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Solidification fronts and magmatic evolution

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Abstract

From G. F. Becker's and L. V. Pirsson's early enunciations linking the dynamics of magma chambers to the rock records of sills and plutons to this day, two features stand at the centre of nearly every magmatic process: solidification fronts and phenocrysts. The structure and behaviour of the envisioned solidification front, however, has been mostly that akin to non-silicate, non-multiply-saturated systems, which has led to confusion in appreciating its role in magmatic evolution. The common habit of intruding magmas to carry significant amounts of phenocrysts, which can lead to efficient fractionation, layering, and interstitial melt flow within extensive mush piles, when coupled with solidification fronts, allows a broad understanding of the processes leading to the rock records of sills and lava lakes. These same processes are fundamental to understanding all magmas.

The spatial manifestation of the liquidus and solidus is the Solidification Front (SF); all magmas, stationary or in transit, are encased by SFs. In the ideal case of an initially crystal-free, cooling magma, crystallinity increases from nucleation on the leading liquidus edge to a holocrystalline rock at the trailing solidus. The package of SF isotherms advances inward, thickening with time and, depending on location — roof, floor, or walls — and the initial crystallinity of the magma, is instrumental in controlling magmatic evolution. Bimodal volcanism as well as much of the structure of the oceanic crust may arise from the behaviour of SFs.

In mafic magmas, somewhere near a crystallinity (N) of 55% (vol), depending on the phase assemblage, the SF changes from a viscous fluid (suspension (0<N<25) and mush (25<N<55%)) to an elastic crystalline network (rigid crust (55<N<100%)) of some strength containing interstitial residual melt. With thickening of the roofward SF of some mafic magmas, the weight of the leading, viscous portion repeatedly tears the crust near N $\sim$55–60%, efficiently segregating the local residual melt into zones of interdigitating silicic lenses. This is SF instability (SFI), a process of possible importance in continental crust initiation and evolution, in producing silicic segregations in oceanic crust, and in recording the inability of the viscous part of the upper SF ever to detach wholly in typical ($<\sim$1 km) sheet-like magmas. These granophyric and pegmatitic segregations, individually reaching 1–2 m in thickness and 30–50 m in length, form thick ($\sim$50–75 m) zones that can be misconstrued as sandwich horizons where the last liquids might have accumulated. In effectively splitting the magma chemically and spatially, SFI is, in essence, a form of chaos (i.e. silicic chaos).

Differentiation of initially crystal-free, stationary magmas is limited to processes occurring *within* SFs, which operate in competition with the rate of inward advancement of solidification. Local processes operating on characteristic time scales longer than the time for the SF to advance a distance equal to its own thickness are suppressed. Enormous increases in viscosity outward within the viscous, leading portion of the SF efficiently partition the distribution of melt accessible to eruption. Eruptible melts lie essentially inward of the SF and are thus severely restricted in silica enrichment. The silica-enriched SFI melts are thus generally inaccessible to collection and eviction unless the host SF is reprocessed or "burned back" through, respectively, later regional magmatism or massive, late-stage re-injection. And because of large viscosity contrasts between SFI melts and host basalts, once freed, SFI melts are literally impossible to homogenize back into the system and may collect and compact against the roof to form large silicic masses. Unusually voluminous, bulbous masses of silicic granophyre present along, and sometimes warping, the roofs of large

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diabase sills may reflect collections of remobilized blobs of SFI melts. These bulbous masses may be later added to the continental crust through solid state creep.

In sheets made of phenocryst-rich, singly saturated magma, most phenocrysts are able through settling or floating to avoid capture by the advancing SFs. Significant differentiation is possible through extensive settling of initial phenocrysts and upward leakage of interstitial residual melt from the associated cumulate pile, which over-thickens the lower SF, greatly tipping the competitive edge against suppression of melt leakage by advancing solidification. Dense interstitial melts may similarly drain from roofward cumulates of light phenocrysts. The variation in crystal size and modal abundance in these cumulate piles are intimate records of prior crystallization, transport, and filling.

Magmas in transit erode SFs and thoroughly charge the magma with crystals, facilitating fractionation and differentiation, especially if the body occasionally comes to rest. The key to protracted differentiation through fractional crystallization is not crystallization in stationary, closed chambers, but the repeated transport and chambering of magma or the periodic resupply to chambers of phenocryst-rich magma. This is punctuated differentiation, which may be the general case. Close corollaries are that thick, closed sheets of initially crystal-free, multiply-saturated magma undergo precious little overall differentiation, and that deciphering the sequence and crystallinity, including in transit phenocryst entrainment, growth, and sorting, of the filling events is central to unravelling intrusive history.

Variations in temperature, whether on phase diagrams or in actual magmas, are intrinsically linked to commensurate variations in space and time in magmatic systems. The spectrum of all physical and chemical processes associated with magma is accordingly strongly partitioned in space and time.

The idea of a magma chamber as a vat of low crystallinity melt crystallizing everywhere within and differentiating through crystal settling is unrealistic. A magma chamber formed of any number of crystal-laden inputs, encased by inward-propagating, dynamic solidification fronts, and where significant differentiation is tied to the dynamics of late-stage, interstitial melt within extensive mush piles is more in accord with the rock record.

KEYWORDS: solidification fronts, magmas, phenocrysts, crystallization.

. . . the writer pointed out the special importance of two stages, the early stage of crystallization, when crystal settling may occur, and the late stage of crystallization, when squeezing out of residual liquid may occur. N. L. Bowen, 1919

1 Introduction

A SOLIDIFICATION FRONT is the partially molten to partially crystalline mass of magma spatially-contained between the solidus and near-liquidus isotherms. All magmatic bodies have solidification fronts, however thick or thin, that grow inward and progressively thicken with cooling and have physical divisions based on local rheology and strength. Such fronts are not unique to magma, for almost every manufactured material involves solidification in some way. The quality and detailed properties of metal alloys, for example, depend critically on a detailed knowledge of the dynamics of so-called mush fronts (e.g. Beckermann and Viskanta, 1993). And so it is also with magma. The evolution of magma is intimately tied to the physical and chemical dynamics of solidification fronts. It may not be too much to say that the evolution of most magmas involves an understanding of the growth and erosion of solidification fronts during transport and holding of magma. The entrainment, sorting, and re-equilibration of phenocrystic debris in ascending magma, the interaction of solidification fronts with phenocryst-laden emplacements and eruptions, and the behaviour of residual melt within these fronts are central to understanding magmatism.

Magmatic solidification fronts are distinct in size and texture over those of most other systems, and it is these features, coupled with the phenocrystic or particulate nature of magma itself, that allows direct insight on the evolution of almost any magmatic system. Yet history shows that our concept of what magma chambers are and how they function clearly stems from very early concepts of solidification fronts and magma crystallinity that are not akin to silicate magmas but to systems of metal alloys and aqueous solutions. Crucible melting experiments on rocks, although giving the impression, perhaps inadvertently, of accurately portraying a magma chamber on a small scale, actually pertain to exceedingly local regions of solidification fronts (see Fig. 1). The resulting phase diagrams thus contain no sense of space and time within the magmatic system itself. Through the linkage of temperature, space, and time, solidification fronts emerge as dynamic features common to the understanding of large chambers, sills, dykes, lava lakes, lavas, and even volcanism itself.

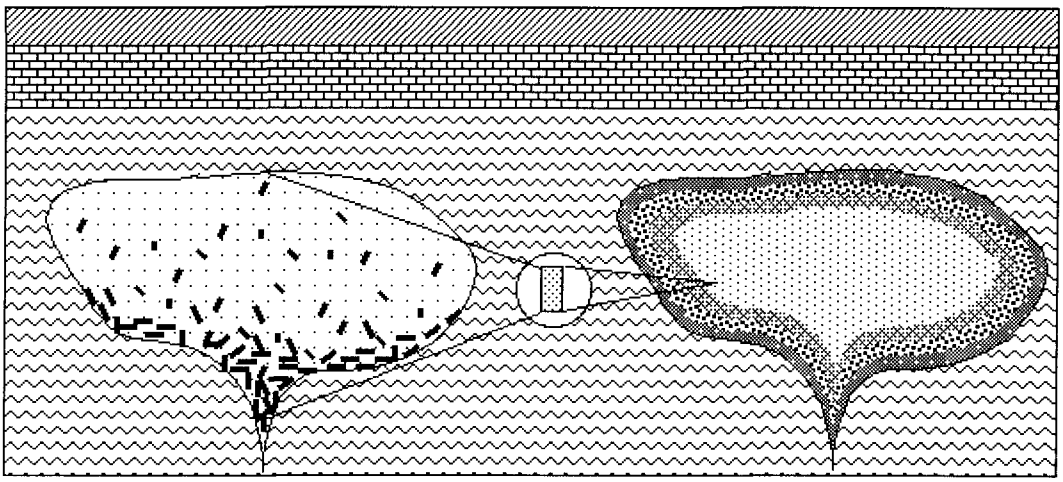

FIG. 1. A schematic illustration of two ideas of magma chambers and their conceptual relation to a crucible used in experimental studies of phase equilibria (inset). The chamber on the left crystallizes throughout and deposits crystals on the floor; it cools mainly from the roof and crystallizes from the floor up. The chamber on the right is surrounded by mushy solidification fronts, cooling and crystallization are approximately evenly distributed about the margins of the body; thick piles of crystals on the floor, should they arise, come from phenocrysts injected with the magma itself.

Recognition of the central role of solidification fronts has been hindered because of the great difficulty of observing magma directly in its natural subsurface setting. Delineating the fundamental principles of solidification front behaviour is possible, however, through the characterization of magma in simple, perhaps unique, situations using complementary field and laboratory studies. My aim here is to use a 100 year record of field and laboratory petrology to weave an unified perspective of some styles of magmatism based on the single concept of solidification front dynamics.

I begin with a description of the rheological character of solidification fronts and how they contrast in silicate and non-silicate systems. The dynamic character of solidification fronts is then described as revealed by the solidification history of various sills and lavas. Next I give an historical perspective of the development of the concept of magma chamber dynamics as motivated by the rock record. This is intended not simply for historical purposes, but more so for identification of the key ideas and facts upon which rest the origins of our fundamental magmatic principles and beliefs. This is followed by an investigation of the origin, nature, and abundance of phenocrysts in magmatic flows, which is a long under-appreciated phenomenon. The remainder of this work presents a case for interpreting the rock record of magma, from the largest to the smallest bodies, through the occurrence of phenocrysts and the dynamics of solidification fronts.

2 Rheological structure of solidification fronts

2.1 *Solidification Fronts.* Magmatic solidification fronts are distinct from those of metal alloys and aqueous solutions. Metal alloys are characterized by single dendritic crystals often spanning the entire solidification front with dendrite tips at the liquidus and dendrite bases at the solidus (see Fig. 2). Directional dendritic crystal growth is uncommon in magmas; it is seen in the spinifex texture of olivine in komatiites, perhaps in plagioclase of the border group of Skaergaard, and possibly in some other places. By far the most common form of crystal growth in magmas is as single, more or less equant, crystals. Crystal size, which is relatively restricted in igneous rocks ($\sim$0.1$-$10 mm), is always much smaller than the size of the solidification front itself. Silicate crystals also are not firmly attached to each other everywhere in the front, but only where the concentration of solids reaches a point of maximum packing. These major differences between magmatic and dendritic solidification fronts reflect fundamental differences in the dynamics of phase equilibria and crystal growth in these systems, points I will return to after more fully introducing solidification fronts.

In traversing silicate solidification fronts magma goes from fully solid to fully liquid (see Fig. 3). With cooling, the defining bundle of isotherms propagates inward, meeting in the interior of the body at the point of final solidification. Although thin and sharp just after emplacement, with time the solidification front thickens in proportion to the square root of

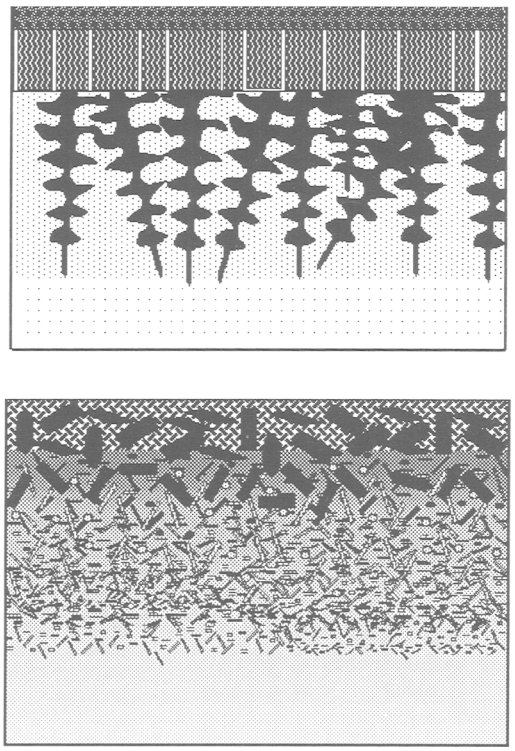

FIG. 2. Two examples of solidification fronts. The upper example is a dendrite front common in singly saturated aqueous and metal alloy solutions. The lower example is dominated by many small crystals and is typical of multiply-saturated magmas. Melt can flow freely in and amongst the crystals of the dendrite front but not in the magmatic front.

time, as is well known, for example, from the drilling of Hawaiian lava lakes (e.g. Wright and Okamura, 1977) and solidification theory in general. In an idealized magma emplaced free of crystals, crystals nucleate and begin growing at the inwardly advancing liquidus; the viscosity there is essentially that of the crystal-free magma and only when the crystallinity increases to about 25% does the viscosity increase by a factor of about ten (Shaw, 1969; Ryerson *et al.*, 1988).

At the trailing edge of the solidification front, near the solidus, the viscosity is enormous as the rock is nearly solid. Inward from this point crystallinity decreases, but it is only where it decreases to about 50–55% that the viscosity begins to decrease dramatically. Here the magma undergoes a transition from a partially molten solid to a mushy liquid (Fig. 2). This is the point of critical crystallinity, marking the region of maximum packing of the solids, where at all higher crystallinities the crystals form an interlocking network of considerable strength. At still lower crystallinities, the network of crystals is not fully interlocking and the material behaves instead as a crystal-laden fluid possessing a large effective viscosity; crystals and bubbles are, nevertheless, unable to move freely relative to one another and the region behaves as a mush. This mush-like behaviour continues inward with decreasing crystallinity (i.e. increasing temperature) until reaching the earlier-described region where the effective viscosity is within a factor of about ten of the viscosity of the crystal-free magma. Within this innermost zone crystallinity (N) is low ($0<N<25\%$) and the relatively sparse crystals are able to move freely relative to one another with little hindrance; the magma here is a suspension. (The term 'slurry' is purposely avoided due to its common characterization of fluids consisting of large concentrations of very small particles as in mortar.) These various rheological divisions can be summarized and classified as follows (Marsh, 1988).

1. **Rigid Crust:** $N>50-55\%$. Bounded by the solidus ($N = 100\%$) and the point of critical crystallinity ($N = 50–55\%$). This is the drillable portion of the solidification front commonly referred to as the "crust" in studies of Hawaiian lava lakes; it has significant strength, and in some respects is the effective wall rock of the magma.

2. **Critical Crystallinity Region:** $N\sim50-55\%$. The relatively abrupt transition from an interlocking assemblage of some strength to a high viscosity mush.

3. **Mush Zone:** $25< N<50-55\%$. Bounded by the Capture Front ($N = 25\%$) and the region of critical crystallinity at maximum packing of the solids where $N = 50–55\%$; crystal migration is difficult due to mutual hindrance and once a crystal reaches this zone it is forever trapped in the solidification front.

4. **Capture Front:** $N\sim25\%$. The region marking the transition from the leading suspension zone to the trailing mush zone and behind which settling crystals are unlikely ever to escape the solidification front.

5. **Suspension Zone:** $0<N<25\%$. Beginning at the liquidus, N=0, and ending at a crystallinity of about $N = 25\%$; small, sparse crystals can move freely relative to one another; the effective viscosity increases by a factor of about ten across this zone.

Two aspects of these subdivisions are important to emphasize. First, these are generalized definitions and are, in reality, closely tied to the specific phase equilibria and dynamics of the magma under study; the defining crystallinities will vary with magma type and style of local deformation. Second, the entire solidification front, which is always present, is dynamic; it propagates inward at an ever decreasing rate and thickens with time. During reinjection of

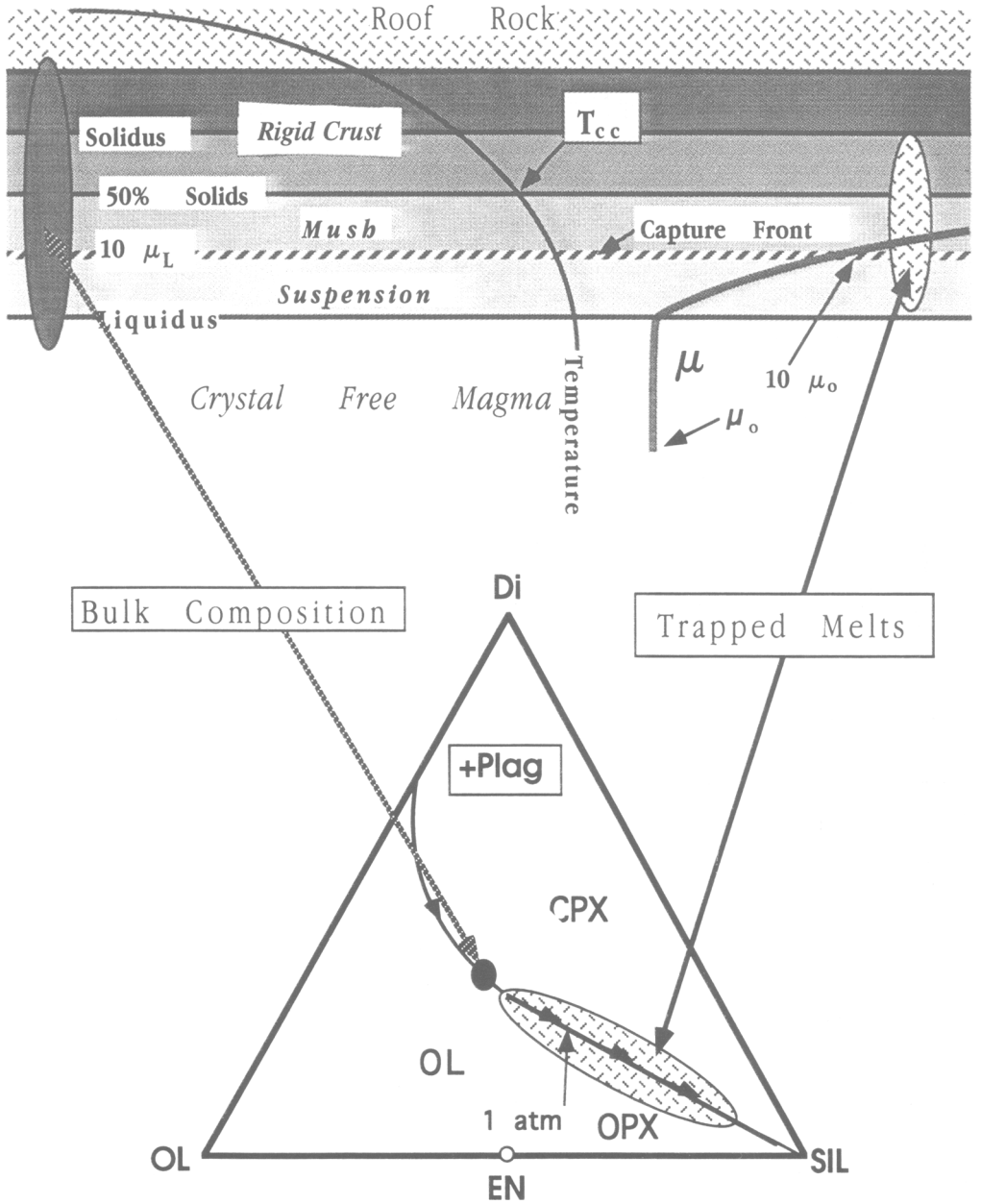

FIG. 3. The nomenclature of a solidification front for a basaltic magma (upper), showing the relative positions of the Suspension, Mush and Rigid Crust zones. The Capture Front separates the Suspension and Mush zones. The chemical composition of the associated melts, beyond and within the front, are indicated on the normative ternary system. The composition of the melt *within* the front changes markedly with increasing crystallization, but because this melt is normally unextractable the bulk composition of the system, in this limit, is unaffected by crystallization.

Chemical variations in the ternary system are intimately linked to *spatial* variations in the magmatic system.

hotter magma the forward progress of the solidification front may be arrested or even reversed. And because it is dynamic it initiates and controls all physical and chemical processes occurring within itself. We shall see presently that the distribution of phenocrysts in sills and the formation of silicic segregations are integral reflections of solidification front behaviour.

It is also important to note that the transition from Mush to Rigid Crust is basically defined in a mechanical sense. At crystallinities below 50–55% the assemblage is no longer drillable, nor are magmas containing more than this quantity of phenocrysts ever erupted as lavas (Marsh, 1981). The strain rates associated with drilling and eruption are clearly extremes relative to *in situ* magmatic processes. The transition to a crystal-locked magma may thus occur at significantly smaller crystallinities. The exact crystallinity associated with the onset of significant strength, therefore, depends on the style and habit of crystal growth, which depends on whether the system is singly or multiply saturated, and on the dynamic behaviour of the magma itself.

2.2 *Roles of multiple and single saturation.* Magmatic crystals often grow under conditions of multiple saturation, which inhibits development of thick chemical boundary layers and keeps crystal size small and uniform. In the extreme of a single component system (e.g. SiO_2), the liquidus and solidus, aside from a small undercooling, are one and the same and the solidification front is a sharp, planar interface. Crystal size is set by the nucleation and growth rates, which are linked to the rate of cooling.

In binary eutectic systems, with or without solid solution, the liquidus and solidus are separated. Crystallization is marked by growth of a single solid phase, often for a long temperature drop, before appearance of the other solid phase. In a very real sense this is analogous to the growth of olivine in dry picrite and plagioclase in dry high alumina basalt, both compositions are characterized by a large separation in temperature between these liquidus phases and the appearance of the next phase. During this period of growth of the first phase, the components rejected by the growing crystals collect about the crystal, with diffusion or fluid flow establishing some form of quasi-steady state exchange to sustain growth. Other phases cannot nucleate and grow in this rejected boundary layer of enriched fluid because their phase boundaries are at much lower temperatures.

Depending on the rate of diffusion and the exact composition of the envelope of rejected components, the boundary layer itself may become gravitationally unstable and separate from the growing crystal, being replaced by nutrient-rich melt which sustains growth. Crystals grow until they impinge on one another and block the supply of nutrients. Under the right conditions there is almost no limit to crystal size, and crystal size is essentially nutrient limited. If boundary layer separation does not occur, and diffusion cannot supply nutrients fast enough to match the growth rate, as induced by the rate of cooling, increased nucleation will limit crystal size as in chilled margins. This is the common form of crystal growth in dendrite systems of metals and aqueous solutions, mentioned above, whereby directed growth crystal tips are continually kept in contact with fresh melt. The non-dendritic, but more or less equant growth of olivine and plagioclase in picrite and high alumina basalts suggests systems characterized by low nucleation rates and single phase growth.

In multiply saturated or near-multiply saturated systems, the nucleation of any one solid phase produces rejected components useful to other phases, spawning local oversaturations, and stimulating nucleation. Nucleation occurs in response to crystal growth, which stabilizes the associated chemical boundary layer and produces interlocking, ophitic to subophitic textures of small crystals. It is remarkable in this respect that this texture and crystal size in a relatively broad range of tholeiitic systems is nearly invariant with body size. Diabase sills ranging in thickness from 20 to 500 m have almost indistinguishable textures; only the phenocrysts brought in with the initial magma stand out as disproportionately large (the character of phenocrysts will be considered later). It is however curious in this regard that phenocryst-free bodies, essentially identical in size and composition, sometimes have much coarser textures than diabase; this reflects an effect due to more than just cooling rate.

The style of saturation has a profound influence on the nature and behaviour of the solidification front. In multiply saturated basalts, which most high level basalts are, crystals grow in three dimensional clusters and strings (e.g. Vogt, 1921), probably forming a loose macrostructure or ganglia, which significantly affects both the physical and chemical properties of the magma and the fluid-solid interactions. The rheology of the magma depends on the characteristic length scale and framework of the ganglia or strings of crystal clusters, and is much less well represented by a fluid containing a non-reactive assembly of spheres, which, *per force*, is the type fluid most readily known. The chemical boundary layers associated with crystal strings are exceedingly small and possess little collective buoyancy, and the settling rates of the ganglia depend on their overall structure. This delicate structure can be destroyed by vigorous stirring, as in the dendrite experiments of Paradies and

Glicksmann (1992), and perhaps also in the picrite experiments of Ryerson *et al.* (1988) where the viscosity hardly increases even after 60% crystallization. Thus the strain rate and style of deformation of the magma may affect the length scales and habits of the crystal strings. The transitions from Suspension to Mush and Mush to Rigid Crust move to smaller overall crystallinities with decreasing rates of strain.

Overall it follows that great caution be exercised in applying results from singly-saturated crystallization in aqueous solutions to basaltic magmas.

The dynamic behaviour of solidification fronts explains a host of observations on the general petrology of sills, and also sets the stage for appreciating the historical development of magma chamber dynamics.

3 Dynamics of solidification fronts

Accepting [post-emplacement] *crystallization alone, it would also be difficult to explain why the lower layer of femic rock is so very thick and the upper one so very thin.*

L.V. Pirsson, 1905, on Shonkin Sag laccolith.

The differentiation of magma by crystal fractionation is central to the chemical diversity of igneous rocks, and crystals are central to the definition and dynamics of solidification fronts. In newly injected magma two features are therefore crucial to controlling differentiation: phenocrysts and solidification fronts. There are two conditions of fundamental importance: magma emplaced carrying phenocrysts and magma emplaced free of phenocrysts.

In magma with phenocrysts, differentiation is controlled by a balance between capture of crystals by the solidification front and crystal settling; in all but the thinnest of sheets massive crystal separation is inevitable. The remaining magma, now phenocryst-free, is less viscous, more eruptible, and has changed its composition almost discontinuously. This is **punctuated differentiation**.

In initially phenocryst-free magma, differentiation is confined to the interstitial liquids of the solidification fronts. The crystal-free central region does not differentiate. This is the **Null Hypothesis**. Here we discuss these processes, along with flow differentiation, to reveal some of the dynamic behaviour of solidification fronts, and also to show how this behaviour controls differentiation in sills and at Kilauea volcano, Hawaii.

3.1 *Punctuated differentiation and S-profiles.* The mafic, potassic phonolite emplaced to form Shonkin Sag laccolith (70 m thick by 3 km wide) of north-central Montana carried 35 vol.% phenocrysts of mainly clinopyroxene ($\sim$ 25–30%), and some olivine, leucite, and phlogopitic mica (Hurlbut, 1939; Congdon, 1990; Marsh *et al.*, 1991). Instead of simply settling to the floor to form a single thick cumulate pile, the downward propagating upper solidification front captured a great many phenocrysts along the upper margin. Slowing of the rate of cooling with time progressively slowed growth of the solidification front such that more and more crystals escaped capture (see Figure 4). The final result left an S-shaped vertical modal distribution of mafic phenocrysts. Moreover, because leucite floated in this magma, it collected in the upper parts of the body, but was also captured near the lower contact by the lower, upward propagating solidification front.

The seminal recognition of the fundamental dynamic interaction between solidification fronts and settling phenocrysts was put forth by Pirsson (1905) to explain the disparity in phenocryst distribution at Shonkin Sag. And although generally recognized early on to be of importance (e.g. Bowen, 1915*c*), phenocryst capture has not become a central tenet of petrology. The first quantification of this process was by Osborne and Roberts (1931). Since that time the mechanism has been repeatedly rediscovered by Jaeger and Joplin (1955), Gray and Crain (1969), Fujii (1974), and Marsh (1988*a*), and its effects have been noted in a great number of sills (Mangan and Marsh, 1992).

The redistribution of phenocrysts after emplacement can strongly differentiate the body chemically through crystal fractionation. This sudden change in composition of the eruptible portion of the body is **punctuated differentiation**. The degree of differentiation depends on the initial content and distribution of phenocrysts. At Shonkin Sag, for example, post-emplacement differentiation is explainable simply in terms of phenocryst settling and phenocryst capture (Hurlbut, 1939; Marsh *et al.*, 1991). Post emplacement crystallization, although important in understanding the final textures, is in essence a second order chemical effect.

3.2 *The Null Hypothesis and I-profiles.* In sills where the first magma does not contain phenocrysts, crystals grow within the solidification fronts. They conceivably can become large enough to settle fast enough to escape the Capture Front. Having escaped, these crystals now enter the inner, hottest portion of the magma and are susceptible to resorption. Jaeger and Joplin (1955) and Mangan and Marsh (1992) modeled this process as single crystals escaping the front. Jaeger and Joplin estimated the resorption time of crystals of a given size relative to their settling rate. Mangan and Marsh considered crystal growth beginning at the liquidus and continuing until approach of the Capture Front whereupon the crystals escape and settle according to Stokes' law; resorption

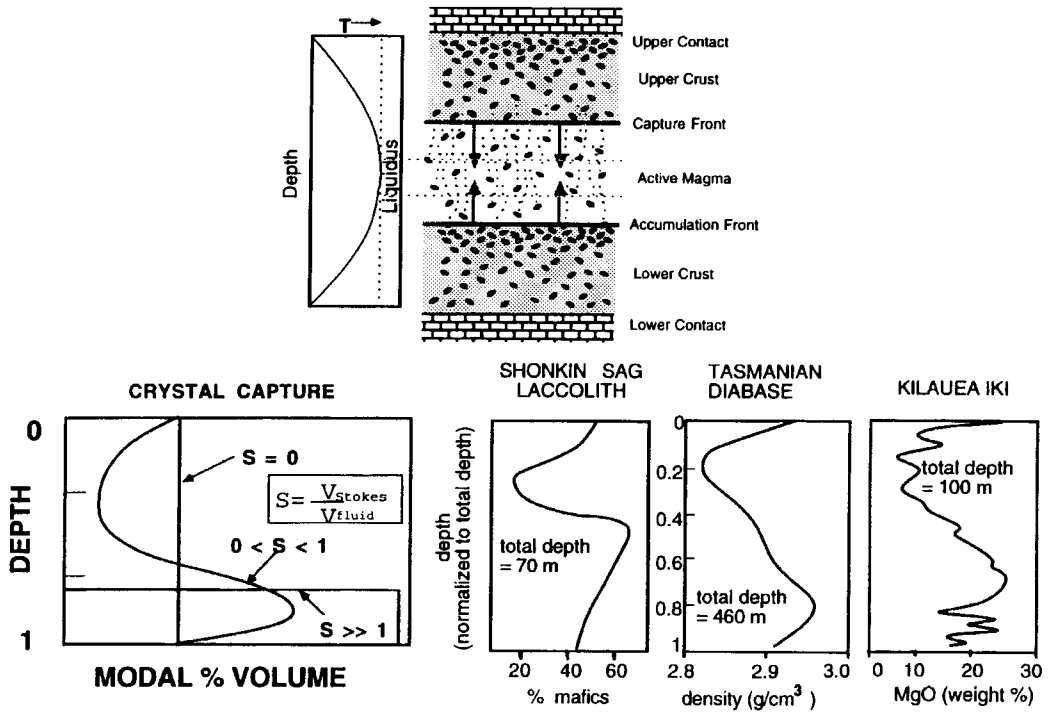

FIG. 4. The process of phenocryst capture by inward-moving solidification fronts, which leads to S-shaped profiles of modal mineralogy in sills. The upper diagram shows the idealized process with the possible end member modal distributions given by the sketch on the lower left. The values of the parameter S denote the relative importance of the rate of crystal settling to the rate of advance of the front. The remaining diagrams on the lower right show modal, density and chemical variations observed in some sheet-like bodies that may reflect this process of capture (after Marsh, 1988).

begins upon crossing the liquidus. Both studies found it unlikely, except in the latest stages of sill solidification, for crystals to survive and be incorporated into the lower front. In essence crystallization and differentiation within the solidification front is self-contained.

Plumes of material from the leading region, Suspension Zone, of the upper solidification front can descend much faster than single crystals, cool much slower, and deposit crystals on the lower front (Brandeis and Jaupart, 1986: Marsh, 1988*a*). But these small crystals would be deposited in the Suspension Zone of the lower front, under the same phase equilibria, and simply increase the volume of the lower front without causing any crystal fractionation. Even if whole blocks fall from the upper front there will be no obvious chemical differentiation, for this simply adds more of the same to the lower front.

The consideration of many other processes leads to a general conclusion: The central or eruptible melts of sill-like bodies formed of magma initially free of phenocrysts undergo very little, if any, differentiation. This is the **Null Hypothesis**.

Although Pirsson (1905) early on noted that,

> *Nothing is more common than to observe great vertical thicknesses of igneous rocks exposed by erosion, often for several thousand feet, and find that they are of uniform character throughout from top to bottom.,*

there is precious little chemical information available on the bodies of basaltic composition where this hypothesis is most readily testable. That is, if no effect is seen in basaltic magma, it is highly likely to also be true for more silicic magma.

Stratigraphic and chemical profiles through the Peneplain sill of the Ferrar dolerites of the Dry Valley Region of Antarctica are shown in Fig. 5 (Gunn, 1966; Marsh and Wheelock, 1994). This 330 m thick, quartz tholeiite sill shows no sign of containing any phenocrysts upon formation, is exceedingly well exposed, and can be traced for about 100 km. The sill was sampled in the area of Solitary Rocks along the Taylor glacier where it intrudes along and above the contact between the Beacon sandstone and the Irizar granite. Although here the upper contact has been eroded away, in comparison with other complete sections apparently only about 20 m is missing.

The rock itself is medium grained dolerite of a uniform texture and color from top to bottom except for an horizon of silicic segregations near the top. These segregations represent tears in the upper solidification front in response to sagging under gravity. As the tears open, local interstitial melt is drawn into the tears (see below). They clearly do not represent a final solidification or 'sandwich' horizon (Wheelock and Marsh, 1993). Some irregularity in composition near the sill base could reflect interaction with the wall rock during sill formation.

The central portion of the sill is strikingly uniform in both major and trace element (not shown here) composition. There is no sign whatsoever of settling of any post-emplacement crystals to produce chemical differentiation. This is in sharp contrast to the much smaller 70 m thick Shonkin Sag laccolith. Although lavas and sills of apparent homogeneous composition have not been popular to study, they may be more common than expected (Mangan and Marsh, 1992). The evidence so far thus seems clear: no phenocrysts, no post-emplacement differentiation.

The understanding of what phenocrysts are and how they behave is central to understanding magmatic evolution, which is an issue we will return to later. There is yet another class of sills that show strong concentrations of phenocrysts at a particular horizon superimposed on a more or less uniform background composition.

3.3 *Flow differentiation and D-profiles.* Heavy particles carried by fluids flowing up pipes and channels migrate towards the centre of the flow (Segre and Silberberg, 1962; Saffman, 1965; Leal, 1980); this segregation of crystals and liquid is flow differentiation. Flow differentiation is common in phenocryst-laden magmatic flows and has been described in some detail by Baragar (1960), Drever and Johnston (1958; 1967), Upton and Wadsworth (1967), Simkin (1967), Gibb (1968), and Gibb and Henderson (1989), among others. Experiments demonstrating flow differentiation during simulated ascent and emplacement of magma have been performed by Bhattacharji and Smith

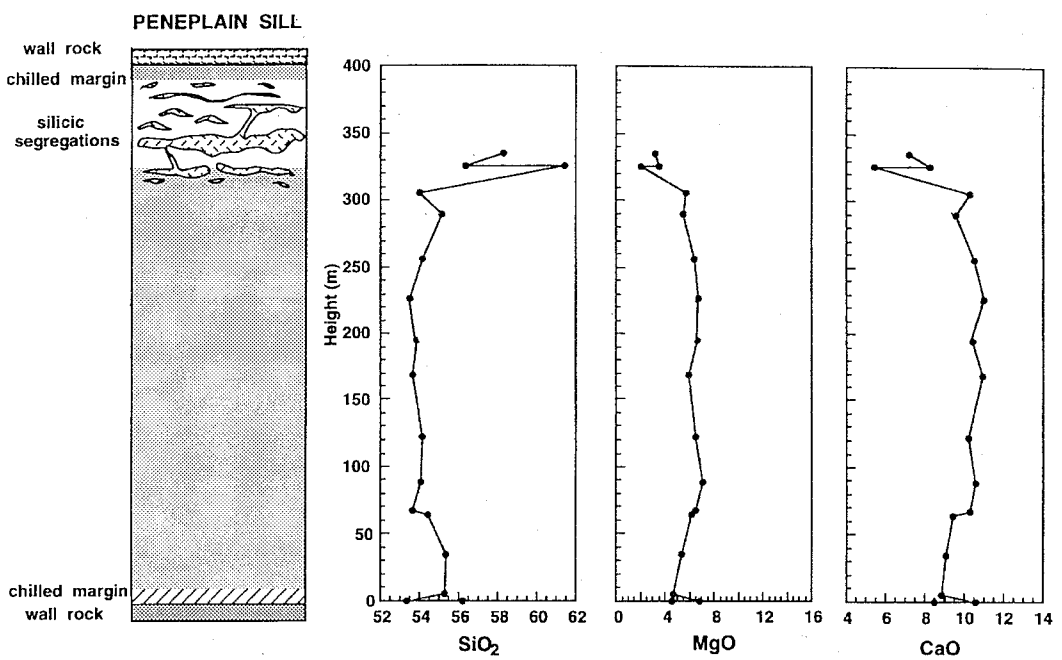

FIG. 5. The stratigraphy and bulk composition of the Peneplain sill, Antarctica (after Wheelock and Marsh, 1994).

(1964), and Komar (1972*a,b*; 1976) has considered the mechanics of phenocryst migration in ascending pipe flows. This process is shown schematically by Fig. 6, where the formation of D-shaped compositional profiles is evident. It is essential to realize that the most primitive bulk compositions in flow-differentiated sills are not at the margins, but, quite the reverse, near the middle; the most evolved compositions are at the chilled margins (Richardson, 1979; Henderson and Gibb, 1987). The chilled margins are also remarkably uniform in composition (Smith *et al.*, 1975), which is consistent with the notion that the leading magma has been largely stripped of crystals and has every chance to approach a uniform composition (see Fig. 6).

Petrologically, flow differentiation appears as concentrations of unusually large crystals (phenocrysts) in the central regions of dykes, or as tongues in sills. In the Ferrar dolerites of Antarctica, for example, Gunn (1966) found the 270 m thick Basement sill to contain a laterally extensive, 30 m thick tongue of large (2–6 mm) orthopyroxene phenocrysts near the midplane of the body. The lack of any orthopyroxene phenocrysts above or below the tongue or in the chilled margins made Gunn recognize this tongue as due to flow differentiation within a rising column of magma. More recent work on this sill (Wheelock and Marsh, 1994) shows that in the region of Bull Pass, some 30 km north of Gunn's location, this tongue swells to

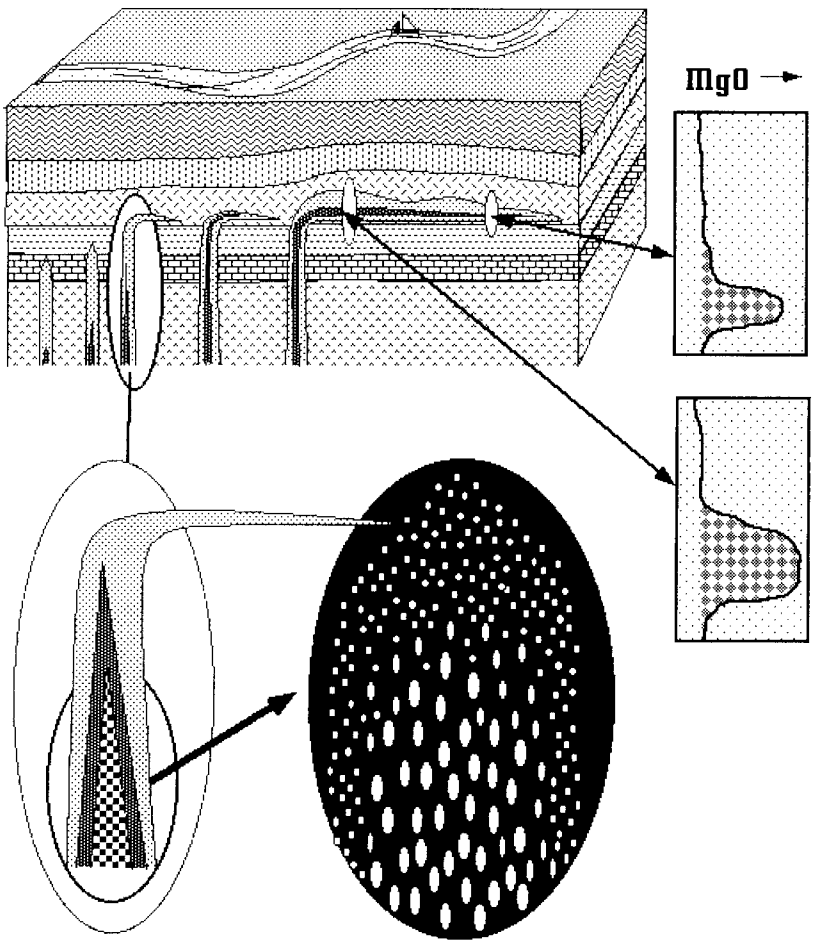

FIG. 6. Sill emplacement by a flow differentiated magma containing phenocrysts that are sorted by size and density during transport. The resulting sill contains a tongue-like mass of phenocrysts which introduce D-shaped compositional and modal profiles within the sill.

fill nearly 75% of the sill, reaching a bulk composition of 24% MgO whereas the margins contain ~7% MgO. A broadly similar tongue of orthopyroxene phenocrysts in the diabase sheet of the Culpeper Basin of Virginia suggests a similar emplacement scenario to Ragland and Arthur (1985). In a sill emplaced with a significant dip, such as the York Haven sill of Pennsylvania (Mangan *et al.*, 1993), the phenocrysts pile up, filling the sill at its lower end, which was evidently the end from which it was supplied.

The position of the tongue in the sill gives an indication of the relative time of arrival of the phenocryst-rich magma during the sill emplacement sequence. The closer the tongue to the midplane of the sill, the later the arrival of the tongue in the filling event. The fact that tongues are found at all horizons in the lower half of sills and even sometimes fill one end of dipping sills, suggests phenocryst arrival is very much dependent on proximity to the phenocryst source and the three-dimensional nature of the filling event itself.

3.4 *Comparisons with models.* The data from the Peneplain sill can be compared with the theoretical model of sill solidification, using the simplified Di-Ab-An system, proposed by Worster *et al.* (1990; 1993). They adopt the converse point of view that rapid thermal convection in the interior of the sill sustains significant cooling and crystallization *throughout* the interior. They assume that these newly-grown crystals always settle to the floor, regardless of size, and thus strong differentiation is produced essentially regardless of the overall style of cooling. The resulting vertical profile of chemical composition is roughly S-shaped, which reflects the overriding dependence of this model on the single assumption of efficient crystal settling. This assumption of efficient crystal settling regardless of size is tantamount to assuming unintentionally that the magma was laden with phenocrysts, which is the singular prerequisite for chemical differentiation. It is very much doubtful that newly-grown crystals cause such extreme differentiation.

A broadly similar model but coupling true whole rock phase equilibria to cooling and crystal settling has been formulated by the late M. Y. Frenkel and associates (1988; 1989). Crystal settling is included by ascribing a Stokesian settling velocity to each precipitating phase, which competes with capture by the upper solidification front. The kinetics of crystal nucleation and growth are, in effect, wrapped up in the ascribed settling velocities, which decouples crystallization from the dynamics of the front itself. S-shaped chemical profiles are produced that closely match those of actual sills. Here, again, the key feature of this model is the ability to make the newly-formed crystals behave as phenocrysts with settling rates large enough to ensure escape from the capture front.

A model incorporating the kinetics of crystal nucleation and growth throughout the interior of a well mixed sheet-like body has been developed by Hort *et al.* (1993). Distinctive cumulate layering, in both crystal size (generally coarsening upward) and abundance, is produced through the effect of oscillatory nucleation about a binary eutectic. Crystal capture is precluded by the absence of an upper solidification front and by the assumption of uniform crystallization throughout the interior. Layering is found possible only for bodies thicker than about 100 m, which is a limit expected to increase when the body is not always well mixed and crystallization is confined to solidification fronts. The key ingredient to promoting layering in this model is the spatial decoupling of crystallization from any control by the solidification fronts.

In summary, the vertical compositional profiles seen in sills result from the development and interaction of solidification fronts with phenocryst-bearing and phenocryst-free magma. S-shaped profiles reflect sill formation from phenocrysts evenly distributed, more or less, throughout the initial magma. D-shaped profiles reflect sill formation from phenocryst-rich, flow-differentiated magma where during ascent the phenocrysts have been concentrated in the axial region of the supply conduit and have settled *en masse* behind the leading edge of the ascending magma. I-shaped profiles result from injection of phenocryst-free magma where crystal growth occurs predominantly within the solidification front, inhibiting differentiation. Sills formed by multiple emplacements may show almost any combination of these end-member compositional profiles. The Shiant sill, for example, shows a pair of internal contacts representing perhaps an I+D profile (Gibb and Henderson, 1989), and Gibb and Henderson (1992) show similar relations in a number of other sills. They suggest that such evidence shows a clear lack of strong convection in these sills, which would tend to homogenize the inputs.

Model calculations verify, if only inadvertently, that strong modal variations, especially S-shaped profiles, are possible only when crystallization is given a distinct mechanical advantage over solidification front advancement. This is possible through the assumptions of rapid crystal settling or through extensive crystallization deep within the magma interior, well beyond the kinetic and spatial controls of the solidification fronts. In actual magmas this clearly points to the central role of phenocrysts and solidification fronts in controlling modal variations. These controls on differentiation are also well exemplified in the eruptive characteristics of Kilauea volcano, Hawaii.

3.5 *Kilauea differentiation.* A remarkable feature of magmatism at Kilauea is the near-uniform composition of the most differentiated magma erupted from the summit; this basalt has a silica content of ~51–52 wt.% and ~7 wt.% MgO (Powers, 1955; Murata and Richter, 1966; Wright and Fiske, 1971). More primitive compositions are not uncommon; they are characterized by abundant olivine phenocrysts, which, when allowed to settle from the magma, repeatedly yield this characteristic composition. In a later section, we will see that these olivine phenocrysts are, by and large, 'tramp' or detrital crystals entrained from an underlying mush column. Suffice it here to show that this special summit composition reflects the attainment, upon loss of olivine phenocrysts, of multiple saturation at the leading edge of the solidification front. That is, this special composition marks the leading edge of the solidification front and is representative of the bulk composition of the entire upper front, whereas all more evolved compositions are, in essence, trapped within the solidification front and are thus inaccessible to eruption (See Fig. 7). This is directly observed, for example, in Kilauea Iki lava lake (Helz, 1986; Marsh *et al.*, 1991). Helz (1989) also makes a case for diapiric loss of interstitial melt from the lower front, which adds melt to the upper front. Magma extracted for eruption from such bodies will be from the least viscous, hottest, and central portion of the system, which is beyond any first order influences of the solidification fronts. Protracted differentiation is thus stifled by the development of solidification fronts. The extreme paucity of siliceous lavas on the island of Hawaii (Peterson and Moore, 1987) reflects the pervasiveness of this fundamental control on differentiation.

Small, isolated lenses or segregations of slightly more siliceous melts are also found in drilling the crusts of Makaopuhi and Kilauea Iki lava lakes (Wright and Okamura, 1977; Helz, 1980), and similar features are common to the upper portions of many basaltic sills, as we have seen already for the

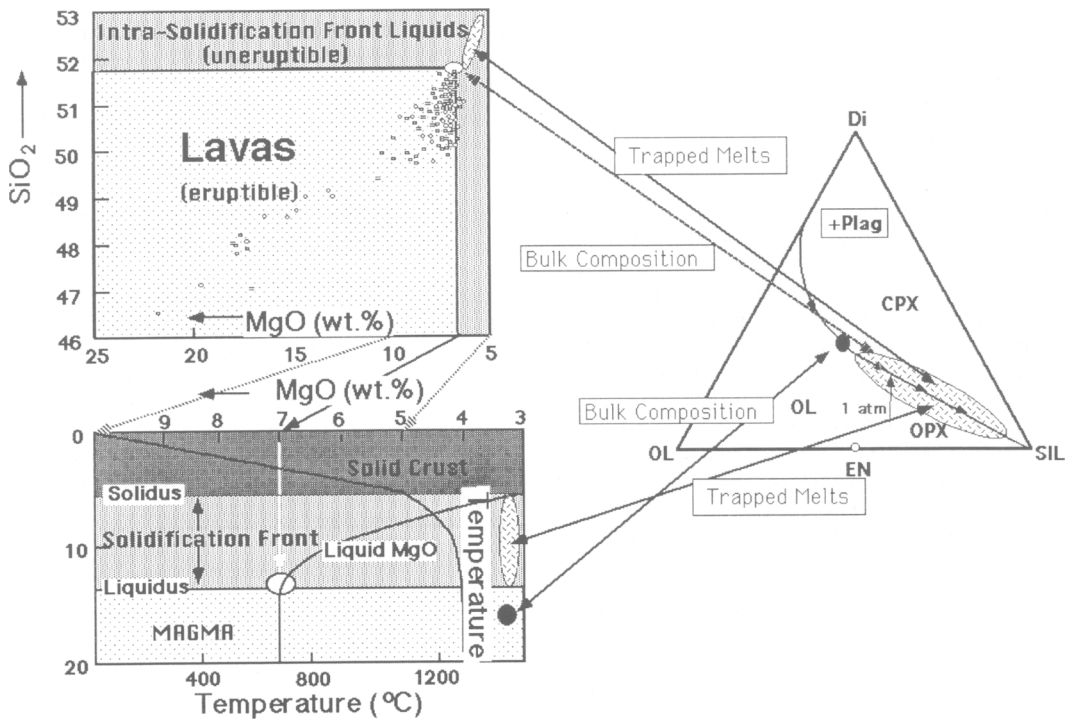

FIG. 7. Solidification front control on the composition of lavas erupted from Kilauea volcano, Hawaii. The diagram at the upper left shows the observed distribution of lava composition observed from the summit of Kilauea. The solidification front on the lower left shows the variation in magnesia concentration in the interstitial melt of the front; although this melt becomes strongly differentiated, it is inaccessible to extraction and eruption. The normative ternary system on the right reflects these compositions, but does not reveal why the more fractionated compositions are not observed as lavas.

Peneplain Sill (Fig. 5). The formation of these features and the possible accumulation of them into a significant mass of eruptible magma is critical to understand as it represents a petrologic process whereby trapped melt occurring *within* the solidification front can be extracted from the front. We next consider this process.

4 Solidification front instability

The pegmatoids do not simply represent the residual liquid as they occur at many horizons. They can only have accumulated by liquid percolating upwards through the crystal mush at a late stage when the whole sill was some 80 percent crystalline and being brought to rest by impermeable rock at successively lower levels.

B. M. Gunn, 1966, in discussing Antarctic sills.

For almost 150 years petrologists have recognized the intimate connection between igneous bimodality and igneous diversity. Bunsen (1851) first saw that combinations of basic and silicic end-member compositions can yield reasonable intermediate compositions. Bowen (1915*a*) saw the great chemical versatility of separating solids from liquids, but he also saw in every clear-cut field example the great limitations of fractional crystallization by crystal settling alone. He thus reintroduced Harker's idea of filter pressing under a regional stress field to gain an extreme siliceous component (Bowen, 1919); but the time scale for regional squeezing relative to solidification is much too long for this to be effective. Nevertheless, rhyolitic magmas do exist, and regardless of the possibility of refinement through partial melting, some hitherto unappreciated process must exist to extract refined melts from solidification fronts.

Fractional crystallization and differentiation by crystal settling, indeed, commonly occurs; it is often highly effective, especially in singly saturated systems laced with phenocrysts, but it has been repeatedly found not to produce rhyolite from basalt in the normal sense of the process. As far as anyone knows, the Moon has no significant volume of highly siliceous lavas; neither does Hawaii or ocean ridges, although at least the latter two do produce bimodal differentiates as discrete, isolated, lens-like bodies (e.g. granophyres, plagiogranites). The process by which these lens-like bodies form may be central to understanding Earth's extreme differentiation (e.g. Morse, 1987).

Along the roof of certain sheet-like basaltic intrusions, the solidification front can thicken to the point of becoming gravitationally unstable, whereupon the crystalline mesh or network begins to sag and tear or delaminate internally. As gashes and lenses open, nearby late-stage interstitial melt is drawn in to fill the potential voids (see Fig. 8). This is solidification front instability (SFI). Large (2 m × 50 m), horizontally interdigitating lenses form a siliceous or granophyric horizon in, for example, the Ferrar dolerite sills of Antarctica (Fig. 5) (Gunn, 1966; Wheelock and Marsh, 1993). Depending on their exact composition, the melt of these lenses can be less dense and much more viscous than the host magma, making them highly resistant to back-assimilation through mixing, and they can thus survive and accumulate during wholesale collapse along the roof and walls of the solidification front. These segregations may also accumulate and grow into volumetrically significant bodies through repeated cycles of melting and solidification either during major episodes of reinjection or during long term tectonic recycling.

In this fashion a fundamentally bimodal silicic chemical signature, or perhaps silicic chaos, is irreversibly entered into basaltic systems and Earth itself. Because the steady, long-term trend toward increasing silica in the residual liquid, due to crystal separation, suddenly bifurcates (due to SFI) to produce simultaneously both strongly siliceous and more mafic compositions, the process is intrinsically chaotic (e.g. Pippard, 1985). The process is not random, but depends on a specific set of prescribed conditions initiating SFI. The more mafic component is made by fresh magma drawn into the solidification front, stimulating continued growth of mafic crystals in the leading region of the front, as shown by Fig. 8. SFI is, in effect, bimodal fractionation, and largely accomplishes what Bowen wanted filter pressing to do. Although a detailed treatment of the mechanics of SFI is beyond the scope of the present work (but see Marsh, 1995), some key observations are important to note.

There are at least two critical petrologic observations establishing that the silicic segregations seen in sills and lava lakes are derived locally from the SFI process and are not blobs of interstitial melt leaking upward from the lower solidification front. The first is that the grain sizes within segregations fine strongly downward. Within any segregation the topmost crystals are unusually coarse and the lowermost crystals are small; upper contacts are sharp and fracture-like and the lower contacts are diffuse and indistinct (Gunn, 1966; Wheelock and Marsh, 1993). The second is that the segregations form laterally extensive horizons of lenses that systematically thicken downward until near the centre of the body where they thin and disappear (Helz, 1980).

Instead of producing siliceous liquids in the classical sense by extensive and protracted near-liquidus crystal fractionation, solidification front instability operates well behind the liquidus and

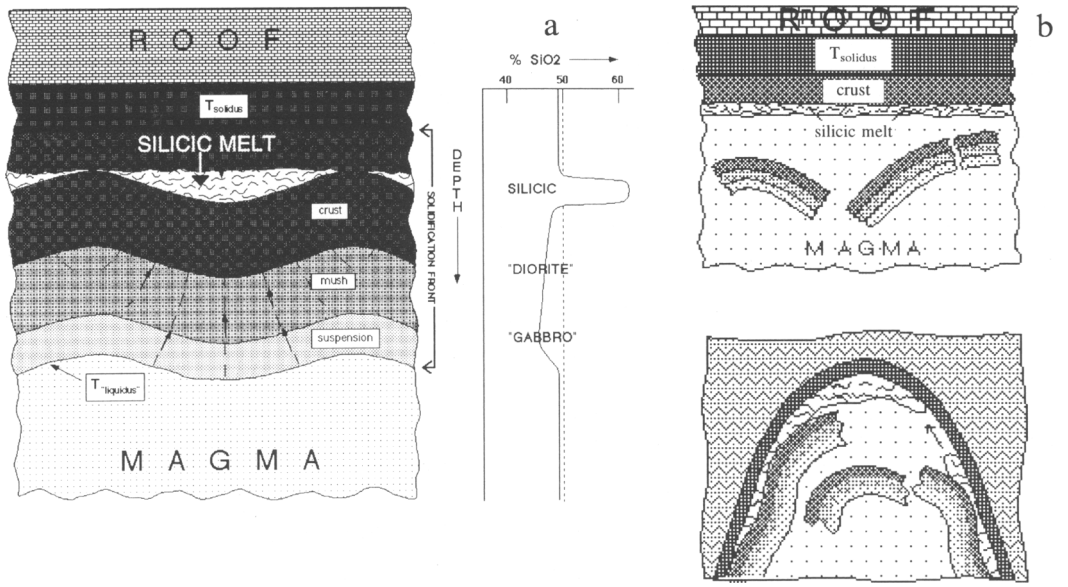

FIG. 8(*a*). Solidification front instability (SFI) whereby siliceous interstitial melt is segregated into lenses through the sagging or internal tearing of the upper solidification front under the influence of gravity. The ensuing silicic segregations are chemically and spatially separated from the underlying magma, and new melt entering the leading edge of the front makes this region relatively more primitive. (*b*). Later stages in the solidification front instability where the front may wholly detach from the roof, freeing the still molten silicic segregations to collect along the roof (upper). When this process occurs along a sloping roof, the silicic melt may migrate upwards and form a separate magma.

thus the separated melts are highly fractionated both spatially (within the chamber) and chemically relative to the host basalt. This overall process, which may occur either during cooling (i.e. retrograde) or reheating (i.e. prograde), may be of central importance in understanding the fundamentals of igneous processes. Here are some possible examples.

1. *Smaller scale process.* The solidification front instability (SFI) is repeatedly arrested due to the thinness of the magmatic body. The solidification front apparently never becomes thick enough to become strongly unstable, or when it does the close proximity of the floor solidification front supports the sagging roof, truncating the instability (see Fig. 8). Granophyric segregations and pegmatoid lenses in thick diabase sills, Hawaiian lava lakes, and the Icelandic crust and oceanic crust (i.e. plagiogranite) may all result from this style of SFI. The idea of generating Icelandic rhyolites through melting of basaltic crust containing granophyre (Sigurdsson, 1977; Oskarsson *et al.*, 1982) needs a process such as SFI to produce the original granophyres (Marsh *et al.*, 1991).

2. *Larger scale process.* In the limit where SFI progresses much further, leading to wholesale detachment of the solidification front itself, the silicic liquids produced during the incipient stages of SFI may collect and locally form relatively large masses of magma. Very large magmatic sheets may thus have repeatedly lost their upper solidification fronts and gained silicic magma near their upper contacts. Large 'stoped' blocks are not uncommon in some large layered plutons (e.g. McBirney and Noyes, 1979; Parsons and Becker, 1987). Many large magmatic bodies, like Dufek (Ford, 1970), Muskox, and Bushveld, have layers of granophyre along their roofs. Although isotopic studies show that these may also be related to melting of roof country rock, the loss of the upper solidification front due to SFI contributes both additional silicic material and fosters the thermal conditions necessary for melting. The small sizes of silicic segregations allows isotopic equilibration through interpenetration and diffusion. If so, it is again curious, however, why the textures of granophyres produced in roof melting are not more like the granitic plutonic rocks from whence they came.

Finally, in chambers with outward-sloping walls the silicic melt, now freed of sluggish porous-media flow, may migrate relatively unencumbered up the walls and collect at the roof; having the effect envisaged independently by Ernst (1960) and McBirney (1980) (see Fig. 8*b*). There are also *prograde SFI* processes.

Progressive heating of a terrain containing silicic lenses allows simultaneous wholesale melting of the silicic rock and only partial melting of the host rock. Structural integrity is lost and inward collapse of the host rock occurs, allowing the now exposed silicic bodies to collect at high points in the system. The initial formation of the proto-continental crust on the young Earth, whether or not a magma ocean was involved, may well have been facilitated by SFI processes. In this regard, there is clear evidence that some sills do carry significant concentrations of granophyric blobs, having been generated upon emplacement in an earlier SFI event.

4.1 *Bulbous granophyres near the upper contact.* A surprising number of basaltic sills contain large, isolated masses of granophyre near the upper contact. The Culpeper Basin granophyres (Froelich and Gottfried, 1988), the granophyres of the Gettysburg region of Pennsylvania (Grossenbacher, 1994), and the Tasmanian granophyres of The Red Hill dolerite (McDougall, 1962) and the Great Lake dolerite sheet (McDougall, 1964) are all quite similar in form and petrology. They occupy high points in the roof of the host sill; they are always separated from the roof rock by a thin ($\sim$10 m) band of dolerite; so far as is known, they are in isotopic equilibrium with the host rock; and the host rock is essentially free of other systematic granophyric segregations.

A likely scenario for the formation of these masses is shown by Fig. 9, which is after that of Grossenbacher and Marsh (1991) and Grossenbacher (1994). Discrete blobs of granophyre from a previous SFI event are remobilized and rise, due to their low relative density, to the head region of the ascending column of magma. Upon emplacement these blobs collect against the roof near the entry point and undergo roofward compaction, efficiently evicting the intervening host melt in a matter of weeks to form a single layer of granophyric melt that flows uphill, along the roof and collects at a high point (Grossenbacher, 1994). The great buoyancy of this mass, possibly enhanced by the similar accumulation of gases exsolved during ascension, further deforms the roof rock. The ubiquitous, thin margin of host diabase along the upper contact gives a measure of the time after emplacement for positioning of the granophyre, and calculations of compaction rate and along roof flow rate show the overall process to be remarkably efficient (Grossenbacher, 1994). The final state of the sill is such that upon burial deep in the continental crust the granophyric mass may in the solid state add itself diapirically to the crust much as in the motion of gneiss domes.

Because silicic blobs should easily migrate to the head region of a propagating dyke, field occurrences should be unusual. Nevertheless, Vogel and Wilband (1978) describe and review a host of occurrences of coexisting acidic and basic melts. A similar process involving transport and assimilation of silicic xenoliths has been postulated by White *et al.* (1989) for the emplacement of the Vandfaldsdalen macro-dyke of the Skaergaard region of East Greenland.

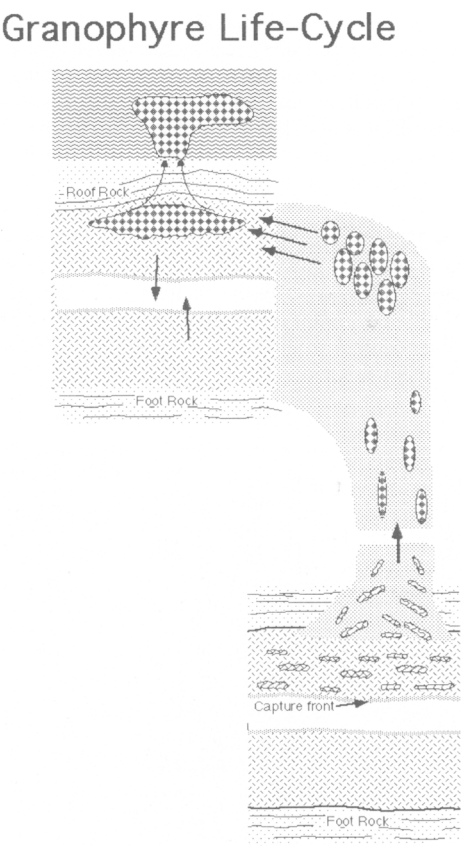

FIG. 9. The possible life cycle of granophyric segregations produced originally by SFI. Wholesale remelting of the sill mobilizes and allows collection and compaction of the blobs against the roof of the new sill; subsequent flow to a high point along the roof may further dome the roof. Deeper burial of the sill in the crust may allow slow diapiric motion, in the solid state, to add the silicic rock to the continental crust.

5 Sills as magma chambers

Basic intrusions that reach thicknesses of more than four or five hundred meters tend to be notably more coarse-grained than thinner dykes and sills, and almost all have some form of layering. It seems that they crystallize in a way that is fundamentally different from thinner bodies. A. R. McBirney, 1984

The principal reason for this long-recognized dichotomy between basaltic sills and much larger intrusions may simply be due to scale. Except for the sure influence of sloping sidewalls in some cases, it is certain that each of the processes so far discussed is not simply specific to sills, but is general and must occur repeatedly and competitively in all igneous bodies regardless of size. The mere size of a body brings with it a set of allowable processes. Small bodies, for example, may cool too fast ever to allow SFI, whereas it is almost certain that bodies beyond a certain volume must be produced by repeated injections of magma, which sometimes contain large amounts of phenocrysts; and SFI may occur to the fullest extent, continually disrupting the upper solidification front such that in the end almost all direct traces of it are lost.

It is difficult not to interpret, for example, the general petrologic features of the enormous (perhaps 500,000 km^3) Dufek intrusion of Antarctica (e.g. Ford, 1970) as essentially similar to those of the adjacent, comagmatic Ferrar dolerite sills, except on a larger scale and the result of many injections. The orthopyroxene layering is the same as the orthopyroxene tongue of the Basement sill except it has been better sorted due to being injected and dispersed into a larger chamber. The anorthosite horizons are broadly similar to anorthosite lenses in the orthopyroxene sill tongues, except more massive and better segregated. Although from a different perspective, this is apparently largely also the sentiment of Ford himself (e.g. Ford and Himmelberg, 1991); and it is difficult not to see the great similarity between Dufek and Stillwater. Detailed isotopic studies of Stillwater cumulates suggest, for example, an involvement of geochemically distinct magmas throughout the history of the body (Lambert *et al.*, 1994). Textural studies of the Kap Edvard Holm layered gabbro complex in eastern Greenland reveal a series of quench zones reflecting sequential injections along the chamber floor (Tegner *et al.*, 1993). Texture maturation in Mush zones efficiently erases all but the most severe evidence of sequential injections; and the absence of internal contacts, especially in large bodies, is no evidence against reinjection. Aside merely from emplacements, there are of course also a host of attendant processes operating to influence the final igneous texture and composition while it is in a porous and permeable state (Lesher and Walker, 1988; McBirney, 1995).

The main lesson here is that these are general processes whose effects are readily recognizable. Misidentifying the occurrence, origin, and role of, for example, phenocrysts in the incoming magma will, *per force*, lead one to construe this rock record as due, for example, to unusual changes in phase equilibria or to crystallization spatially decoupled from solidification fronts. The location and nature of the silicic segregations may wrongly lead one to interpret them as representing a so-called sandwich horizon containing the final, residual liquid. And not realizing that basaltic magma cannot easily, in and of itself, fractionate *in situ* into a core region of eruptible rhyolite, will cloud the wonder of why magmatic systems as vast as Hawaii do not produce rhyolite lava.

The problem of recognizing the important roles of phenocrysts and solidification fronts in magma have plagued petrologists from the very first attempt to explain the diversity of igneous rocks. The fundamental beliefs of petrologists concerning the crystallization of chambered magma are rooted in age old interpretations that carry through to this day. It is essential to appreciate the ideas and facts upon which rest the origins of many of our fundamental magmatic principles and beliefs.

6 Some historical perspectives

6.1 *Beginnings.* As soon as it became clear that chemical diffusion in silicate liquids was very slow (e.g. Becker, 1897*a*), hope faded that the previously popular Ludwig-Soret effect could separate magma into distinct compositional fractions resembling the common igneous rocks. This was the theory of differentiation that was replaced in 1897 by

"the simple principle of fractional crystallization, which is the very opposite of magmatic differentiation."

Becker (1897*b*) further put together two fundamental observations from the freezing of alcoholic liquids and the fact that magmas may be eutectic liquids.

"A bottle of wine or a barrel of cider exposed to low temperature deposits nearly pure ice on the walls, while a stronger liquor may be tapped from the centre. If still a lower temperature were applied the central and more fusible portion would also solidify. Such a mass would be, so far as I can see, a very perfect analogue to a laccolith."

In his melting studies of metal alloys Frederick Guthrie (1884) reinstituted Aristotle's term of *eutexia* to describe eutectic phase equilibria. Guthrie also concluded that:

"I submit that, according to analogy, we should regard compound rocks and minerals [i.e. multi-component systems], other than sedimentary rocks, as representing various kinds of eutectic alloys."

Becker saw that to grow crystals on the walls and at the same time concentrate a homogeneous residual eutectic liquid by diffusion over any significant distance is ineffective and that:

"The fractional crystallization process depends essentially upon convection currents.....The only function of diffusion in this case would be to preserve homogeneity of the residual or mother liquor, so that the eutectic state could be any sensible part of the fluid until the whole mother liquor was reduced to this condition."

A central tenet of this idea is that convection freely circulates magma in and about an interconnected forest of crystals growing inward on the walls. Ice crystals in alcoholic solutions and dendrites in metal alloys solidify very much in this fashion (e.g. Kurz and Fisher, 1992; Davis *et al.*, 1992). It is an exceedingly efficient form of fractional crystallization, and it is this model that, perhaps unknowingly, has been the core principle in the development of models of magma chambers for the past 100 years.

L. V. Pirsson (1905), although initially intrigued by Soret differentiation, adapted Becker's model of convection and fractional crystallization to explain the Shonkin Sag laccolith in north-central Montana. He added to it the enduring concept of crystallization from the floor up with the residual melt collecting near the roof in what later became known as a sandwich horizon. The details of this model are worth knowing (Pirsson, 1905; p. 187–8).

It seems almost impossible to resist the view that in an enclosed mass of magma sufficiently mobile for local differentiation to take place convection currents due to unequal cooling would occur. On the upper surface and along the outer walls cooling would take place more rapidly; on the floor of the chamber, protected by the heated mass above and with heated rocks below, less rapidly. Thus there would be a tendency along the top and sides for the magma to grow heavier and to descend. Material from the more highly heated central part would tend to rise and replace this, and thus currents would be established in the magma, rising in the centre, flowing off to the sides at the top, and descending along the cooler walls. ..... Such currents, once established, would continue as long as sufficient mobility remained in the magma to permit them.

At some period crystallization would take place, and this most naturally would begin at the outer walls. It would not begin at the top because the material would arrive there from below at its highest temperature. Moving off toward the sides the material begins to cool and descend and becomes coolest as it nears the floor; here crystallization would commence. The first substance to crystallize is the solvent, which in this case would be the femic minerals, chiefly augite. Part of the material solidified would remain attached to the outer wall and form a gradually increasing crust, and part would be in the form of free crystals swimming in the liquid and carried on in the current. Probably at first, as the liquid moved inward over the floor of the laccolith and became reheated, these crystals would remelt, giving rise to numerous small spots of magma of a different composition, which would slowly diffuse. As time went on, however, there would be a constantly increasing tendency for the crystals to endure; they would be carried greater and greater distances. But as they are solid objects and of greater specific gravity than the liquid, there might be a tendency for the crystals to drag behind and accumulate on the floor of the chamber. Moreover, from the heat set free at the time of their crystallization and from the resulting concentration of the chemically combined water vapor in the magma, the residual liquid would tend to have its mobility kept undiminished, since these would be factors which would tend to counteract the increase in viscosity due to cooling. In this manner it may be possible to understand how there would form a femic marginal crust and a great thickness of the femic material at the bottom of the laccolith. As the cooling went on the edges of the outer crust would rise more and more toward the top, finally spreading over it, and as a result the crust should be thinner on the top than elsewhere, as in the Shonkin Sag laccolith, in which the upper crust of femic rock is still preserved.

That intruding magmas, like that of Shonkin Sag, often carry large concentrations of phenocrysts was not recognized at this time. Thick cumulate piles of these phenocrysts motivated the idea of extensive crystallization from the bottom up.

In his magnum opus of 1915 Bowen (1915*a*) reiterated Pirsson's model, but found fault with the roles of convection and diffusion, which Pirsson actually did not favor, in supplying nutrients to crystals growing on the border (p. 12).

"The result should be the growth inward from the contact of huge crystals of the minerals of early separation.

Instead he argued that discrete, unattached crystals will be the rule and these need not remain near the border, but are free to grow and settle. Apparently he did not realize that Pirsson's model already contained this feature. His presentation of the albite-diopside-anorthite ternary system convinced him that protracted crystal settling in basalt cooling along a cotectic phase boundary in large, slowly cooled bodies could produce the full diversity of igneous rocks (p. 90).

The decision is reached that this differentiation is controlled entirely by crystallization. The sinking of crystals and the squeezing out of residual liquid are considered the all-important instruments of differentiation, and experimental evidence is adduced to show that under the action of these processes typical igneous rock series would be formed from basaltic magma if it crystallized (cooled) slowly enough.

Instead of a steady displacement inward of the residual liquid, crystals settle to the floor and displace the melt upward. Slow cooling is necessary to allow the crystals to grow large enough to escape and settle to the floor. Fast cooling of basalt gives fine-grained diabase with the only granitic differentiate being the local residual, interstial melt. The common finding of silicic differentiates at the top of diabase sills convinced Bowen that crystallization was dominantly from the bottom up. He himself when a student had mapped such a diabase sill in the Gowganda Lake District (Bowen, 1910) and had interpreted the silicic pod at the upper contact as some involvement with the overlying sediment. Now (1915*a*) he reversed his earlier finding and read the rocks as clear evidence of extreme differentiation due to protracted crystal settling.

In sum, to maximize fractional crystallization along a cotectic Bowen needed extensive crystal settling, and thus the crystals could not be attached to a solidification front. And to deposit the last liquid at the upper contact, crystallization had to be from the bottom up. Both of these features, for broadly similar reasons, are features of Pirsson's model.

That the marginal zone of crystallization is a dynamic entity, thickening and propagating inward and consisting of an outward increasing concentration of discrete crystals causing an enormous variation in viscosity, was cleanly introduced by Grout (1918). By approximating convection as flow through a pipe in the shape of a large, chamber-filling cell driven by thermal and compositional buoyancy, Grout showed the great effectiveness of convection over single crystal Stokes settling in depositing wall- and roof-grown crystals on the floor. He also noted that the density differences driving convection down a vertical wall increase outward with increasing crystallinity but so does viscosity, which resists the flow.

The active force [of convection] *is much greater than the resisting viscosity up to the point where crystals begin to touch each other.*

Crystals carried along and deposited by down-wall flows may also contribute to banded and layered rock, reasoned Grout, and convection may even be rhythmic in accord with volcanic events or in response to variations in cooling.

6.2 *The Becker-Pirsson-Grout magma.* This basic model of crystallization and convection has survived to this day essentially intact from Becker's (1897), Pirsson's (1905), and Grout's (1918) original descriptions; some important modifications will be mentioned presently. It is therefore important to note the central features of the model and the facts on which these features are based.

1. *Convection.* Becker correctly realized that uneven cooling along level (i.e. equipotential) surfaces and along walls will promote convection. This is sound reasoning. But the convection that most early studies want and employ is really that associated with crystal sedimentation, not necessarily thermal convection.

2. *Floor up crystallization.* Although Becker originally imagined crystals growing inward from nutrients supplied by convection, Pirsson was motivated by the unusually thick cumulate pile on the floor of Shonkin Sag laccolith to propose strong roofward cooling and crystallization but crystal transport by convection to the floor. The syenite residual melt ends up disproportionately near the roof and the upper border series of shonkinite is much thinner than the thick shonkinite pile on the floor. The fact that the thick pile of basal cumulates is due to massive settling of 35 vol.% phenocrysts already in the magma upon emplacement apparently never occurred to Pirrson (Hurlbut, 1939). The striking mineralogical variations in Shonkin Sag are simply due to *in situ* settling of these original phenocrysts (Congdon, 1990; Marsh *et al.*, 1991).

Bowen (1915*a*) wanted crystallization from the floor up in order to place the silicic differentiates in diabase sills near or at the upper contact. As we have seen, these silicic segregations do not represent the last liquid or sandwich horizon, but instead form in response to internal tearing of the upper solidification front; and some such segregations may be injected with the initial magma. These sills are chemically bimodal in this respect and a series of intermediate compositions is not generally found. Moreover, laterally adjacent diabase shows a good chilled margin stemming from the initial emplacement.

6.3 *The Skaergaard influence.* The discovery of the Skaergaard rocks again brought forward the features of Shonkin Sag laccolith, namely, an unusually thick basal pile of cumulates and a relatively thin upper border series. After careful

consideration, Wager and Deer (1939) adopted the Becker-Pirrson-Grout model of roof and wall crystallization with deposition by convection, sometimes rhythmic, to explain the whole body right down to the remarkable trough bands. They needed to explain a wide spectrum of layer scale and character, and to do this they needed three things: crystals, crystal settling, and crystal sorting.

The presence of crystals is a given fact in any plutonic problem, even though their ultimate origin (i.e. initial phenocrysts or not) is often not clear. Although there was some general hesitancy among petrologists to adopt pervasive crystal settling, Bowen's (1915*b*) experimental demonstration of the sinking of olivine deeply impressed Wager and Deer. The most obvious process of crystal sorting is through sedimentation and sedimentary currents, which is clearly a form of convection. Sorting in sedimentation or during magma ascent is convection, it is depositional or two-phase convection (Grout, 1918). Convection itself has always been a petrologic process, but its occurrence and understanding has been often needlessly interwoven with thermal and compositional convection. Once the seminal thermal problem was laid out by Rayleigh in 1919, the stability criterion of thermal convection became justification for employing convection, which for sedimentation is unnecessary. And so it was with Wager and Deer; they invoked crystallization, convection, and crystal sorting in a tank of magma injected all at once and containing some "primocrysts" whose subsequent importance was unclear. Nothing better could be asked for.

The concept of magma freely flowing in and amongst a dendritic-style assemblage of crystals growing outward from the walls, as suggested by Becker (1897*a*), is adopted by Wager and Deer (1939) to explain the border series, especially the perpendicular feldspar rock. These crystals grew attached to the walls and could not participate in fractionation and layering (*ibid.*, p. 281), but the crystals that produced the layering were produced unattached, in the fashion of Pirsson (1905), to be sorted in essentially gravity currents along the walls. There is no real appreciation here of the spatial characteristics of multiply-saturated silicate solidification fronts.

Hess (1960) used the same model to explain Stillwater, but he also brought back diffusion in the liquid state through a thin ($\sim$2 m) layer of mush on the floor; and G.B. Hess (1972) presented a modern analysis of the implications of convection and crystal settling in the large Stillwater system. Irvine (1970) gave the first internally consistent conceptualization of this basic model based on roofward heat transfer, phase equilibria in a circulating magma, and floor-up crystallization quantitatively-linked to the rate of roofward heat loss.

The original field interpretations upon which the Becker-Pirrson-Grout model was based were clearly flawed, and we might expect the ensuing model itself to be flawed. But later observations from large systems, like Bushveld and Muskox, provide convincing evidence that crystal accumulation is clearly from the floor up and that there is little or no indication of any significant upper solidification front or border series. Interpretation of the type bodies may have been wrong (e.g. Shonkin Sag), but the observations are correct. It is the magmatic conditions that produce these relations that may yet be open to question.

6.4 *Filling sequence and phenocrysts.* Two seemingly obvious and elementary features missing from essentially all the above considerations are: 1) a lack of appreciation of the phenocryst content of the initial magma, and 2) the nature of the filling event or events themselves. Only Hurlbut (1939) explicitly deduced, based on the seminal innovations of Osborne and Roberts (1931), that the filling magma was crystal-laden, which is the key to understanding Shonkin Sag. The evolution of the entire body to a very great extent simply involves settling of crystals from a phenocryst-laden initial magma. Knowledge of body shape, style of filling, and initial phenocryst content is essential to understanding any magmatic body. Yet this is information generally not easy to obtain, which commonly leads to conflict over whether the final rock sequence was produced by a single injection or by multiple injections (e.g. Barksdale, 1937; Hurlbut, 1939). The choice is perhaps clearest in sills, where the course of emplacement and solidification is often clear.

In this context, through the earlier presentation, we have been able to isolate a number of fundamental processes involving solidification fronts alone, the interplay of solidification fronts and phenocrysts, and the behaviour of phenocrysts in ascending magma. The effect of these processes on the evolution of a magma chamber is shown by Fig. 10. We have, however, until now said little about the origin of phenocrysts themselves.

7 New crystals, phenocrysts, and mush columns

The sharp break in grain size between phenocrysts and groundmass is correlated with some corresponding change in conditions prevailing during freezing of the magma. Such a break occurs where slow cooling of magma deep within the crust has given way to rapid cooling following uprise of the magma and extrusion at the surface or injection into cooler rocks of the upper crust.

I.S.E. Carmichael, 1974

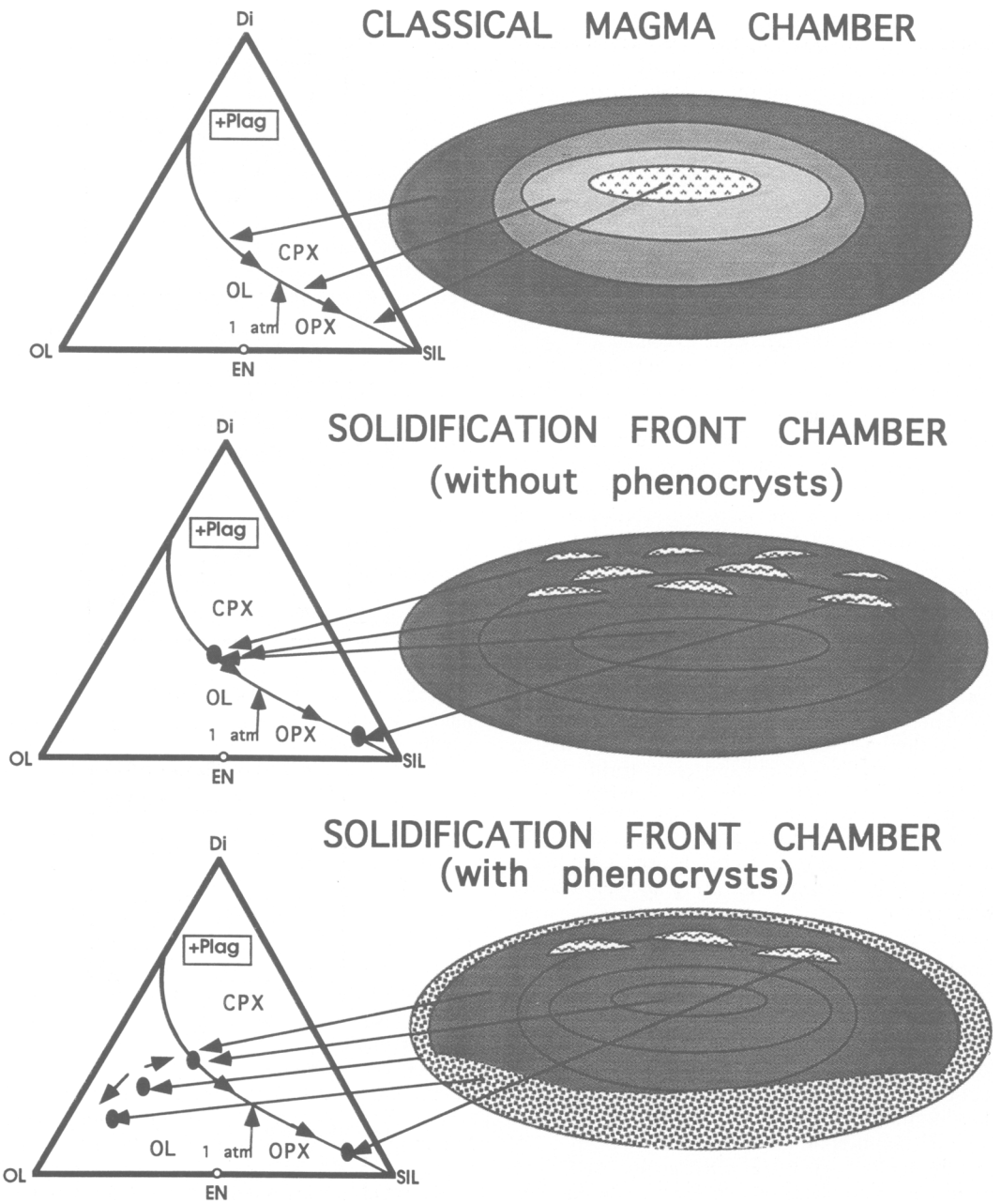

FIG. 10. Magma chamber evolution for magmas free of phenocrysts (middle) and carrying phenocrysts (lower) in comparison to the classical concept of magma chambers (upper).

7.1 *Nucleation and growth.* That magmas commonly carry crystals, large numbers and of large size, is a fairly new concept. To be sure, the description of a magma as porphyritic has been freely used for a hundred years. But the more common historical perception has been that most crystals grow after

emplacement or upon eruption. The thermal state of magma was not appreciated until the use of geothermometers and phase equilibria became widespread along with attempts to link the petrographic character of lava to magma (e.g. Carmichael *et al.*, 1970). Earlier studies most often implicitly assumed a state of significant superheat and thus an absence of any initial crystals. Hurlbut's 1939 deduction, mentioned already, of a phenocryst-laden magma forming Shonkin Sag is exceptional in this respect; as also is Gibb's (1976) deduction that phenocryst-laden ultramafic magma seen in dykes may give rise to regionally-associated layered intrusions.

Evidence that an actively crystallizing magma always contains a wide spectrum of crystal sizes, from nuclei to phenocrysts, is apparent with the measurement of crystal size distributions (CSDs) (e.g. Cashman, 1993; Marsh, 1988*a*). Post-emplacement or newly-grown crystals have characteristics intimately reflecting the most recent conditions of nucleation and growth, whereas phenocrysts arriving with the magma reflect a markedly different kinetic history. CSDs provide a quantitative connection between petrographic observations and the kinetics of crystallization, which allows an assessment of the distinguishing characteristics (e.g. number, size, and distribution) of crystal populations in magmatic systems.

Crystal size distributions measure the population density (n; in units of number of crystals per vol. of rock per crystal size, or no./L^4) of crystals as a function of crystal size (L). Although any number of physical and chemical processes can affect the CSD, many lavas, sill rocks, and plutons exhibit log-linear or near log-linear correlations of population density with crystal size. CSDs for clinopyroxene in the upper shonkinite of Shonkin Sag laccolith (Marsh *et al.*, 1991), for plagioclase in a basalt-mugearite sill on Reunion Island (Upton and Wadsworth, 1967; McCormick and Marsh, 1995), and for plagioclase in Makaopuhi lava lake (Cashman and Marsh, 1988) are shown by Fig. 11.

There are three basic features of CSDs that are useful to understanding crystal size and crystal abundance in magma. The first is simply the dependence of typical crystal sizes on the rates of nucleation and growth. The second is the total number of crystals (N_T, of any specific phase), which is a property of rocks not usually considered, but is intimately related to both nucleation rate (J) and growth rate (G). And the third is the relation of linearity in the CSD to the history of nucleation and growth.

CSD linearity. There are many dynamic crystallization models that can give rise to this style of broadly linear CSD's (e.g. Marsh, 1988*a*), but perhaps the most obvious is simply an exponential

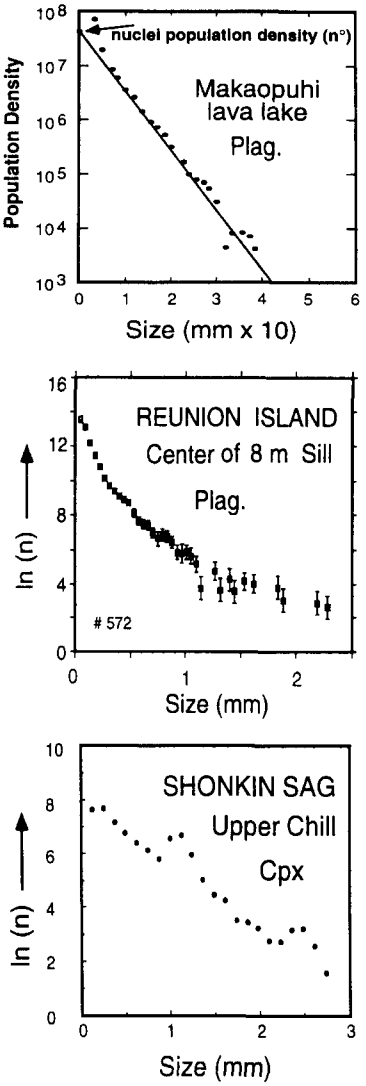

FIG. 11. Typical crystal size distributions (CSDs) for Makaopuhi lava lake (85 m), a sill on Reunion Island (8 m) and Shonkin Sag laccolith (70 m); the first two are for plagioclase and the third is for clinopyroxene. The log-linear CSD of Makaopuhi is typical for a system undergoing batch crystallization in response to a nucleation rate that increases exponentially with time. The first part of the Reunion Island CSD is also linear but the upward curvature at larger sizes suggests some accumulation of plagioclase phenocrysts from an earlier nucleation event at depth. The Shonkin Sag CSD shows a broadly linear trend but with a concentration of crystals at L~1.25, which reflects the abundance of clinopyroxene phenocrysts, from a much earlier nucleation event, carried in the initial magma.

rise in nucleation rate with cooling. Coupled with a linear growth law for crystals (e.g. L = Gt, where G is growth rate and t is time), which by all accounts is reasonable, any exponential nucleation event propagates across the CSD diagram producing a linear CSD. That is, the time variation in nucleation density ($n^0(t) = n(t,L = 0)$) establishes nuclei along the L = 0 axis and growth moves these populations horizontally across the CSD diagram (see Fig. 12). Each value of n along the L = 0 axis represents n^0 at the time of the nucleation event. The largest crystals are the oldest and represent the lowest nucleation rates (J(t)), which are formally given by $J(t) = n^0(t)G$ (in units of no. per unit vol. per sec.).

The direct relation of CSD slope to a time variation in nucleation rate is best seen by considering the governing CSD population balance equation for a batch system (e.g. Marsh, 1988*a*),

$$\left(\frac{1}{n}\right)\frac{\partial n}{\partial t} + G\frac{\partial \ln(n)}{\partial L} = 0 \qquad (1)$$

where n is based on the bulk volume. Taking G to be constant and substituting n = J/G in the first term, this becomes

$$\frac{\partial \ln(J)}{\partial t} = -\frac{G\partial \ln(n)}{\partial L} \qquad (2)$$

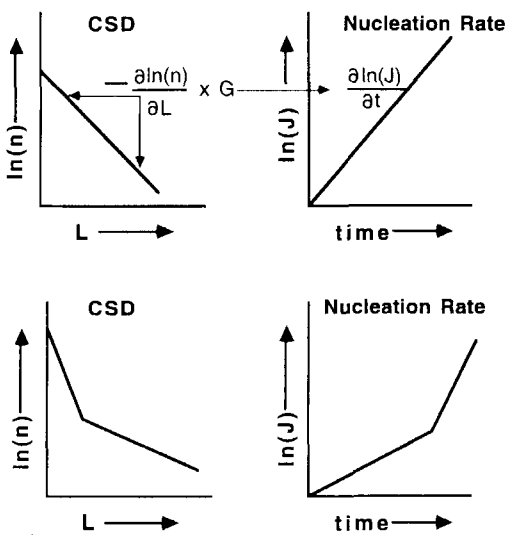

FIG. 12. The history of nucleation in a batch system is directly related to the slope of the CSD on a log-linear plot (upper pair). A CSD made up of two linear segments reflects a history of nucleation marked by two distinct cooling events from which the nucleation history can be reconstructed (lower pair).

The derivative in the last term is just the slope (= S) of the CSD, which when constant allows (2) to be integrated to give

$$J = J_0 \exp(-GSt) \qquad (3)$$

where J_0 is the initial or beginning nucleation rate. Thus a log-linear CSD with a constant growth rate represents an exponential increase in nucleation rate; and a CSD made up of more than a single straight segment, which is not uncommon, represents more than a single distinct nucleation event.

Growth rate can vary with time or crystal size, but in instances where this can be checked this effect is not large (Cashman and Marsh, 1988). To be clear, growth rate does seem dependent on cooling rate (Cashman, 1993), but this effect is only significant in relatively narrow dykes ($\sim$10–20 m). Growth rates for minerals, both silicates and oxides, in larger basaltic bodies are typically $\sim 10^{-10}$ cm/s (Resmini and Marsh, 1995). Nucleation rates, on the other hand, are observed to vary greatly with time and cooling rate (Cashman, 1990; 1992; 1993). It is of central interest to investigate the fundamental control by nucleation and growth on the total number of crystals.

Number of crystals. From a log-linear CSD the total number of crystals (N_T) of any specific phase is the first moment of the CSD itself for that phase (e.g. Marsh, 1988*a*; eqs. 32–34),

$$N_T = n^0 G t_c^2 = J t_c^2 \qquad (4)$$

The total number of crystals is the product of the last nucleation rate and a characteristic time (t_c) to complete crystallization. An estimate of t_c can be found from the growth law and a characteristic crystal size (i.e. $t_c = L_c/G$), but this introduces a third parameter in (4) when what is desired is a function dependent on simply G and J. An independent measure of t_c can be obtained using an explicit crystallization model such as that due to Avrami (e.g. Avrami, 1939; 1940), which is frequently used in petrology (e.g. Kirkpatrick, 1983; Cashman, 1993; Hort and Spohn, 1991*a*,*b*).

For *constant* rates of nucleation (J) and growth (G), fractional crystallinity (ϕ) varies with time according to the well-known Johnson-Mehl-Avrami (JMA) equation.

$$\phi = 1 - \exp\left(-\frac{\pi}{3} J_0 G_0^3 t^4\right) \qquad (5)$$

Although homogeneous nucleation of spherical crystals is assumed throughout a simple binary system, which is most certainly not the case in magma, the parametric relations it provides are, on dimensional grounds, applicable to multiply-saturated magmas. Moreover, even when the nucleation rate is not constant but varies with time according to

(3), (5) has the same general form but with a larger numerical constant. Equation (5) predicts strong, sigmoidal changes in ϕ over time as a function of the rates of nucleation and growth.

A characteristic time (t_c) for the system to reach, say, 55% crystallization, which completes the crystallization of a single phase, can be defined by setting $\phi = 0.55$ and solving for $t = t_c$.

$$t_c = 0.93(JG^3)^{-1/4} \tag{6}$$

The variation in crystallinity with time (5) becomes

$$\phi = 1-\exp\left(-0.98\left(\frac{t}{t_c}\right)^4\right) \tag{7}$$

Using this time (6) scale in (4) gives N_T as

$$N_T = 0.93\left[\frac{J}{G}\right]^{3/4} \tag{8}$$

Taking $J = 10^{-3}$ cm^{-3} s^{-1} and $G = 10^{-10}$ cm/s, for example, $N_T = 1.8 \times 10^5$ crystals per cm^3, which is a reasonable result for basaltic volcanics (e.g. Resmini and Marsh, 1995).

Typical crystal size. From either the assumed growth law or the dominant size based on the CSD moment number density fractions (Marsh, 1988*a*; fig. 5), a typical crystal size is given by

$$L = Gt = 0.93\left(\frac{G}{J}\right)^{1/4} \tag{9}$$

This is the typical crystal size (i.e., radius) for Avrami-style crystallization under constant rates of crystal nucleation (J) and growth (G). This is essentially the functional relationship used by Winkler (see discussion in Shaw, 1965), by Shaw himself (1965) and rediscovered by Brandeis *et al.* (1984). Because of the small fractional exponent, crystal size is relatively insensitive to reasonable variations in J and G.

For the example cited above where $J = 10^{-3}$ cm^{-3}s^{-1} and $G = 10^{-10}$ cm/s, L = 0.17 mm (0.34 mm diameter). Although this is a reasonable size, it is too large by a factor of about 2 for the major phases grown after emplacement in Hawaiian lava lakes, dykes and many diabase sills, but for most plutons this size is too small by a factor of about 10. Phenocrysts in lavas and sills and crystals in basic plutons are commonly 2-8 mm in size (i.e. diameter, length, etc.). The combination of G/J yielding such a size can be estimated from (9). That is, for, say, at the very least L = 1 mm

$$\frac{G}{J} \sim 10^4 \tag{10}$$

And since, as has already been mentioned, growth rates seem to vary much less than nucleation rates, for $G = 10^{-10}$ cm/s, the companion nucleation rate is 10^{-6} cm^{-3} s$^{-1,}$ which is a factor of 10^3 smaller than that commonly found for lavas and sills (Cashman and Marsh, 1988*a*; Cashman 1990). But since growth rates in plutons are probably even smaller, this is an upper bound on the effective nucleation rate.

There is clearly some relation between the kinetics of crystallization and cooling. The common gauge linking crystallization to cooling has been the degree of undercooling. In laboratory experiments where charges are very small in size this is the obvious choice; the response time for imposed changes in temperature is very small. Experiments on silicate systems do show a quasi-classical dependence on undercooling, but these are at undercoolings much larger (i.e. typically 50–200°C) than can be expected in most any natural setting even near a contact beyond the first instants of emplacement. A more appropriate link between kinetics of crystallization and temperature in igneous bodies is the local rate of cooling. Cashman (1993) considers this approach for both natural systems and laboratory experiments and finds correlations between nucleation and growth rates and local cooling rate. Although chilled margins certainly point to rapid cooling and high nucleation rates, for the most part grain size in basaltic sills does not increase with sill size. Beyond the margins, uniformity in final crystal size and number suggests a relative constancy in G/J.

The important finding of these considerations is in the variation of numbers of crystals (N_T) and crystal size (L) with G and J, as shown by Fig. 13; calculated using (8) and (9). For mafic plutonic rocks, the observed grain size and population densities indicate very small nucleation rates $< \sim 10^{-7}$ cm^{-3}s^{-1}, much lower than expected from cooling rates, which strongly suggests that these assemblages have undergone a major, near solidus annealing event. This has systematically coarsened the assemblage by destroying the smallest crystals to the benefit of the larger crystals. It is exactly these crystals, which are generally called phenocrysts, that can be expected to be entrained by magmas ascending the lithospheric mush columns. We now turn to the evidence for this process.

7.2 *Phenocryst abundance.* The recognition of extensive flow differentiation in magma forming many sills and dykes (see 3.3) makes using the chilled margin an unreliable or 'fallacious' (Upton and Wadsworth, 1967) indicator of phenocryst abundance in the initial magma. Many basaltic sills of all sizes contain dense accumulations of large (2–10 mm) olivines or orthopyroxenes in the middle when the outer margins of the sill are essentially phenocryst free; their size alone precludes them from

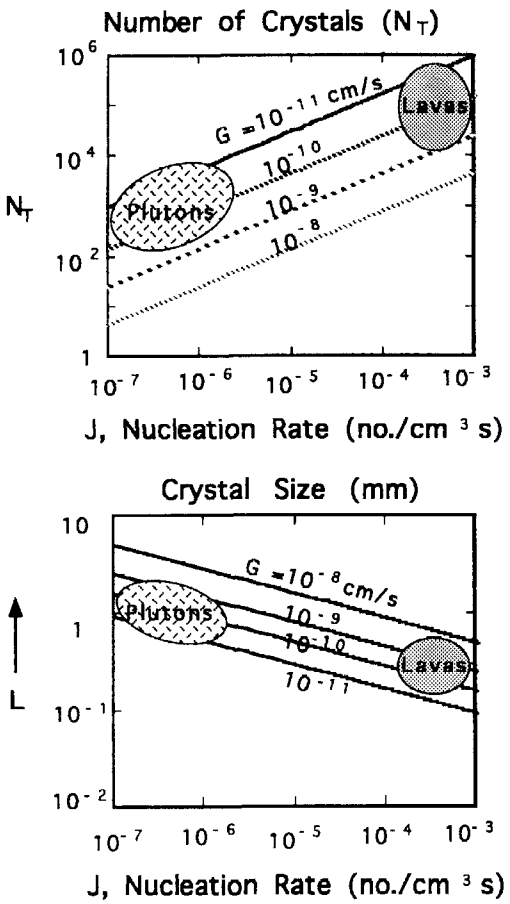

FIG. 13. The total number (N_T) and size (L) of crystals as a function of the rates of nucleation and growth expected in magmatic systems.

having grown *in situ* (see 7.1). The column of magma rising in a fissure or conduit is itself flow differentiated or sorted, with the leading batch and margins being poor or devoid of phenocrysts; the heavy crystals move to the centre and downward due to gravity and grain dispersive pressures. Vertical sorting in response to hydraulic equivalence (i.e. size and density contrast) also takes place, as in sedimentation (e.g. Greenspan and Ungarish, 1982; Marsh, 1988*a*), such that the filling magma may deliver a stream of phenocrysts already fairly well sorted (see Fig. 6). The degree of sorting clearly depends on the contrast in hydraulic properties of the crystals and liquid and on the time available for sorting, which is set by the ascent rate and distance from the point of entrainment of the crystals themselves.

In the eruption of Kilauea, for example, Murata and Richter (1966) have documented a clear correlation between eruptive flux and olivine size and concentration. They liken this transport of olivine to bedload transport of sediment in rivers. The Hawaiian lithosphere may be likened to a gravel bed of crystals through which magma ascends, the number of crystals entrained depends on their size and the rate of ascent of the magma. It is almost impossible to imagine magma traversing any such system without entraining some phenocrysts, presumably through fluidization. The overall volume and intensity of delivery of phenocrysts in any single episode of eruption or chamber-filling depends on the route of the magma and its chances of encountering a slug of older crystals. The magmatic system itself may be a tall stack of interconnected sills traversing the lithosphere (see Fig. 14) as has been depicted by Ryan (1988; 1994) using seismicity and buoyancy arguments. The narrower interconnecting passageways cool and congest, acquiring strength and becoming fractureable, sooner than the sills themselves where loose cumulates on the floors are relatively easily entrained.

Small volume, weak inputs to these sills or chambers may be almost free of phenocrysts, and large volume, vigorous inputs are likely to be heavily laced with phenocrysts. The vigor of delivery is measured relative to the sedimentation rate of the phenocryst swarm. In a suspension of four particle sizes of spheres of equal concentration (5 vol% each), for example, sedimentation typically occurs at a rate of about $V_S = 0.175\ V_L$, where V_L is the Stokes settling velocity of the largest particle (Greenspan and Ungarish, 1982). For entrainment and delivery of phenocryst-rich magmas at or near Earth's surface the ascent rate must be significantly larger than V_S, and must pass through several sills (Anderson, 1995).

That much of the olivine in picritic Kilauea and Mauna Loa lavas is entrained debris is also suggested by the strong olivine control lines exhibited by magnesia variation diagrams (Powers, 1955; Murata and Richter, 1966; Wright, 1971), which are characterized by a single olivine composition ($Fo_{87.5}$). The actual composition of olivine phenocrysts in the lavas themselves, however, is highly variable and ranges from Fo_{75} to Fo_{91}; moreover, the maximum foresterite content does not decrease with decreasing MgO content of the lava (Maaloe and Hansen, 1982). These findings are not restricted to Hawaii, for broadly similar relationships between lava composition and phenocryst content are observed for Jan Mayen (Maaloe *et al.*, 1988) and Reunion Island volcanics (Upton and Wadsworth, 1967); and the average olivine composition of the

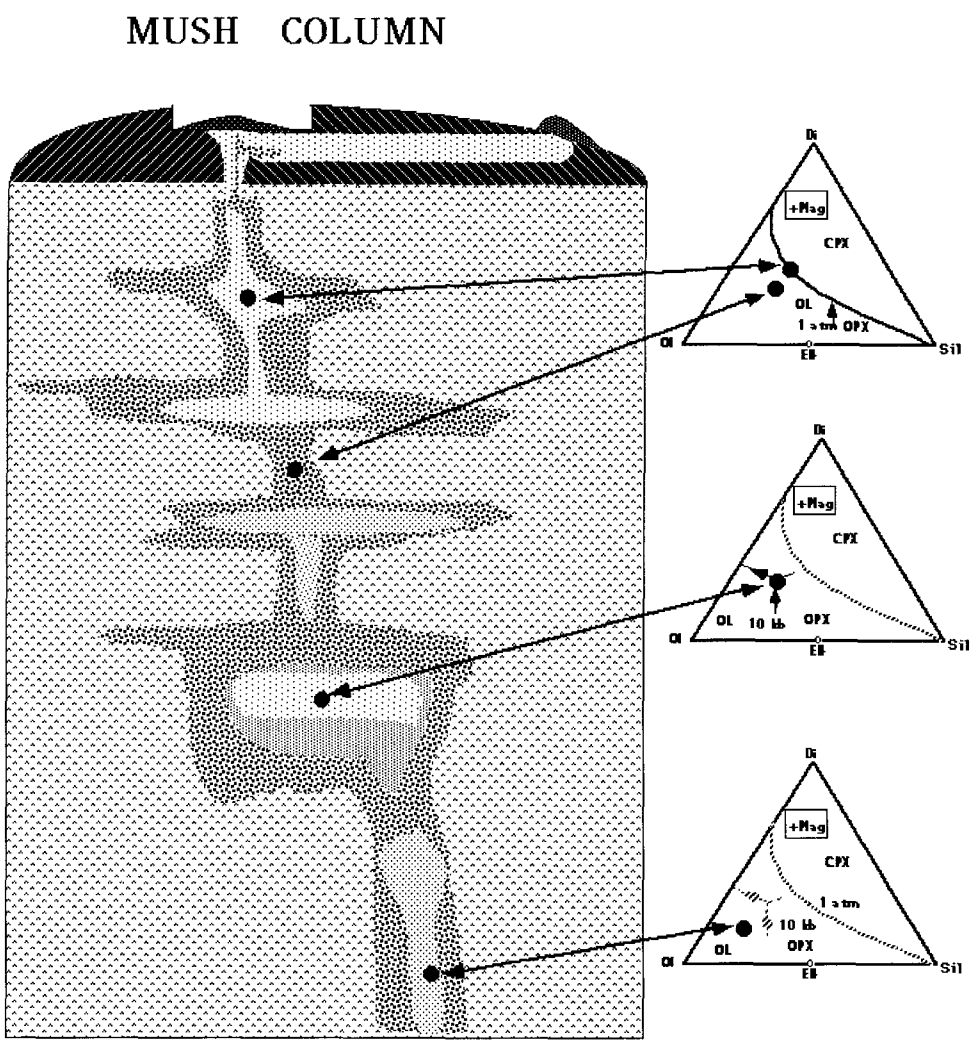

FIG. 14. A schematic illustration of a possible magmatic mush column existing beneath a highly active system such as Kilauea volcano, Hawaii. The mush column consists of a variety of local crystallization environments characterized by contrasting time scales for cooling and contrasting regimes of phase equilibria. Magma coursing this column will inherit crystals from a host of earlier crystallization events. There is also a tendency for magma of a low phenocryst content to be thermally preserved.

Rum intrusion is also about Fo_{88} within a similar span of compositions (Hunter and Cheadle, 1995). The synneusis of diverse olivine phenocrysts also reflects entrainment from multiple cumulate sources (Schwindinger and Anderson, 1989; Schwindinger and Marsh, 1994).

The full spectrum of observed lava composition from 20 to 7 wt.% MgO for the Kilauea eruption of 1959, for example, can be produced simply through addition or subtraction of olivine of composition $Fo_{87.5}$, yet the eruption traversed at least three sills during ascent (Schwindinger, 1987).

.....in Kilauea lavas, this process is reversible, since olivine is always the first mineral on the liquidus. .....where "tramp olivine" is continually being transferred from place to place by 2-phase (crystal-liquid) movement.....

(Wright and Fiske, 1971).

The initial position of the olivine control line is slightly different for each Hawaiian eruption phase, reflecting slight compositional differences in each dominant magma. But the most differentiated end member summit composition is never more siliceous and less magnesian than, respectively, about 51.5 and 6.8 wt.%.

It seems remarkable that the eruptive mechanism of Kilauea should function so precisely as to deliver repeatedly to the summit a differentiated magma of such a circumscribed composition

(Murata and Richter, 1966).

This composition marks the transition from a singly saturated system (i.e., olivine) to a multiply saturated system (olivine, plagioclase and clinopyroxene), and it is remarkable that separation of the entrained olivine always brings the magma to this same composition. Remarkable, because if the initial crystal-free magma contained, say, 12 wt.% MgO and then entrained a batch of cumulate olivine, upon loss of this olivine the bulk composition should revert to the same initial composition. That this evidently does not happen implies the process is not entirely reversible. The entrainment of olivine seeds the magma, inducing further precipitation of olivine, dissolves some olivine while undergoing Fe/Mg diffusional exchange with the magma, of which all three processes drive the bulk composition to the three phase cotectic. Direct evidence of these three processes occurring concurrently is found in the olivine phenocrysts of the 1959 Kilauea Iki eruption (Schwindinger, 1987). Subsequent loss of olivine during intermittent chambering brings about punctuated differentiation and yields a phenocryst-free magma in which further fractionation is controlled by both multiphase nucleation and solidification front propagation (see Fig. 7).

We have already seen that the three phase tholeiite cotectic marks the leading edge of the Kilauea solidification front. The solidification front contains the more fractionated melts, which, although inaccessible to eruption, can be locally segregated through solidification front instability (see section 4). The effects of these various processes on the final sequence of rock is illustrated for Kilauea Iki lava lake by Fig. 15. This series of interrelated processes, largely reflects the influence of two central features of the magmatic system: magmatism through a column of crystal-liquid mush with concurrent re-equilibration and differentiation of phenocryst-bearing magma by punctuated differentiation. We will return to this basic system of a sill on a mush column later in relation to ocean ridge magmatism.

It is important to realize that regardless of the ultimate origin of the erupted phenocrysts, magma or mantle wall rock, diffusional exchange with the melt may all but obliterate any information on source provenance. In a purely chemical sense, at one extreme the mush column is much like an exchange column (e.g. Navon and Stolper, 1987), although the common occurrence of a wide spectrum of olivine compositions clearly shows that this extreme is never attained. In this respect, the absence of other phases like clinopyroxene and orthopyroxene, which are essentially ubiquitous to mantle rock, is curious. This absence almost certainly reflects the low pressure ($< \sim 8$ kb) phase equilibria in the picritic-tholeiitic system (Walker *et al.*, 1979; Stolper, 1980;). These phases are simply not stable in an essentially olivine dominated mush column (see Fig. 14).

7.3 *Phenocrysts from solidification front erosion.* The compositions found in Kilauea Iki lava lake can be spatially separated into two basic regions (see Fig. 16): those due to the filling or eruptive event, which are the olivine controlled compositions; and those differentiated melts residing within the solidification front, which are due to multiple saturation. As we have remarked previously (Marsh *et al.* 1991), this variation in composition is similar to that shown for the whole island of Hawaii itself (see Fig. 16) (Peterson and Moore, 1987). Because more differentiated compositions do exist, however small in volume, some mechanism must exist to make these melts accessible to eruption.

Kilauea lavas having less than 6.8 wt.% MgO are found only along the rift zones extending east and southwest from the summit (Wright and Fiske, 1971; Ho and Garcia, 1988; Helz and Wright, 1992). Lavas reaching 3.9 % MgO and 56 % SiO_2 are found, and all such lavas show no olivine control on bulk composition, but instead show effects of clinopyroxene, olivine, and/or plagioclase fractionation. The presence of these additional phases in the fractionation sequence and the spatial restriction of these lavas to the flanks of Kilauea clearly suggests that further transport of these magmas from the summit region to the rift zones has partly eroded the developing solidification fronts. The eroded material mixes with the interior summit magma, slightly lowering the overall magma temperature, which disperses and stabilizes the new three phase assemblage throughout the magma, allowing further crystal growth and significant crystal settling. The net result is a

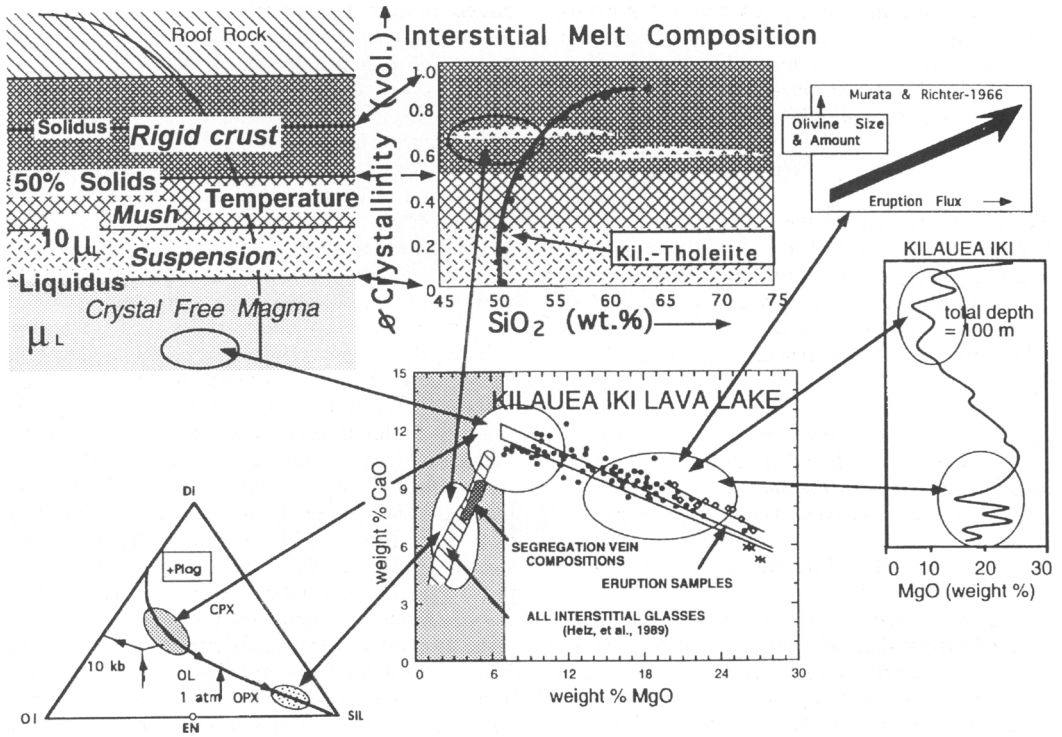

FIG. 15. The observed variation in composition of lavas filling Kilauea Iki lava lake and of the melts found within the partially crystalline upper solidification front (lower centre). The scatter in the compositions of the lavas at high magnesia contents reflects the variation in the size and modal content of olivine phenocrysts in the initial erupted magma, which is dependent on the eruptive flux (upper right). The wiggles in the S-shaped distribution of phenocryst content reflects this same variation in the magma filling the Kilauea Iki (lower right). The compositions of melts contained within the solidification front (see regions within shaded band, lower centre) appear as segregations within the Rigid Crust (upper centre) of the solidification front (upper left). Note that the composition of these melts increases strongly in silica content only after about 50 % crystallization (heavy curve, top centre). The transition from olivine control to multiple saturation is near 7% MgO, which marks the leading edge of the solidification front after loss of phenocrysts (lower left). Chemical variations among a group of co-magmatic lavas reflect spatial variations in the physical processes of solidification fronts.

differentiated magma that shows evidence of mixing of summit and flank magmas, as observed at Kilauea (Wright and Fiske, 1970). The degree of differentiation depends on the volume of the front eroded relative to the volume of relatively crystal-free summit magma available for mixing. But because the front becomes rigid near a crystallinity of ~50 %, it is unlikely in any single episode of transport, erosion, and holding to attain more than about 25 % fractionation.

Gabbroic xenoliths in 1960 Kilauea basalt, in fact, may be cumulate materials from solidification fronts that produced the earlier evolved 1955 lavas of the east rift zone (Fodor and Moore, 1994). Based on the highly open textures of some of these xenoliths, Fodor and Moore suggest derivation from the Mush Zone and from near the transition to the Rigid Crust within the front. Broadly similar samples of eroded solidification fronts have been identified in basalts from Kahoolawe Island, Hawaii (Fodor *et. al.*, 1993; Rudek *et al.*, 1992).

This process is of limited effectiveness in accomplishing much silica enrichment, and only by repetition can significant differentiation occur. Although fissure transport of magma from the central vent is well recognized in both Hawaiian and Icelandic systems (Ryan, 1988; 1990) and solidification front erosion is clearly a natural consequence of this dispersal, the fact that fractionation calculations most often show about 35–50 wt.%

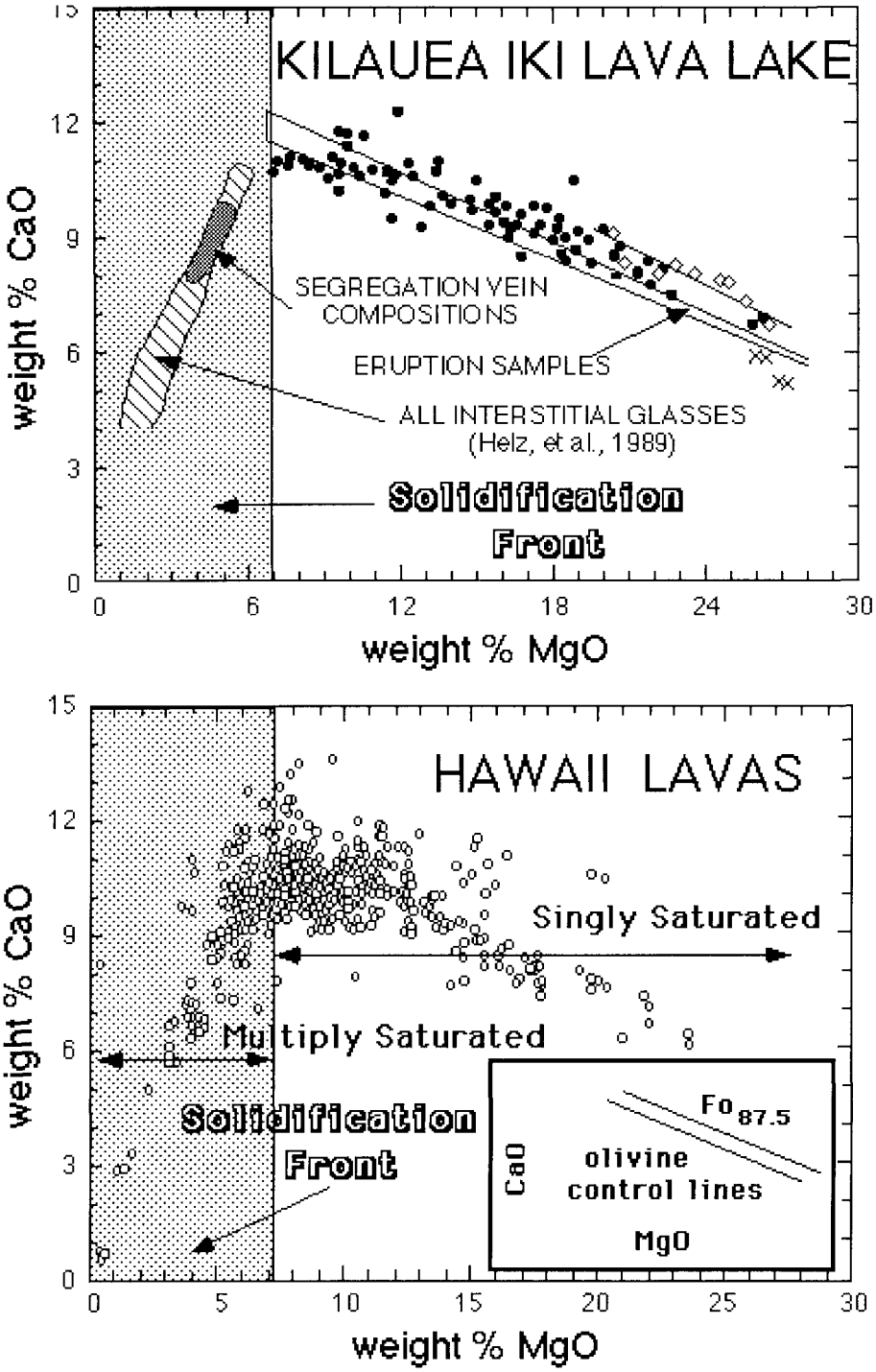

FIG. 16. A comparison of the composition and location of melts in Kilauea Iki lava lake (upper) with observed compositions of lavas from the whole island of Hawaii (see text).

removal of solids suggests another mechanism of melt extraction from the solidification front.

7.4 *Filter flow fractionation.* When magma is held long enough in a fissure system or conduit to allow not only the solidification fronts in certain locations to meet but for the Rigid Crusts to meet, the interstitial melt can be removed through flushing or replacement by invading juvenile melt from deeper in the system. Once the entire fissure is overtaken by Rigid Crust, the solidification front has considerable strength (see section 4) and becomes a porous medium saturated with highly fractionated melt. The porous flank fissure may be reactivated by the arrival of new magma from the mush column, which purges the fissure of its fractionated melt, pushing it forward into less solidified parts of the fissure system. Ho and Garcia (1988) call this action an hydraulic plunger. Subsequent eruption produces three compositions: 1) differentiated magma; 2) mixed magma, consisting of purged and juvenile magma; and 3) juvenile or primitive magma. These magma types are each characteristic of Kilauea volcanism (Wright and Fiske, 1971), but the greatest volumes will be of 1) and 3). The petrologic evidence of this process will show strong crystal fractionation, which is what is observed at Kilauea (Ho and Garcia, 1988; Garcia *et al.*, 1989; Russell and Stanley, 1990). In the 1955 eruption of Kilauea, this was the first major eruption on the east zone for a period of 115 years; this long repose time alone points to extensive solidification front development and suggests an effective congested fissure thickness of about 200 m.

7.5 *Summary.* The composition and differentiation of basaltic magmas in vigorous, large volume systems, like Hawaii, are often controlled by the abundance of large concentrations of cumulate debris entrained during ascent. These crystals represent a wide range of conditions of cooling and fractionation, and continually strive to reach equilibrium with the local prevailing magma composition. Because these crystals are often large (e.g. $\sim$2 mm, but are sometimes up to $\sim$8 mm) and settle easily in basalt, whenever the ascending magma hesitates or comes to rest during ascent, strong crystal fractionation occurs and the bulk composition is modified almost discontinuously. This is punctuated differentiation. The process is quasi-reversible through further ascent and entrainment of cumulates. In near-surface intrusions it is clearly of importance to understand the filling event not only in terms of the input flux as a function of time but especially in terms of the abundance and size of entrained phenocrysts. The final crystallinity and differentiation of the body must be divided into those effects due to the behaviour of the initial phenocrysts, those due to crystals grown or recrystallized after emplacement, those due to the internal dynamics of the solidification fronts (e.g. SFI), and those due to the tectonic style of the magmatic regime, which determines the eruption and replenishment rates. A good example of the results of these effects may be found in the products of ocean ridge magmatism.

8 Sill on a mush column: ocean ridge magmatism

Our model includes a thin, narrow, sill-like body of melt overlying a thicker, wider crystal mush zone at fast spreading ridges ..... and a short-lived, largely mush-filled chamber without a melt lens at slow spreading ridges.....

(J. M. Sinton and R. S. Detrick, 1992)

It is difficult not to see a remarkable similarity between the basic petrological features of the oceanic crust and ophiolites and the dynamics of sill solidification. The overall gabbroic composition, the medium scale grain sizes, the position, form, and composition of the plagiogranites, and the limited downward detailed layering each suggests a close dynamic correspondence to diabase sills. There are also striking differences: the enormous thickness of oceanic crust relative to any continental sill, the pillow lavas and sheeted dykes, and the textures in the gabbros. These differences may, however, mainly reflect the long-term record left by a sill-like body, attached to a mush column, which is continually stretched, erupted and replenished. The ridge setting may thus be of critical importance to understanding the role of relatively small, periodically fed and continuously deformed sills in providing a fundamental organizational process for enormous plutonic complexes.

8.1 *Ridge sills.* There is good geophysical evidence to believe that active magma chambers beneath some ocean ridges may in fact be sill-like magma lenses some 100–300 m thick and up to perhaps 2 km wide (e.g. Detrick *et al.*, 1987; Sinton and Detrick, 1992). They are essentially exactly analogous to diabase sills in dimension, level of emplacement, and even composition, although most diabase sills are slightly more siliceous and contain orthopyroxene instead of olivine. This broad similarity should not be too surprising being that both arise from similar rifting processes. The position (i.e. depth), viability (i.e. size), and along strike continuity of these ocean ridge sills each depends critically on the associated thermal regime, which is reflected, in the broadest sense, by the spreading rate.

Fast spreading ridges like the East Pacific Rise (EPR) strongly squeeze the bundle of lithosphere or thermal boundary layer isotherms close to the sea floor, promoting high heat flow, thin ridge-area

lithosphere, low topography, and thermal conditions hospitable to sustaining a relatively thick sill. Whether or not magma is supplied continuously along the ridge or spreads from preferred locations, possibly reflected by an hierarchy of segmentation (e.g. Macdonald *et al.*, 1993), is still unclear. With slower spreading, the thermal regime systematically relaxes, producing a cooler overall near-ridge structure, lower heat flow, a thicker near-ridge lithosphere, more severe near-ridge topography, and generally shrinks and pushes deeper the conditions conducive for long term sill viability. Numerical studies suggest no steady state sill at all for half spreading rates less than about 2–3 cm/yr. (Phipps Morgan and Chen, 1993*a,b*, and Parmentier and Phipps Morgan, 1990).

Barth *et al.* (1994) have carefully pointed out the basic geophysical and petrological similarities between Kilauea Iki lava lake and ocean ridge magmatism. Although they find the analogy imperfect, they infer a number of small scale processes and features of the ridge sill to be analogous to process in lava lakes. We shall see that the analogy is closer if a substantial mush column is added to the lava lake model.

Ophiolites, long used for insight into generation of the fundamental structure of oceanic crust, now seem to be dynamically more closely allied to the ephemeral magmatism of the cooler slow spreading ridges (Quick and Denlinger, 1993; Malpas, 1993). Magmatism here is characterized by sporadic emplacement of spatially and chemically inhomogenous, relatively cool magmatic suspensions. Ongoing deformation leads to "syntectonic differentiation", which squeezes late-stage, iron-rich interstitial melt into thin shear zones producing strings of Fe-Ti oxides (Dick *et al.*, 1991). Similar structures are seen in diabase sills in association with plagiogranites and granophyres (Kanaris-Sotiriou and Gibb, 1989; Grossenbacher, 1994). These local chemical variations may indicate filter flow fractionation (see 7.4), which also may be discernible in the detailed sampling of lavas of fast spreading ridges (Perfit *et al.*, 1994). The progressive deepening of ODP hole 504B to over 2 km has produced a rock record of diabase sheeted dykes of a bulk composition matching the overlying lavas. Coupled with the Hole 735B gabbro record, the sheeted dykes are underlain by a thick sequence of unlayered, albeit syntectonically differentiated, gabbros followed by (as yet unsampled in drilling) layered gabbros, before encountering the underlying ultramafic assemblages.

This sequence is exactly that expected from a periodic and abrupt stretching of a diabase sill overlying a crystal-gorged feeding column. (see Fig. 17). That is, rifting spawns broad upwelling, deep partial melting, magma formation, ascent and sill initiation. Subsequent abrupt spreading pulls aliquot after aliquot of magma in the form of dykes and lavas from the sill while, simultaneously, more magma is pulled into the sill from the underlying mush column. The deeper portion of which is ultramafic and the upper portion is gabbroic. The degree of causality between spreading, injection rates, and magma composition and crystallinity is clearly a fertile ground of interplay. The influxing magma may, depending on its rise rate, carry phenocrysts that enter the sill and commonly settle to form layering. The bulk of the eruptable sill itself, however, on a time averaged basis is mostly of low phenocryst content; which means that the crystal settling time is less than the repetition rate of the abrupt spreading-dyke formation events themselves.

The fact that the sheeted dykes are nearly phenocryst free, fine-grained diabases reflects the fact that the sills are, at least in their eruptable region, on average low ($< \sim 25\%$) in phenocryst contents. Even in MORB itself, less than 6% of the lavas studied by Bryan (1985) have more than 25% phenocrysts (Sinton and Detrick, 1992). This limit of $\sim 25\%$ crystals defines the edge of the Suspension Zone (see 2.1), which marks the transition between eruptable magma and the surrounding envelope of the Mush Zone. There is also every reason to believe, in analogy to diabase sills, that ridge sills sometimes also contain tongues of phenocrysts entrained from the underlying mush column. The extent of entrainment depends on the rate of magma transport, which is dependent on two distinct, but interrelated processes.

One process is a pervasive overpressure attending the regional mantle upwelling associated with global thermal convection. Decompression melting and buoyancy create a persistent overpressure, tending to drive melt upward regardless of the local details of spreading, dyke formation, and extraction of melt from the ridge sill itself. Dyke formation in response to spreading is thus the second process. That these two processes are not one and the same is seen in the rock record furnished by the sheeted dykes and, to a lesser extent, the lavas. That is, if magma transport, injection, and eruption were all due to a regional overpressure, dykes and lavas would commonly be found carrying large amounts of entrained phenocrysts. The great paucity of picritic MORB lavas, for example, attests to the difficulty of entraining crystals from the deeper mush column during fresh injections. It instead appears that dyking is a tectonically-induced process of magma withdrawal from a passive sill or mush pile. Injections to the sill are only in the broadest sense coupled to dyke and eruptive withdrawals. In this respect, the sheeted dyke complex is massive testimony to the Null Hypothesis (see 3.2): the sill at any time is essentially

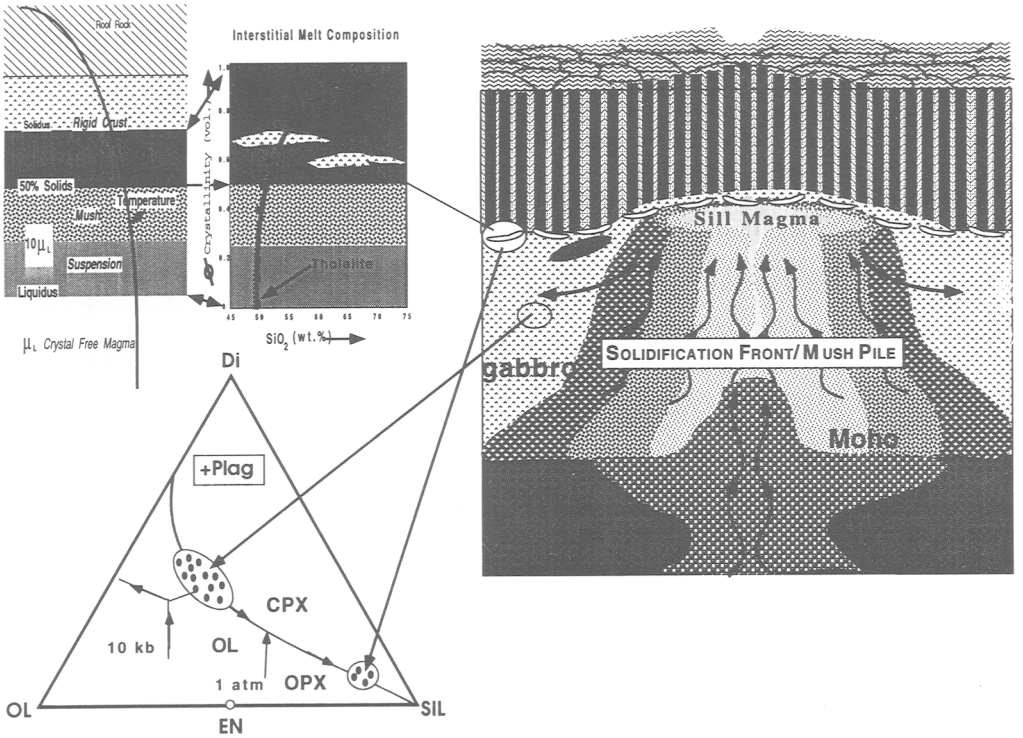

FIG. 17. The magmatic system at a **fast** spreading ocean ridge (right) as related to the structure of solidification fronts (upper left) and the associated phase equilibria (lower left). The ridge magma chamber is dominated by a thick mush pile or solidification front arranged symmetrically about the summit axial region, which compacts and spreads laterally, evicting melt and processing melt ascending from the underlying mantle mush column. The small sill-like magma chamber reflects an excess of melt evicted relative to that crystallized. At **slow** spreading ridges there is no steady state existence of the sill, yet enough low crystallinity melt must still exist to form the sheeted dikes. A broadly similar high level mush pile may exist beneath large continental silicic systems. (See text for further discussion.)

free of phenocrysts and fractionation processes are of limited effectiveness in differentiating the magma.

As mentioned earlier, however, lateral transport at Kilauea may lead, through filter flow fractionation (see 7.4), to more fractionated lavas being at the distal parts of the system. At fast ridges, vestiges of the sill existing beyond the axial summit caldera are contained within the Mush Zone and Rigid Crust of the solidification front. They are wholly congested with crystals, but the interstitial, residual melt can be expelled by more primitive melts invading laterally or from below in response to overpressuring in the main system, or from upward eviction in response to compaction and deformation in the contiguous gabbroic mush (Langmuir *et al.*, 1992; Langmuir, 1989; Quick and Denlinger, 1993). The eruptive record may thus show any combination of lavas from unfractionated to fractionated. The detailed sampling by Perfit *et al.* (1994) on the East Pacific Rise shows significant diversity on a small (~600 m) scale.

Just as at Hawaii, there is no sign of any silicic *lava* in the ridge system. The stratigraphic position, overall texture, and composition of the plagiogranites/trondhjemites correspond fairly well with silicic segregations in mafic sills; although the former can be even more siliceous. Sharp upper contacts, often forming a 3-dimensional fractured network of silicic rock and diabase and gabbro (Pallister and Hopson, 1981; Dick *et al.*, 1991), is, for all intents and purposes, exactly what is seen in the Ferrar dolerite sills of Antarctica (Wheelock and Marsh, 1993). What is different in the plagiogranites, however, is that they show in cross-cutting relationships signs of remobilization; occasionally they are

even found entrained in dyke rock. This is strong evidence for continual growth and erosion, probably through cycles of cooling and reheating, of the roofward solidification front. The front thickens to the point of undergoing instability (see section 4), producing zones of silicic differentiates, then replenishment of the system with new magma in essence 'burns back' the front, remobilizing the segregations and collecting them at high levels near the sheeted dykes. There is little chance of evicting them as lavas for they are always controlled by the solidification front. The eventual subduction of plagiogranite lenses presents the distinct possibility that they could reappear in the ridge system as dykes or lenses at essentially any level in the oceanic crust. The scale of plagiogranites is too small to allow for significant shear and diffusional assimilation back into mantle peridotite.

The high silica contents of some plagiogranites are difficult to match unless perhaps 95% of the magma is fractionated at which point small regions of immiscibility may arise (Philpotts, 1982; Spulber *et al.*, 1983; Dixon and Rutherford, 1979). It is doubtful, however, whether these patches can be collected without wholesale instability of the upper solidification front (see Fig. 8*b*), which is especially difficult at slow spreading ridges.

It is, indeed, curious that the structure and composition of the oceanic crust can be, broadly speaking, so similar wherever it is seen, be it as ophiolites or in ocean drilling, when the nature of the magmatic regime can be so different. Whether or not an actual layer or zone of low crystallinity magma exists seems to be of secondary importance to forming a generic variety of oceanic crust. This incongruity may be of fundamental importance to the concept of magma chambers in general.

9 The ridge magma chamber as a general concept

The most differentiated compositions of plagiogranite and diorite in the Oman ophiolite being near the base of the sheeted dyke complex, compelled Pallister and Hopson (1981) to interpret this as the sandwich horizon of a massive ocean ridge magma chamber. A single, 5 km thick by 30 km wide, laterally wedging chamber riding on the Moho, and solidifying as a crystal-free magma from the floor up satisfies the ophiolite rock record and produces the oceanic crust. With no clear plutonic evidence of any record of the phenocryst content and frequency of the filling events, there is no choice but to adopt such a model. With no chilled margins, the kinetic record of inputs of nucleating and growing crystals has been annealed from the initial CSDs. There is no easy way to discern new crystals from old; and the record left by the lavas and dykes is interpretable only through a model of what is going on in the immediate staging area below. The overall problem and evidence are almost exactly that available to Becker, Pirrson, and Grout at the beginning of this century, and the most reasonable solution, until recently, is essentially the same.

Although the new model of ocean ridge magma chambers appears as a distinct break from the age-old concept of magma chambers, there is still a formidable problem at hand. The enormous mush pile that goes to form the oceanic crust cannot be simply crystals entrained from the mantle mush column, because the bulk composition does not fit. At the same time, the whole gabbroic mush pile cannot be generated in the capping sill, for this involves bringing all magma into the same small, ephemeral region for processing and cumulate formation, which is especially difficult at slow spreading ridges where the sill may exist only a fraction of the time. If the structure of the oceanic crust were in fact due to the conventional processing of primary melts in a sill-like body, it would be expected to be highly variable in basic form and composition, much more so than is actually observed.

It is instead possible that the sill, although functioning as a real high level chamber when actually present, merely represents pooled or excess melt evicted during compaction of the underlying gabbroic mush. New magma enters the mush pile and reacts variously with the existing crystals, which buffers the melt composition and allows interaction with earlier fractionated melts. Melt eviction during compaction cannot be too efficient at high crystallinities, for the pooled sill magma, aside from blobs of plagiogranite, evidently never issues highly fractionated dykes and lavas. Textures in the mush pile are set by a population density of crystals prescribed by packing, deformation, and local annealing; nucleation rate in the usual sense is not coupled to the cooling rate, but is set by the existing crystal population in the mush pile; and crystal size is determined by residence time in the mush pile. Once established the texture is self-sustaining.

The magma chamber, *per se*, is essentially a mush pile or thick solidification front (see Fig. 17). The magmatic sill may be merely a secondary result of a more hospitable thermal regime and more compaction-driven melt eviction than can be extracted to form dykes and lavas. With an increase in spreading rate, the melt flux from the mantle increases, the near-surface thermal regime intensifies, reaching near-liquidus temperatures just below the sheeted dykes, and the sill begins forming. The heightened thermal regime shrinks the thickness of the basal solidification front. The actual magma chamber, where the most extensive crystal-liquid chemical interactions occur, is in the mush pile, not in the sill.

Distinctive geochemical signals come both from the mantle and the mush pile. But when the sill is present it functions essentially as an homogenizer by tending to dampen these signals through mixing. That it clearly does not erase these signals may reflect a relatively low level of dynamic activity in the sill, as has been suggested for continental sills by Marsh (1989) and Gibb and Henderson (1992). The gabbroic mush and the sill both, in effect, serve as compositional buffers. Broadly similar processes may also dominate the chemical diversity of continental magma chambers.

10 Magma chambers as solidification fronts

Magma in any setting must ultimately have some connection to a mush column, mush region, or mush pile. The more voluminous the system the closer the spatial connection between the high level chamber and the column, and the more extensive the interaction. Magma repeatedly traversing the lithosphere at select locations, as in island arcs and major intra-plate volcanic centres, must hydraulically see the lithosphere as essentially a gravel pile of crystals, which may become entrained and reworked. As much as is possible within the limitations of time and phase equilibria, the melt and tramp crystals buffer the compositions of one another. Only through repeated, long term use do the ultimate source charateristics become clear (e.g. Myers *et al.*, 1985). And the restricted size range of mafic minerals in both magmas and cumulate piles may ultimately reflect (1) a delicate balance between magma ascent rate and the size of crystals entrained, and (2) the time spent by crystals in propagating solidification fronts.

The single basic fact that the composition of basaltic magma does not dramatically change until about 50 vol.% crystallization has occurred (e.g. Ghiorso and Carmichael, 1987; see also Fig. 15 top centre), at which point the crystals are at maximum packing and unsortable, restricts significant differentiation to processes occurring deep within the solidification front. Recall that a chamber at 50% crystallinity at the centre is fully congested everywhere and, on average, saturated throughout with, phase equilibria permitting, highly fractionated melt. Differentiation thus may be a problem of understanding the life of late-stage interstitial melts relative to the local rate of solidification. Thin fronts are fast propagating, stabilizing fronts, allowing little chance for melt migration and chemical diversity. Voluminous fronts are thick, slow fronts, allowing ample time for late-stage melt migration initiation of chemical diversity.

Magmatic diversity in any single body is therefore critically dependent on the volume of the active mush or solidification front reservoir relative to the volume of crystal-free magma. This fundamental necessary condition for spawning diversity is made sufficient only when coupled with an environment of phase equilibria conducive to both diversity and migration of the interstitial melt, as McBirney (1995) has found for Skaergaard. Moreover, disproportionately thick, in essence, artificial mush piles are made by filling large, slow-cooling chambers with crystal-laden magma, which is a process at the heart of the difference between lavas, sills and large layered intrusions (Gibb, 1976). Chambers formed deeper in the crust, where the contact temperature may approach that of the solidus itself, will also have unusually thick solidification fronts. In this respect, there may be an overall similarity between the gross structure of ocean ridge chambers, aside from rifting, and those chambers beneath the large continental silicic calderas.

The especially clear records of the nature and behaviour of phenocryst-laden ascending magmas of dykes, sills, and volcanoes have generally been considered a world apart from plutons, especially the layered intrusions. But they are clearly parts of the same continuum. The ideas that magmatic systems are periodically resupplied (Brown, 1956) by magmas carrying important amounts of crystals (Gibb, 1976) that undergo extensive sedimentological sorting (Irvine, 1987), reworking, and recrystallization are concepts central to understanding magma.

The century-old, versatile, and insightful magmatic process originally delineated by Becker, Pirsson, and Grout and carried on to this day to foster petrologic diversity has emphasized the importance of a highly mobile chambered magma with a paucity of crystals. Igneous diversity in both basaltic and granitic magmas is probably more intimately connected to magma that is mostly mush.

Both the questions and the answers are in the rocks.

Being driven by the rocks is something I especially admire ...., ... because we both know that without the constraints of the real world we are only doing what Robert Frost said about free verse — playing tennis with the net down.

S. A. Morse, Letter to B.M., 1993.

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