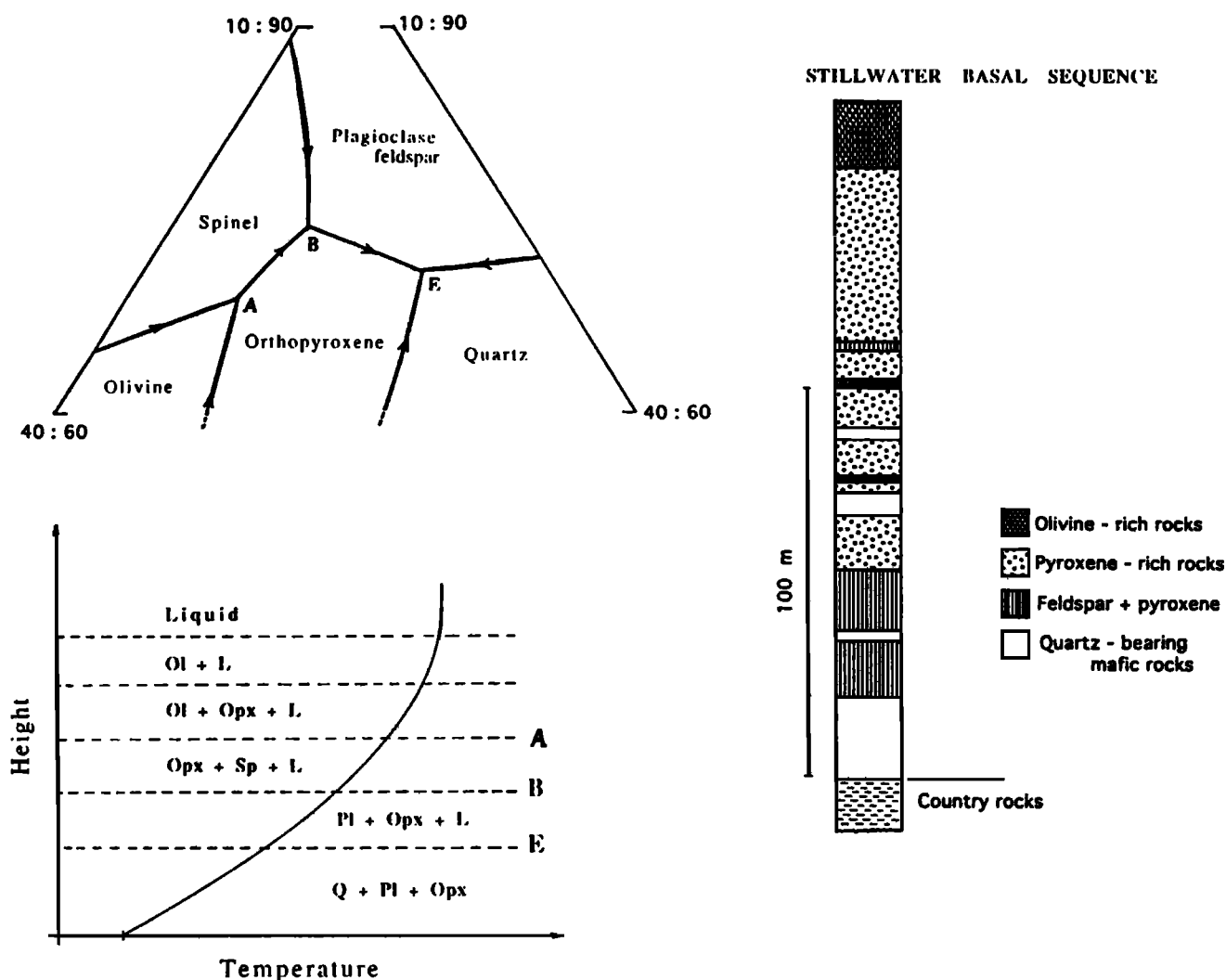


Figure 14. (left) An enlarged portion of the phase diagram forsterite-anorthite-silica at 10 kbar. The predicted structure of the conductive boundary layer prior to the onset of compositional convection for such a system is given, showing coexisting mineral phases for an initial liquid composition lying in the olivine field. (right) A simplified profile of the mineralogic variations in the Basal Zone of the Stillwater Complex, modified from Page [1979].

A common feature of volcanic eruptions is that the chemical composition of the lava changes with time as an eruption proceeds. One interpretation of such data is that the reservoir that is being tapped by the eruption is vertically zoned in composition because of fractional crystallisation [Michael, 1983; Worner and Schmincke, 1984; Bogaard and Schmincke, 1985]. Although lack of knowledge of parameters such as reservoir geometry make a quantitative comparison with chemical data extremely hard, the process of sidewall compositional convection is the best candidate as a mechanism for producing this type of stratification.

An interesting aspect of the dynamics of boundary layers at the sidewalls of crystallizing reservoirs is whether or not flow takes place in a predominantly laminar or turbulent regime, because the stratification that results, in terms of the form of the vertical profile of a chemical component, is not the same. In the case of turbulent flow, the boundary layer increases in horizontal

dimension with height due to entrainment of ambient fluid by eddies, whereas in the case of laminar flow, horizontal growth occurs only by molecular diffusion. In the case of the Laacher See magma chamber, flow is likely to have taken place in a laminar regime [Tait *et al.*, 1987].

Perhaps the major impact of these studies comes from the conclusion that evolved melt is collected at the top of the chamber almost immediately after cooling and boundary layer flow begin. We contrast this with a mechanism for generating evolved melt by "homogeneous" cooling of the entire chamber and crystal settling. A layer of evolved melt at the top of a chamber can be generated by the sidewall boundary layer mechanism in a time that is vastly shorter than the characteristic timescale of cooling for the whole body of liquid.

An important aspect of the dynamic behavior still to be addressed for sidewall flows concerns the possible interactions between these and flows generated from the horizontal boundary

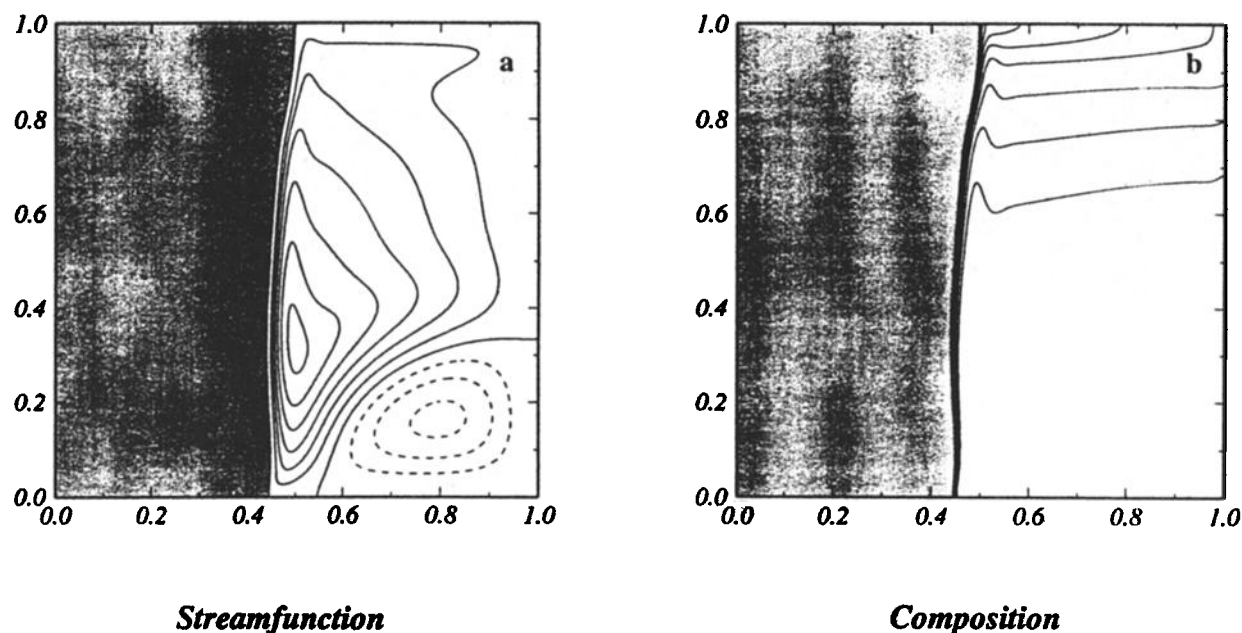


Figure 15. The flow pattern and composition isopleths formed by cooling, crystallization and release of light residual fluid at a vertical sidewall [from *Jarvis, 1991*]. The dashed lines show the presence of a thermal vortex that is rotating in the opposite sense to the boundary layer flow, as negative thermal buoyancy is also released.

layers at the floor and the roof in crystallizing systems. The effects of variable viscosity are also significant but do not change the qualitative picture [*Spera et al., 1982; Jarvis, 1991*].

Dynamics of Crystal Settling

Views on crystal settling have come around a full circle. This process was first thought to be dominant, was then neglected, and has been reevaluated recently. The first level of approximation is that crystals are passive tracers and do not modify the dynamics of a given large-scale convective flow. The second level of approximation requires specification of the behaviour of particles close to the solid margins.

Clearly, high flow velocities act to keep crystals from sedimenting, and this was the basis of early models where crystals remain suspended and well-mixed in the chamber interior. The chamber then evolves only in temperature and crystal content: the bulk composition is not changing. Eventually, crystals grow and the intensity of convection decreases, and sedimentation occurs when the settling velocity of the crystal exceeds the local fluid velocity [*Huppert and Sparks, 1980*]. Such a model is applicable to a well-stirred reservoir where turbulence is high even close to the boundaries. However, in most cases, and especially in reservoirs cooled from above, flow velocities take small values at the floor and do not allow suspension. This emphasizes that the flow field must be well known, especially in the vicinity of the solid margins. Theory, numerical calculations, and laboratory experiments have established how crystals settle in relatively simple flow fields, i.e., those that are dominated by large-scale eddies [*Maxey and Riley, 1983; Marsh and Maxey, 1985; Weinstein et al., 1988; Martin and Nokes, 1989*]. The main result is that close to a solid boundary, the flow velocity decreases to zero and hence is locally smaller than the crystal settling velocity. Thus sedimentation occurs even in an actively convecting system. The application of these studies to magmatic systems has been criticized on the grounds that they are not in the relevant

dynamic regime [*Tonks and Melosh, 1990; Solomatov and Stevenson, 1993*]. Some knowledge of the local structure of the convective flow near the boundary is required, and it is obvious from the previous sections that this is not easily specified. For thermal convection flows in magmatic conditions, it is likely that crystals settle almost as fast as in the absence of convection [*Martin and Nokes, 1989; Solomatov et al., 1993*]. However, if compositional convection is generated by crystallization at the floor, crystals may be kept in suspension in the chamber interior by the strong upward currents.

Regardless of the details of such models, the contribution of crystal settling to the growth of a cumulate pile is measured by the amount of crystals which are kept in suspension, which is necessarily limited by the heat budget of the reservoir. To crystallize, cooling must occur and latent heat must be released. The balance between the two may be achieved in several different regimes which lead to different kinds of size layering [*Hort et al., 1993; Jarvis and Woods, 1994*]. One general result is that crystal content stays small in the reservoir.

As the focus of research changed to boundary layers, where in situ nucleation and growth contribute a large amount of crystals, crystal settling was questioned as a mechanism responsible for the most important features of magma chambers. It was not thought, for example, to be a significant agent to explain layering. However, more recent experiments demonstrate that as soon as the crystal content exceeds a few percent, the interaction between the flow field and the crystals is strong [*Huppert et al., 1991; Koyaguchi et al., 1993; Sparks et al., 1993*]. The flow field becomes complicated, and the history of settling may become chaotic, with periods of steady sedimentation interspersed with periods of catastrophic deposition as the velocity of convection drops to small values. An initially homogeneous convecting layer may separate into two layers with different bulk densities, one devoid of crystals and the other with a larger concentration of crystals. The former layer has a small effective viscosity and convects rapidly, while the other layer is more viscous and convects at a smaller rate and hence with a smaller rate of heat

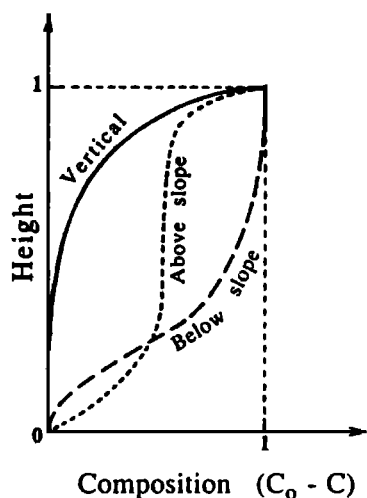


Figure 16. The vertical profiles of composition that developed in experiments in which aqueous solutions of simple salts were cooled at a vertical [after *Leitch*, 1987] or sloping boundary [after *Huppert et al.*, 1986a]. Fluid depleted in the heavy component that enters preferentially the crystals accumulates at the top of the tank due to the buoyant boundary layer flow. In the experiments with a sloping boundary, the profiles given are those that formed both when flow occurred beneath a cooled, downward facing boundary and above an upward facing boundary.

loss. One consequence is that a temperature difference builds up between the two layers, which may lead to a density reversal and an overturn of the layered system [Sparks *et al.*, 1993]. Thus crystal settling may, in fact, be responsible for large-scale flow phenomena and for intermittent deposition of crystals at the margins. Research in this area has barely begun, and one may expect significant developments in the future.

Open System Behavior

The issue of how crustal magma reservoirs are fed by melts ascending from a deeper level chamber or a mantle source is a fundamental one. The possibility that magma reservoirs receive periodic reinjections of new melt has therefore been much debated. Initially, this was not considered a key issue, probably because of the enormous influence of the early Skaergaard work, which suggested that this chamber had essentially crystallized as a closed system. However, the recognition of cyclic layering in large intrusions and some observations of magma mixing in volcanic suites have led to a number of studies of reinjection. Indeed, one may argue that fluid dynamic studies of reinjection are more complete than those of closed systems, and there is now a tendency to attribute a great many observations to reinjection as a wide range of behaviors seem to be possible, as a function of the different parameters of the problem. The work that has been done on the fluid mechanics of reinjection flows has paid very little attention to how the new liquid is accommodated into the finite volume of the preexisting reservoir. The two possibilities, generally speaking, are first that a volume of liquid equivalent to that reinjected can be ejected from the chamber by eruption and second that the volume of the chamber could be increased by deformation of the walls.

Reinjection

The most fundamental division of such flows comes from the relative densities of the resident and reinjected liquids. When the

reinjected liquid is the less dense of the two, a buoyant plume forms, whereas when the reinjected liquid is the denser of the two, the tendency is to form a layered system.

When the reinjected liquid forms a buoyant plume that rises into the chamber, the character of the mixing that occurs depends above all on the Reynolds number of the flow. The basic possibilities were mapped out by *Huppert et al.* [1986b]. At low Reynolds number, a laminar plume is formed which rises up to the top of the chamber and spreads out to form a discrete layer and mixing with the resident liquid is limited. On the other hand, a high Reynolds number plume will become turbulent and as it rises will entrain the resident liquid leading to intimate mixing. When such a turbulent plume reaches the top of the chamber, its mean composition will be somewhere between the initial compositions of the reinjected and resident liquids. It will spread out and generate a downward return flow and lead to the build up of stratification similar, from a fluid dynamic point of view, to the sidewall boundary layer flows discussed earlier. The characteristic "convex up" form of the profile of vertical stratification generated by a turbulent plume in a finite box [Baines and Turner, 1969] will result in this case.

Sparks *et al.* [1980] thought that this mechanism might explain the formation of mixed magmas observed at mid-ocean ridges. The extensive mixing that can be associated with a turbulent, buoyant plume may result in the scavenging of trace elements from a large body of melt. This has been proposed as a mechanism for the formation of stratigraphic horizons in large mafic intrusions that are rich in precious metals such as the Merensky Reef [Campbell *et al.*, 1983]. Note that the stratification that is established by these buoyant flows is such that the less differentiated and hotter magma will be uppermost in the chamber. This situation does not seem to be common, as in eruptive sequences it is generally found that cooler, less differentiated melt is the first to be erupted. Furthermore, it is probable that these flows are limited to a restricted compositional range. One circumstance in which turbulent injection of buoyant primitive magma seems plausible is in tholeiitic systems for which it is widely thought that a density maximum occurs in the differentiation sequence around the composition of ferrobasalt [Snyder *et al.*, 1993].

When new magma that is denser than the resident liquid enters a chamber and the Reynolds number of the flow is low, a two-layer system will form. Several studies have addressed the fluid dynamic behavior of these two-layer systems and how they may eventually break down such that the two layers become mixed [Huppert and Sparks, 1980; Huppert and Turner, 1981]. Both layers undergo thermal convection, with rapid heat transfer across the interface heating the upper layer and cooling the lower layer. Fractional crystallization takes place within the lower layer causing the composition of the liquid to evolve and hence its density to steadily decrease. At some point the density of the lower layer becomes equal to that of the upper layer, and the interface breaks down leading to rapid mixing of the two liquids. Mixing of the two liquids is delayed with respect to the replenishment event both in time and in the sense that considerable separate chemical evolution of the liquids can take place prior to mixing.

A number of variations on this basic theme have subsequently been carried out with the mixing taking place at a different moment or in different geometries according to the flow responsible. At low input rate, the incoming liquid becomes rapidly cooled and crystallized, releasing light liquid more or less continuously and preventing the formation of a two-layer system [Huppert *et al.*, 1982b]. If the resident liquid is stratified

in density, a two-layer system forms, but the final mixing is restricted in vertical extent. *Campbell and Turner* [1989] studied the effect on mixing of injecting a dense layer of liquid but with some appreciable amount of initial momentum. The incoming liquid formed a turbulent jet into which some of the resident liquid was entrained and mixed, before spreading out on the floor of the chamber. *Huppert et al.* [1982a] and *Thomas et al.* [1993] considered circumstances, thought to be relevant to calc-alkaline systems in which gas bubbles are exsolved in the lower layer, which generates a density reversal. Continuous mixing of the two liquids also occurs when there is a large viscosity contrast between the two layers, driven by strong viscous coupling at the interface.

Replenishment phenomena may be responsible for large-scale cyclic mineralogical layering in the cumulate sequence of large mafic intrusions [e.g., *Brown*, 1956; *Irvine and Smith*, 1967; *Campbell*, 1977; *Raedeke and McCallum*, 1984; *Wilson*, 1982]. The tendency has been to analyze how a reinjection event could lead to the formation of a specific mineralogical sequence in one cyclic layer [e.g., *Irvine* 1975; *Raedeke and McCallum*, 1984; *Tait*, 1985]. *Brandeis* [1992] has shown, however, using an overall mass balance for the whole sequence of cyclic layers in the specific case of the Stillwater Complex, that such models require more restrictive assumptions than had previously been realized on the amounts and compositions of the reinjected liquids.

Eruption

The effects of eruption on the internal structure of magma chambers has not been studied in much detail. Most research has looked at how the eruption process induces mixing in the volcanic conduit. For example, starting from a reservoir made of two superimposed layers of two different magmas, a basalt overlain by a less dense and more viscous granitic magma, say, withdrawal from the chamber leads to the lower layer being entrained into the volcanic conduit and mixed with the upper layer liquid [*Freundt and Tait*, 1986; *Blake and Ivey*, 1986; *Spera*, 1984]. Once the eruption has stopped, the layering remains unperturbed. With a continuously stratified reservoir, however, the eruption process acts to homogenize at least the upper part, which leads to a different compositional profile in the chamber and may induce a different crystallization sequence [*Spera et al.*, 1986].

Conclusion

This brief overview has dealt with results which, in the main, have been obtained over the last 10 years and shows that the door to an extremely rich physical world has been opened. The fluid dynamics of differentiation in magma chambers is a mature field, such that the main concepts are sufficiently developed to be used in the interpretation of petrological and geochemical data. The key lesson is perhaps that over a large volume, it is extremely difficult to prevent motions. Most static models are found to be unstable to small disturbances which are always present in natural systems. In fact, many dynamic models are also found to be unstable. Thus the ultimate test for a physical model is that of stability. As a consequence, it is of paramount importance to specify the model in such a way as to allow a stability analysis.

Another conclusion is that physical models have not yet been able to answer several important petrological questions, for example, how specific textures are generated. The reasons are

partly that the questions are too difficult for our present modeling abilities and partly that comprehensive phase diagrams are not available for many melts of geological interest.

Models have now been developed to a quite sophisticated level and may well correctly describe the essential physics. However, caution must be exercised when making direct comparisons with the geological record because accurate determinations of the physical properties of melts are too few. For example, we still rely on a restricted number of thermal conductivity determinations [*Murase and McBirney*, 1983; *Snyder et al.*, 1994].

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