

On some fundamentals of igneous petrology

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Received: 5 October 2012 / Accepted: 28 May 2013 / Published online: 22 August 2013
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Abstract The age-old process of crystal fractionation leading to the diversity of the igneous rocks and Earth itself is an exceedingly well-understood chemical process in magmatism and physical chemistry. But the broader physical aspects of this and related processes have proven elusive on many fronts, especially in its relation to the spatial variations in rock composition, texture, and macroscopic features like layering. Magmatic systems, be they volcanic, dikes, sills, or plutons, are generally analyzed with a problem at hand and an end result in mind. The processes invoked to solve these problems, which are most often purely chemical, are often unique to each problem with few if any general principles emerging that are central to understanding the wider perspective of magmatic processes and problems. An attempt is made at the outset to provide a list of inviolate Magmatic First Principles that are relevant to analyzing most magmatic problems. These involve: initial conditions; critical crystallinity; solidification fronts; transport and emplacement

fluxes; phenocrysts, xenocrysts, primocrysts; crystal size; layering and crystal sorting; thermal convection; magmatic processes are physical. Along with these principles, two reference magmatic systems are suggested where the initial conditions and outcome are unequivocal: the Sudbury impact melt sheet and the Hawaiian lava lakes. Sudbury formed in ~ 5 min by superheated magma crystallized to a near uniform sequence, while the tiny lava lakes, formed of crystal-laden slurries, form a highly differentiated layered sequence. The major difference is in the initial conditions of formation, especially the nature of the input materials. The challenge is to construct and analyze magmatic systems (i.e., magma chambers, sills, dikes, and lavas) using these reference end members and the suggested principles. The Hawaiian 500,000 year volcanic record exhibits what can be expected as input materials, namely a highly varied output of magma of an overall composition reflecting the abundance of entrained olivine primocrysts. The provenance of these crystals is varied, and within any single sample, the population may be highly heterogeneous in composition from crystal to crystal, yet the overall pattern of chemical fractionation is exceedingly regular and well defined. If similar inputs go to form large intrusions, these systems will undoubtedly be dominated by crystal-rich slurries, which provide a vast set of physical processes promoting exotic layering and, at the same time, given the effects of annealing and continued crystal growth, a final chemical record adhering to all the time-honored effects of crystal fractionation. The long assumed initial condition of instantaneously emplaced crystal-free magmas cannot reasonably produce the observed rock records.

To I. S. E. Carmichael, who enthusiastically nucleated and grew a band of young warrior petrologists to go throughout the world seeking adventure in igneous rocks.

Communicated by G. Moore.

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Keywords Magma · Magma chambers · Magmatic principles · Slurries

Introduction

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 uprising of the magma and extrusion at the surface or injection into cooler rocks of the upper crust.

I.S.E. Carmichael, 1974

It goes almost without saying that magmatic processes have formed and shaped Earth from its very beginnings. The major divisions, right down to the formation of continental crust and the diversity of rocks themselves, owe their origins to magmatic differentiation. At the core of differentiation is the seemingly simple process of crystal fractionation, which involves the physical separation of crystals from the associated melt. The chemical aspect of this is abundantly clear. The melt composition evolves in direct response to the removal of the chemical mass held by the crystals. This is a common age-old physical chemistry technique, used by Madam Curie to isolate radium. In complex systems, the crystals are always in compositional contrast with the melt, so the melt composition evolves. Yet, a key ingredient to gaining new and unusual insight into the broader fundamentals of planetary magmatism and differentiation is the appreciation of the intimate spatial connection of chemical and physical processes. It is the specific physical processes that largely remain elusive, namely those governing the general functioning of magmatic systems leading to protracted differentiation via crystal fractionation, the distinctive layering in mafic plutons, and the connection between deep magmatic processes and volcanic processes.

Early in the last century Norman Bowen (e.g., 1928) repeatedly demonstrated that prolonged fractional crystallization of basalt can lead to residual melts of granitic composition. Bowen's reaction series became synonymous with fractional crystallization. Continents, with an average composition of ~60 wt% silica, can be made from basalt. And partial melting of more primitive rocks, like peridotite, can make basalt. Continents can thus come from peridotite. In retrospect, one might have said: 'Of course, what else is there?' And Bowen and untold workers since have shown in wonderful detail how the physical chemistry of common magmas truly reflects Earth's diversity of igneous rocks. And peridotites themselves are not the most primitive rock; they can be made from carbonaceous chondrites, the very building blocks of Earth itself, by draining away the excess iron to the core and expelling the volatiles. Moreover, carbonaceous chondrites have solar elemental abundances. So, from nucleosynthesis to continents, everything looks good, and

this is pretty much exactly correct in a strict chemical sense, even though there are many specific second-order unsolved problems here and there throughout this sequence. But, by and large, there is no basic chemical problem. The fundamental problem is physical. How does it actually happen?

In massive basaltic systems possessing the greatest potential for protracted fractional crystallization little to no appreciable silica enrichment occurs. And that which does occur is mainly not due to the conventionally held scenario of simple cycles of crystal nucleation and separation. It is more closely tied to the entrainment of masses of pre-existing crystals, their re-equilibration, loss, and re-involvement in a process closely tied to the actual rate and style of transport of the magma itself. The recognition and appreciation of the physical processes attending this life cycle show it to be also intimately connected with the deep-rooted perplexing problems of the origin of magmatic layering, the origin of continents, and even the basic compartmentalization of chemical divisions in Earth.

The general *association* of silicic or rhyolitic magma with massive basaltic systems is sometimes abundantly clear (e.g., Snake River, Thingmuli, Torfajökull), and convincing chemical evidence, commonly invoking permissive AFC processes, can be presented linking the entire compositional sequence. But in each of these instances, there is no clear physical or detailed spatial connection, and when physically and chemically scrutinized, as at Torfajökull, the largest outpouring of rhyolite in Iceland, the genetic heritage of the silicic magma is found not to be simply related to the basalt in any conventional sense (e.g., Gunnarsson et al. 1998). Bimodality without the requisite intermediate compositions in these systems is a clear indication of conflict with simple fractional crystallization. Bimodality can clearly be produced by wholesale fractionation related to tearing or gravitational instability of roof-ward solidification fronts (e.g., Marsh 1996, 2002), which leads to a wide variety of granophyric associations. The spatial-physical evidence for this is abundant in sill-like bodies, but the key evolutionary physical process in going forward to make viable bodies of voluminous silicic magma is that of ongoing reprocessing where the silicic signal can be separated from the basalt and accumulated (Gunnarsson et al. 1998; Marsh 1996). Reprocessing, as in Iceland, spawns silicic magma, but without it, as at Ocean Ridges, leads only to barren granophyre-spotted terrains.

The severity of this problem came to light originally with the era of the so-called 'Granite Controversy' (e.g., Read 1957). Field researchers pointed out that many types of granite seem to blend seamlessly into gneissic metamorphic rocks that may originally have come from granitic sediments. And large basaltic intrusions, like Skaergaard, and huge basaltic extrusives, like Hawaii, show little tendency to make granites, however strong the evidence of

crystal fractionation into beautifully layered sequences. In fact, there is little tendency to deviate from basalt/gabbro itself, let alone granite, or even intermediate rocks. Bowen (1947) argued that Skaergaard must have lost its volatiles too early to allow micas and hornblende to fractionate and bring the residual melt to strongly silicic compositions. But when plotted on his own Qz-Ab-Or ternary system, granites appear as minimum or near minimum melts, which he saw as residual melts (Tuttle and Bowen 1958). He also saw that any small mass of granitic melt would be trapped interstitially deep in the solidifying assembly, and invoked tectonics to knead it out for collection, but he then realized that deformational time scales are much too long relative to solidification to effect this process.

Bowen was masterful in experiment, theory, and field, and he recognized the problem but was unable to breach this obstacle. Due to Bowen's success and the obvious clear relation of phase diagrams to rocks, in terms of solving problems of chemical variations, the field has ever since been dominated by phase equilibria (e.g., Morse 1980). Once significant volumes of granitic material exist, there is almost no end of secondary processes that can be invoked to further refine and enhance it (e.g., zone refining, mixing, assimilation, partial melting, etc.). The problem is getting from basalt to intermediate rock to start with in a primary magmatic process; granites come later. The physical problem remains; the evidence is all around us.

Magmatic First Principles and reference systems

In thermodynamics, the properties of idealized, end member crystals, free of defects, dislocations, and extraneous atoms are found, and through straightforward models, the properties of real crystals are closely approximated. Phase diagrams can be calculated and chemical mass balance scenarios understood. The converse, the never-ending chore of measuring and explaining the detailed properties of every unique variety of commonplace crystal, is tantamount to explaining the origin of every magmatic body using a unique application of chemical and physical principles. Scenarios are presented that match the evidence, but the scenario itself is most often unique to that body or that researcher. These principles are unique in the sense that different ones are routinely separately applied to understand volcanic sequences, sills, dikes, small intrusions, large intrusions, and granitic versus basaltic magmas. In this context, it is not the separate specific processes that are in question; it is in the choice of the inclusion and exclusion in their stringing together to make a satisfactory scenario that is in question. Certain processes are inviolate. Heat flows from hot to cold, heavy objects sink in lighter fluid, and crystals and melt try to achieve chemical

equilibrium. It is the next set of principles that are more problematic. These principles are more aptly called 'bundled principles.' Just as all the quantum physics and chemistry behind building atoms and combining them to form a crystal need not be intimately known to understand the structure or thermodynamics of a crystal, so too is it not necessary to understand all the physics of fluids and chemistry of crystal growth to understand magmatic evolution. But certain firmly grounded aspects of these subjects must be soundly understood to decipher magmatic processes. Systems exhibiting these bundled first principles abound.

Magmatic First Principles

1. *Initial Conditions:* The exact timing sequence in the establishment of any magmatic body is generally unknown to a degree in direct proportion to the size of the body. The larger the body the less is known on how it was formed, and to simply assume 'reasonable' initial conditions, such as 'instantaneous emplacement of crystal-free magma' severely vitiates any further analysis of genesis. Yet the initial conditions, as in all physical processes, are absolutely critical to understanding the end product. Improbable or farfetched assumed initial conditions can only be connected to the given final rock record by unreasonable processes.
2. *Critical crystallinity:* The crystallinity, usually near 55 % (vol.), of maximum packing whereupon the assemblage becomes under shear a dilatant solid. The exact modal volume % strictly depends on the size distribution of crystals, and an assemblage can be designed where each succeeding size *ad infinitum* is ten times smaller such that the full volume can be occupied, but most commonly this does not occur. The magma viscosity is exceedingly sensitive to crystal content and increases essentially without limit with approach to critical crystallinity.
3. *Solidification fronts:* All magmas are surrounded by solidification fronts the thickness of which is a measure of the local cooling history and transport regime. Solidification fronts are defined by the mass contained between the solidus and liquidus, are dynamic features, relentlessly thickening inward in direct response to the thermal regime, and in phenocryst-free magma control all crystal nucleation and growth.
4. *Transport and emplacement fluxes:* The rate at which magma can be transported and emplaced to form a magmatic reservoir or eruption is limited. Small bodies can be emplaced almost instantaneously, the larger the body the longer and the more emplacements it takes to build. Large bodies cannot be

emplaced instantaneously. The longer it takes to build a body, the higher the probability of the string of emplacements being heterogeneous in composition and/or crystal load.

5. *Phenocrysts, Xenocrysts, Primocrysts*: There are broadly two types of magmas: Those laden with already large crystals, whatever their origin—cognate or not, and those not carrying any noticeable crystals of unusual size. The transport, and associated sorting and separation of entrained crystals, is critical to understanding differentiation, the textural structure of plutons, and the origin of the magma itself.
6. *Crystal size*: The typical sizes of crystals in any pluton must reflect the overall cooling time of that body. Small bodies have small crystals; unusually large crystals relative to cooling time reflect injection of slurries. Annealing, which is proportional to cooling time and body size, exacerbates initial characteristics.
7. *Layering and crystal sorting*: Any slurry of crystals, given the chance in avalanching or other shear flows, will sort itself by size, density, or shape into well-formed layers. Layering reflects injection of slurries over rhythmic crystallization.
8. *Compaction*: Compaction and interstitial melt eviction works best in thick sequences of poorly welded mush or slurries possessing a marked basal discontinuity in porosity or crystallinity and weakly, or perhaps not at all, in basal solidification fronts grown under equilibrium conditions where crystals weld together in a 3D network as they grow.
9. *Thermal convection*: Normally only occurs in sheets of superheated magma, not otherwise.
10. *Magmatic processes are physical*: Magmatic processes are characteristically physical processes buttressed by chemistry, not the reverse.

Magmatic reference systems

Perhaps the greatest mystery in all of igneous petrology is knowing explicitly from beginning to end exactly how any given batch of magma, regardless of size, has crystallized to achieve its final spatial state. Bodies of subterranean magma in their natural habitat have proven incredibly elusive to find, let alone examine. Postmortem or post-solidification studies after protracted crystallization foster a certain style or culture of analysis (e.g., dynamic implications of certain textures) that has proven difficult to critically analyze without a magmatic firmament or reference system to guide a broader analysis. Many scenarios, however unlikely and unconvincing, satisfying chemical evidence can be presented. Yet enough is now known of certain distinctive magmatic systems that an end-member

reference system can be constructed to aid in understanding most all magmas.

Small magmatic bodies, as in thin (<10 m) dikes and sills, are commonly fine grained, featureless, and deemed uninteresting. Left unstudied, just how uniform is the texture and composition is generally unknown. That some sheets as thin as a few cm can be hundreds of meters long suggests an injection rate, to avoid solidification, approaching the propagation rate of a shear wave. To travel this fast, the magma viscosity must be as low as possible: the magma must be near its liquidus and free of large crystals. The initial conditions associated with the subsequent solidification are not in question nor is the history of differentiation. The body cools too quickly relative to crystal nucleation and growth to allow any collective physical segregation of the crystals from the melt in which they were grown. It is only as magmatic bodies become larger that the spatial variation in texture and composition becomes highly varied, almost in proportion to size, and the history of solidification becomes problematic.

And here enters an enigma: The smallest bodies achieve rigidity too quickly to allow any subsequent injection, but with increasing size, as in a dike or sill, once the cooling time becomes commensurate with the reinjection time, the body may still be partially molten and weak in its center, inviting a subsequent injection. For smaller bodies, perhaps up to 200 m, this process may repeat itself many times, and because of successive high temperature chilling (so-called ‘hot chills’), the rock record may be distinctly noticeable. This has been shown particularly well in the Shiant Sill by Gibb and Henderson (1989), and has recently been delineated in some detail using CSD variations for the Beacon Sill (Zieg and Marsh 2012). But with increasing body size, since the cooling time increases with the square of the body size, the reinjection times—perhaps only weakly dependent on body size—become short relative to the solidification time. Injections take place in high temperature mushes without producing any final chill. These events thus become exceedingly difficult to recognize in the rock record, especially with subsequent annealing. And as bodies become unusually large, kilometers in thickness, although the number of injections is surely large, standard evidence of reinjection is scant, yet it does exist. The enigma is thus: In the smallest of bodies, the initial conditions are perfectly clear, but the outcome is uninteresting. As the bodies become large and interesting, the initial conditions become obscure, yet remain critically indispensable to deciphering the magmatic history and cannot simply be casually assumed. Reasonable bounds can be placed on the initial conditions of most bodies using magmatic reference systems. The two most accessible magmatic systems with the clearest, unequivocal initial

conditions are the Sudbury impact melt sheet and the Hawaiian lava lakes.

Sudbury and the null hypothesis

In less than 5 min, a massive ($\sim 50,000 \text{ km}^3$) crystal-free, superheated ($\sim 1,700 \text{ }^\circ\text{C}$) magma was formed at Sudbury 1.85 Ga ago. With a depth of 3 km, a diameter of $\sim 200 \text{ km}$ and the aspect ratio of a compact disc it has, above all, relatively speaking, precisely known initial conditions: Instantaneously emplaced and crystal free. These are exactly the initial conditions commonly assumed for many large intrusions (e.g., Wager and Brown 1968). These are perfect initial conditions for a laboratory experiment. What happened? Although as in any large body, there are endless intriguing detailed processes, in terms of magmatic differentiation, the short answer is relatively nothing.

The impactor landed in heterogeneous largely granitic crust laced with dolerite dikes and sills and perhaps small, layered gabbroic intrusions (e.g., Zieg and Marsh 2005). The initial transient cavity reaching the 30 km deep Moho developed in $\sim 2 \text{ min}$ and with a diameter of $\sim 90 \text{ km}$ relaxed within $\sim 5 \text{ min}$ to a 200–250 km diameter crater. Everything associated with impacts of this size is breccias produced by the intense ~ 2 megabar shock waves; atomic bonds in common minerals are broken and the crystalline structures scrambled (e.g., French 1998); the bolide itself was vaporized. The initial melt was the equivalent of a high temperature molten breccia resembling a viscous emulsion, separating in days to months under the action of gravity into two superheated layers: an upper $\sim 2 \text{ km}$ granitic layer and a lower $\sim 1 \text{ km}$ basaltic (norite) layer; all covered by $\sim 2 \text{ km}$ of fall-back breccia (Fig. 1). The up and down streaming of blobs of granitic and basaltic melts of all sizes gave ample opportunity for intimate chemical exchange. Vigorous thermal convection commenced in each layer, dissipating $\sim 500 \text{ }^\circ\text{C}$ of superheat down to the liquidus in tens to hundreds of years and homogenizing each layer locally on scales of ten km. Crystallization then commenced within upper and lower solidification fronts moving inward and meeting in the center after $\sim 80,000$ years. Because of the unusually high temperature of the melt sheet, the lower contact temperature was initially at the norite solidus, which pinned the cool side of the solidification front at the contact until the liquidus reached the center. This produced an unusually thick ($\sim 1 \text{ km}$) solidification front extending throughout the norite, giving ample opportunity for crystal growth, compaction, and differentiation. Yet, the norite itself is essentially uniform and featureless (Fig. 2), a characteristic that perplexed all petrologists dedicated to Bowen's prevailing model of central crystallization and progressive crystal settling. Hess

(1960, p. 173) noted, 'The mafic portion of this intrusive shows surprisingly little differentiation in spite of its size and hence long time available for differentiation. Apparently no crystal sorting has occurred.' Wager and Brown (1968, p. 522) similarly puzzled over the general uniformity of the Sudbury in their global survey of large layered intrusions: 'The general impression gained from the descriptions of the norite intrusion and examination of the rock types is that it is a relatively uniform norite with little conspicuous layering.' And the upper, granitic unit as a massive granophyre has lost its textural identity, perhaps due to superheating, while the thin central unit or sandwich horizon, the Transition Zone (quartz gabbro), is a dense collection of materials rising from the crater floor, falling from the roof, and perhaps slightly squeezed out from the underlying norite.

The missing piece of the puzzle for a long time was no knowledge of the special initial conditions of Sudbury's formation by impact. And especially since these conditions—Instantaneous Emplacement and Crystal Free—are exactly what became the standard assumed initial conditions for major magmatic bodies like Skaergaard, Stillwater, and Bushveld. After all, how important could an accurate knowledge of initial conditions possibly be? In short, initial conditions are everything. In fact, the converse is much more accurate to work from: The final states of magmatic bodies are, by and large, essentially what they start out as. This gives rise to the *Null Hypothesis*: Sheet-like magmatic bodies, regardless of size, born uniform and crystal free solidify to near uniform homogenous bodies, free of layering, and any strong effects of large-scale differentiation. This is obvious as discussed above for small dikes and sills, and it also appears so for large sheet-like bodies as exemplified by Sudbury. Many large uniform sills, like the $\sim 350 \text{ m}$ Penepine Sill, also attest to this. But as bodies get large, the time to build them becomes large and an accurate knowledge of the initial conditions of formation becomes obscure. The converse is especially interesting: Nonuniform, exotically layered, and chemically differentiated bodies come from noncrystal-free and, most likely, noninstantaneously emplaced magma. The only unequivocal, accessible bodies of this type are the Hawaiian lava lakes.

Hawaiian lava lakes

These tholeiitic to picritic lava lakes are remarkable for their known initial conditions of formation from crystal-rich magma. They are thin (14–125 m) strongly differentiated, layered sheets that have been extensively studied by direct sampling during the course of filling and solidification (e.g., Wright and Okamura 1977). They furnish an invaluable and indisputable connection between volcanism

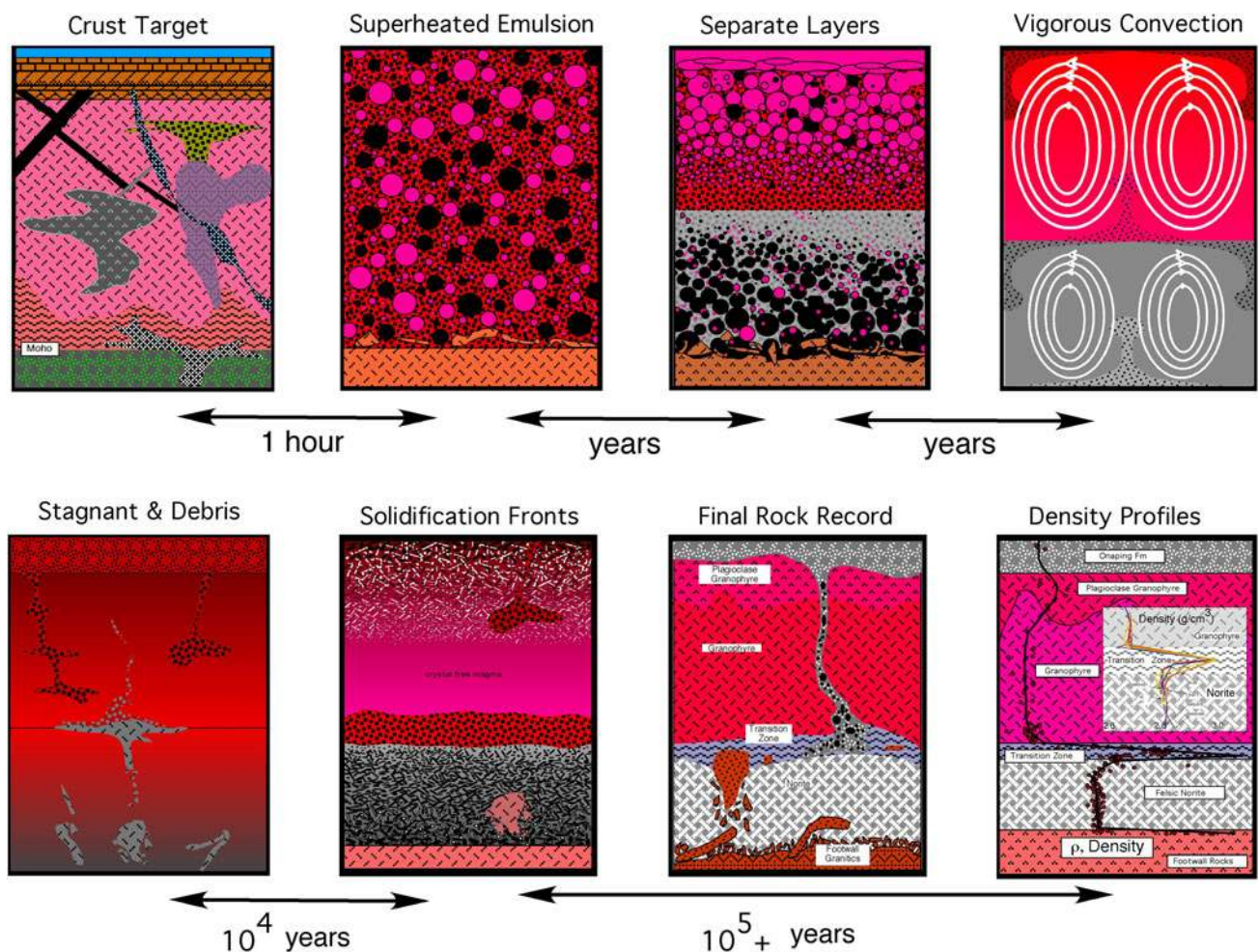


Fig. 1 The magmatic history of the Sudbury impact melt sheet; beginning with the impact into heterogeneous country rock (*upper left*) and continuing with separation of a viscous emulsion into upper granitic and lower noritic layers that were strongly mixed through

vigorous thermal convection, inward propagation of upper and lower solidification fronts, and rock debris rising from the crater floor and falling from the upper mat of fallback breccia. After Zieg and Marsh (2005)

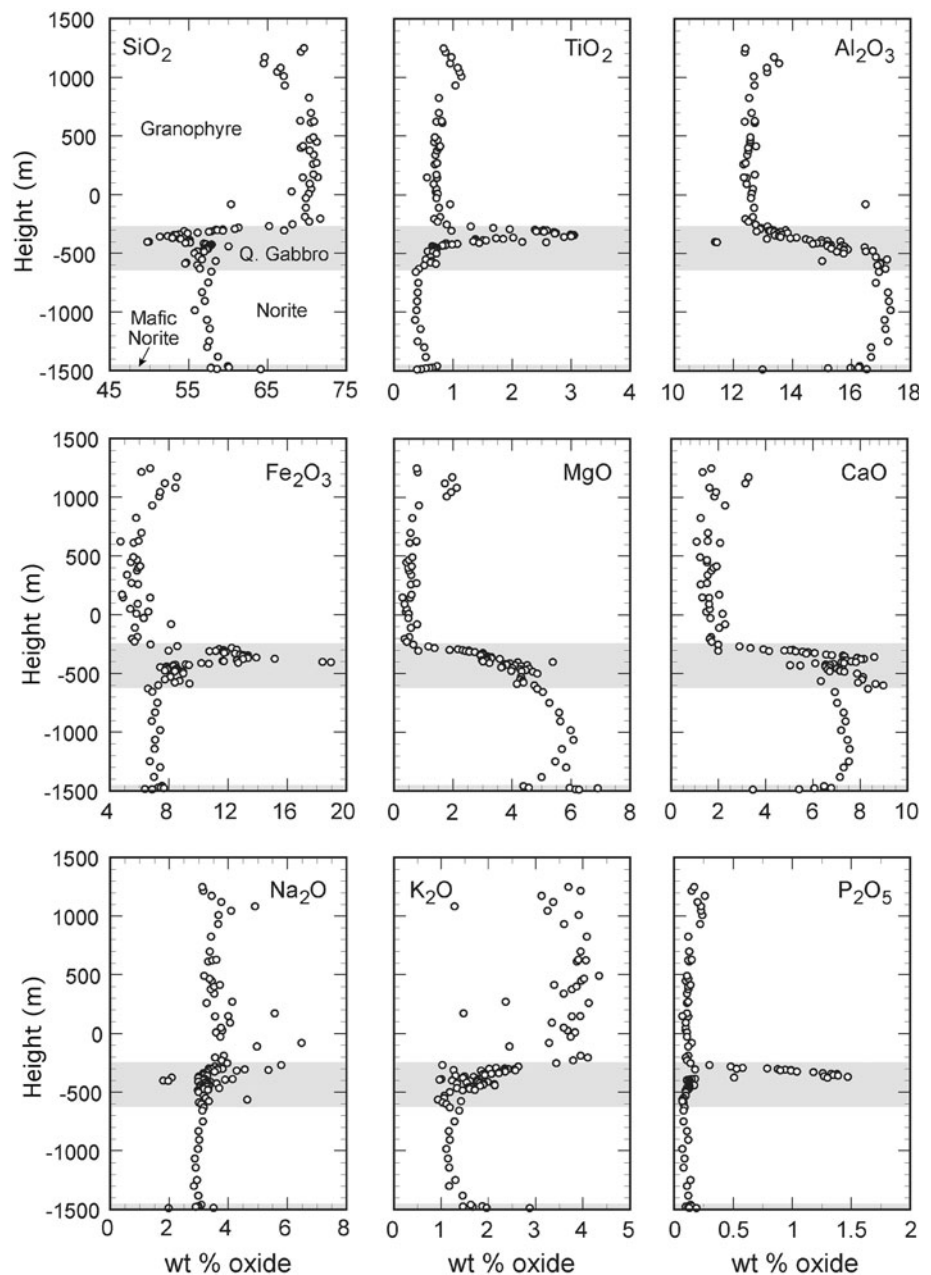
and the vast family of bodies known as differentiated sills (e.g., Upton and Wadsworth 1967; Gunn 1966; Gunn and Warren 1962) that have been so aptly generalized by Simkin 1967).

The 1959 eruption from Kilauea Iki, for example, took place over about a month (November 14 to December 20, 1959) with the main filling of the lava lake taking place over 5 days (16–21 November) at a peak flux of 380,000 m³/h (3.34 km³/year) culminating in a maximum volume of 44.3 million m³ (Fig. 3; Richter et al. 1970). The subsequent drilling of this and especially Makaopuhi lava lake (Wright and Okamura 1977) gave the impetus for the establishment of the solidification front concept (Marsh 1981, 1996). Much of the incoming lava for Kilauea Iki was picritic with MgO as high as 25 wt% and forsteritic olivine crystals as large as 2 mm. The lake level fluctuated in concert with filling and drain-back, leaving a ~125 m thick sheet with a bathtub-berm ring 10–20 m above the

final lake level. Once the lake stabilized after filling, a crust or solidification front began to grow inward from all contacts capturing some of the olivine in the upper crust while phenocrysts deeper in the lake escaped to form a cumulate sequence on the floor. The net result is a classic S-shaped sedimentation sequence of MgO variation reflecting the modal abundance (up to 37 vol%) of olivine (Fig. 3) in the incoming lava. Temporal changes in filling flux delivered an uneven modal amount of olivine, which modulates the initial spatial distribution of bulk composition and phenocryst content (e.g., Murata and Richter 1966) and any subsequent layering reflects this temporal input sequence.

Beyond the obvious dynamic sequence of filling with olivine phenocryst-xenocryst-laden lava leading to a layered magmatic body, even more basic is the monitoring of the full cooling and solidification history along with evolution of melt chemistry due to fractionation of the olivine, however heterogeneous (see below) in composition it may

Fig. 2 Sudbury bulk rock major oxide compositions plotted against height (from Zieg and Marsh 2005)



be (Fig. 4; Richter and Moore 1966; Helz 1980, 1987a, b; Helz and Thornber 1987). That is, when the bulk composition is examined on a standard major oxide subtraction diagram the correlations are clean and clear in spite of the mineral assemblage being essentially debris. And, as carefully as Jaeger (e.g., 1968) treated and explained the cooling history of sheet-like bodies, there has been widespread confusion over exactly how sheet-like bodies cool and crystallize. Ever since Pirsson (1905) stated for Shonkin Sag that ‘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.’ there has been the general impression that cooling is much more rapid through the roofs of intrusions and, similarly, crystallization is much more extensive, generating a heavy flux of crystals to the floor, forming massive cumulates. Cooling is, indeed, often slightly more rapid from the roofs of sheets, not because they are more exposed to Earth’s surface, but because it is easier for hydrothermal systems to set up there. This is seen for the lava lakes (Fig. 5) due to water from rainfalls more effectively cooling the upper crust by hydrothermal convection as demonstrated in detail for Alae lava lake by Peck et al. (1977). But the attending

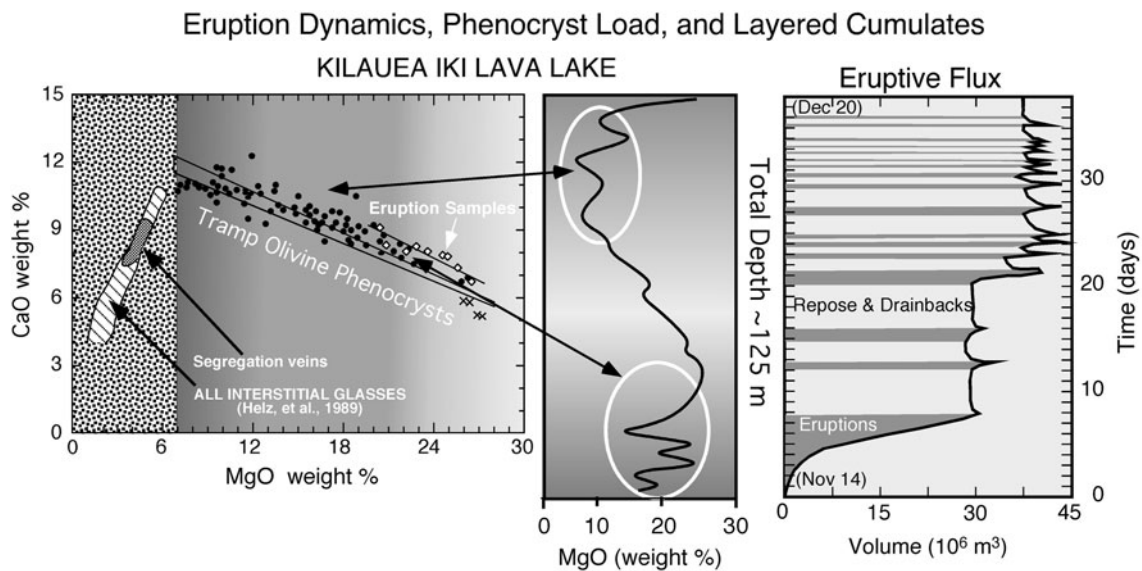


Fig. 3 The composition of the lavas forming Kilauea Iki Lava Lake in 1959. Left, shows the overall bulk composition in terms of CaO and MgO exhibiting the large variation in MgO due to the uneven abundance of primocrysts of olivine and the olivine fractionation control lines. Center, the spatial or stratigraphic S-shaped distribution

crystallization does not send a stream of crystals to the floor; it simply produces a more rapidly growing, thicker upper solidification front, which displaces the thermal sandwich horizon deeper into the body. Perhaps the only way to raise the sandwich horizon above the center of the body is to pour slurries of large crystals on the floor or detach the upper solidification front. Thus, the actual position of the petrologic ‘sandwich horizon’ may be, in the case of systems supplied with abundant crystals in the input emplacements, as in the lava lakes, higher in the system than the thermal center. This reflects the rapid buildup by sedimentation of crystals on the floor, producing a deceptively thick lower solidification front that is often mistaken for crystals produced at the roof by rapid cooling. Any body where the sandwich horizon can be credibly located above the center is a clear candidate for slurry emplacements. The Palisades Sill and Skaergaard are likely examples of this phenomenon, which is also well exhibited in prehistoric Makaopuhi lava lake (Evans and Moore 1967, 1968). And with slurry inputs and the possible consequent compaction, the thermal regime at the floor can be distinct from that at the roof.

Another effect of dumping loads of incoming primocrysts on the floor is the ongoing attempt of this mass to reach equilibrium with the interstitial trapped melt. The packing of the cumulate assemblage, prior to any compaction, is ~50 vol% and the cumulate olivine, regardless of its initial composition, will within months equilibrate with this melt; giving up magnesia in exchange for iron (Moore and Evans 1967). The final assemblage when used

of olivine as reflected by the bulk rock MgO content. Right, the volumetric buildup over time reflecting the variation in eruptive flux with time over the period of Lava Lake formation, showing the strong variations, which are directly related to the entrainment and transport of olivine and are similarly reflected in the stratigraphic profile

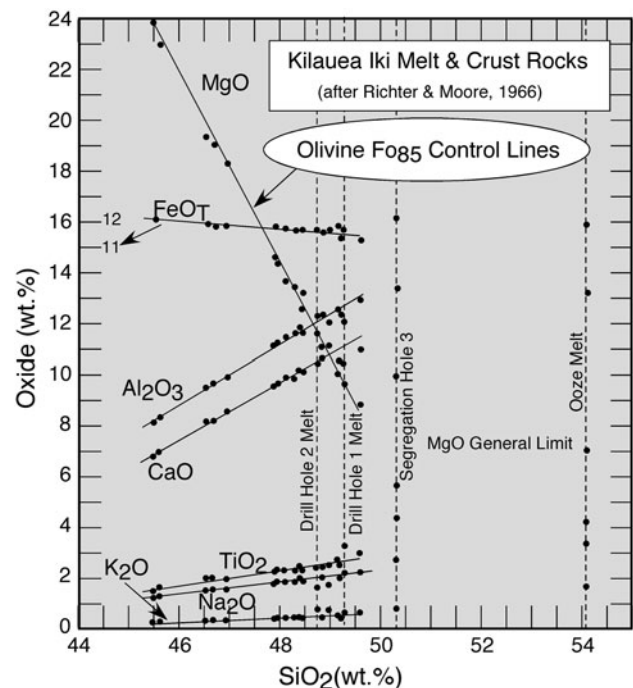
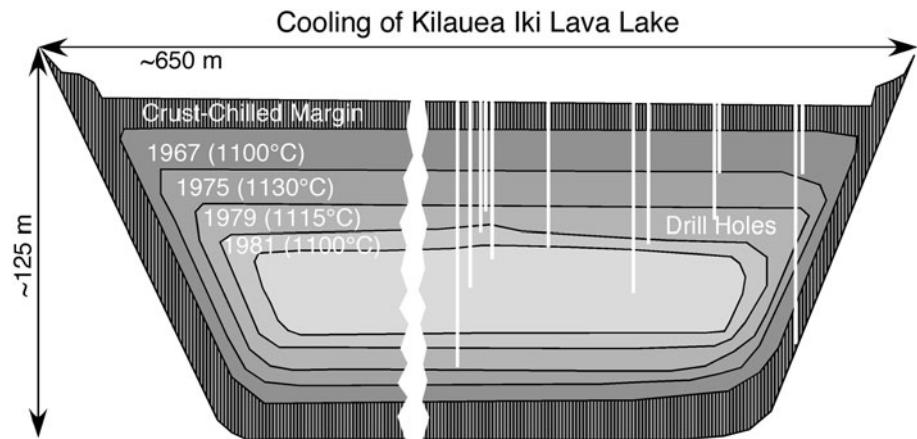


Fig. 4 Kilauea Iki bulk rock chemical fractionation diagram exhibiting the strong control of olivine fractionation on the bulk composition. The overall effective controlling olivine composition is Fo_{85} even though the actual crystal compositions are highly varied (from Richter and Moore 1966). All compositions to the left of 50 % SiO_2 (i.e., richer in MgO than ~8 %) are either from the infilling eruptive or the deepest part of the two indicated drill holes. The two more enriched compositions are from interstitial melts or oozes collected by drilling within the crust or solidification front at crystallinities greater than ~55 vol%

Fig. 5 The thermal history of Kilauea Iki Lava Lake showing the inward migration of isotherms and the position of drill holes. The final horizon of magma is displaced downward due to the effect of hydrothermal cooling in the upper crust from rainfall, and be reminded that the true aspect ratio is that of a slice of bread



to gauge the initial ‘primary’ magma composition will give misleading information reflecting a melt more evolved than that which produced the initial crystals.

The locations of vents that supply magma to intrusions are commonly assumed to be in the bases of the body, but the Hawaiian lava lakes are commonly fed from overland lavas or vents in the sides or near pit craters. The vent for Kilauea Iki was in the subsidence crater wall where the stress field due to topography attracts propagating dikes, which then vent. The lake was thus fed from the side, which provides an opportunity for waves of magma to be spread laterally and for piles of cumulate crystals to avalanche into the greater part of the system. In multiply dispersed solid–liquid systems (see below), any shear motion whatsoever will automatically begin ordering the solids into layers according to size and density. In a crystal-laden fluid ascending in conduits, fissures, or dikes, the crystals become spatially sorted with the largest and heaviest particles in the flow center, forming a trailing central plug or tongue-like flow, encased in crystal-free, highly fractionated, melt (Simkin 1967). Upon delivery to the final reservoir, the incoming mix of crystals and melt is thus already ordered, although the eruptive and filling process may make this difficult to discern. As the lake filled, the initial sub-aerial vent became flooded and sub-surface fountaining initiated surface gravity waves, which propagated across the lake repeatedly disrupting nascent crust formation (Richter et al. 1970). In sum, regardless of the mode of filling of a magmatic body, abundant physical processes can be expected to operate to manipulate the magma to produce an ordered rock record that will anneal into an even cleaner deceptive rock record (e.g., Boudreau 1994).

Given these two end-member magmatic systems, Sudbury and Hawaiian lava lakes, there are any number of combined states that exist between them. Regardless of size, from initially crystal-free magma comes a featureless rock record; from crystal-laden slurries comes highly

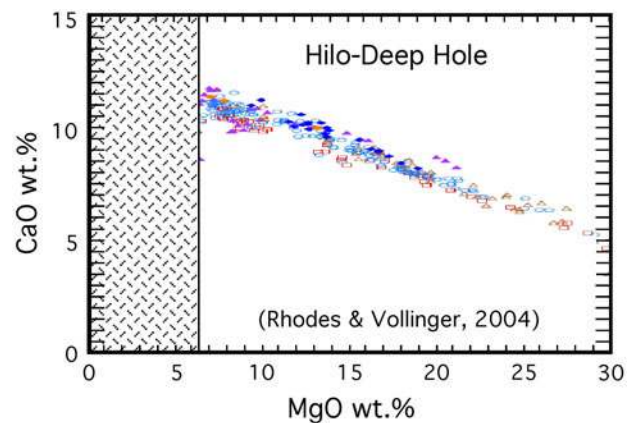


Fig. 6 The variation in CaO and MgO in lavas encountered in the 3 km Hilo, Hawaii drill hole covering a time span of ~500,000 years (after Rhodes and Vollinger 2004). The large range in MgO content reflects a similar range in modal olivine. No lavas are found with MgO less than ~7 wt% MgO, which is the residual composition after loss of all primocrysts of olivine

differentiated, layered sequences. The vast possibilities can be appreciated by considering the building of a magma chamber (Fig. 6).

Building a magma chamber

Basic conditions

The question can now be asked: How does one build a magmatic body? The simple answer is that, given an order for a certain size or volume (V_T) of body, all that is needed is a certain flux (Q) of magma sustained for a period of time (t) equaling the needed volume (i.e., $V_T = Q \times t$). Herein, this simple equation presents the basic problem. Both V and Q are functions of time, that is, $V(t)$ and $Q(t)$; and since $Q(t)$ is supplied by Nature and cannot be prescribed, in approaching the final volume V_T , $V(t)$, as already seen for Kilauea Iki, may be a highly variable time

series. Some idea of the nature of this time series can be appreciated by the periodic nature of eruptions leading to building a volcano. And in a real way, this is akin to building a magma chamber. Eruptions or deliveries proceed perhaps in a rhythm with fortnightly tides interspersed with periods of repose that may last tens to hundreds of years. And all the while, regardless of the building process, solidification relentlessly proceeds, processing magma from the outside inward. The final product is thus a combination of the nature of the deliveries, both in volume and character (e.g., crystal content), and the rate and pattern of solidification. But the specific functional features of the delivery sequence are unknown. When filling started, how it proceeded and how long it lasted are unknown except for the smallest bodies. Yet the broad characteristics of the most probable patterns of filling, in both flux rate and character, can be ascertained, and these give clear insights into how large magmatic bodies are built.

Nature of deliveries

The only readily available record on the nature of the inputs that go to building magma chambers of massive magmatic bodies is in the output of large, long-lived active volcanoes, or volcanic centers like Hawaii. In the past 1 Ma, about 10^6 km³ of lava have been erupted to form the present Hawaiian volcanic center. The recent eruptive record coupled with the 3.1 km deep drill hole near Hilo (e.g., Stolper et al. 2009) furnishes a 500,000-year record of the character of possible chamber inputs. The chemical diversity of the lavas encountered in drilling is large, as is illustrated by Fig. 8a (after Rhodes and Vollinger 2004); the variation in MgO alone spans nearly 30 wt%. This style of variation is typical for Hawaii not only for the long-term eruptive record (Peterson and Moore 1987) but also for recent voluminous eruptions as already mentioned in building Kilauea Iki lava lake. A distinctive example of this universal variation is seen in the modal abundances of olivine (see Fig. 8b) found in submarine Mauna Loa lavas by Garcia et al. (1995). The modal abundance is a direct reflection of the power (transport flux) to entrain olivine, which largely dictates the bulk MgO composition. At the lower end of this correlation, at 0 vol% olivine, are simple tholeiitic lavas with ~7 wt% MgO, and at the upper end at the maximum load of olivine (~50 vol%) are the most magnesian lavas, including komatiites. The importance of this deceptively simple diagram in linking the bulk composition to the dynamics of ascent and entrainment cannot be overemphasized.

Thus, the large variation in MgO does not represent actual melt compositions, but generally reflects the overall large crystal content, mainly olivine, most of which are so-called ‘Tramp Crystals’ (Wright and Fiske 1971; Wright

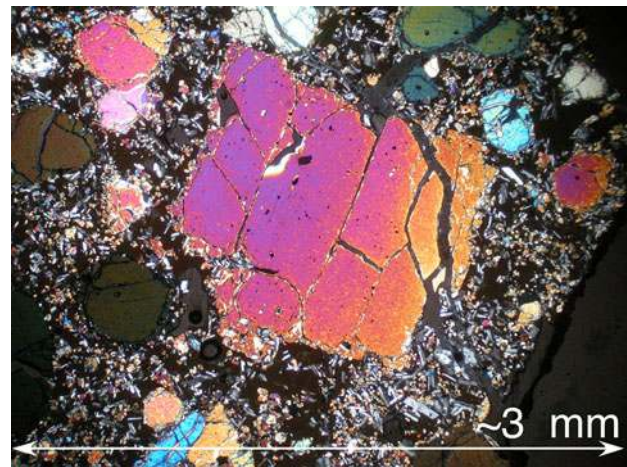


Fig. 7 Photomicrograph of large olivine primocrysts in Kilauea Iki Lava Lake. The large fragmentary and unzoned character of these highly forsteritic crystals are diagnostic of crystals entrained from within or around the underlying magmatic mush column

1971). These are highly forsteritic (often Fo_{85–92}) olivine crystals entrained during the eruption process. That they have been entrained as extraneous crystals is obvious in thin section (see Fig. 7). The load and size of these crystals are proportional to the eruptive flux and are similar to the bed-load carried by rivers in flood stage (Murata and Richter 1966; H. A. Einstein, personal communication 1974). An experimental system fluid dynamically scaled to simulate similar magmatic slurries reproduces the general observed relations between flux and particle abundance, size, and density (Flanagan-Brown and Marsh, in prep).

Olivine slurries

These high-Mg lavas are thus physical slurries of entrained crystals the compositions of which vary widely at all scales down to that of a single thin section (e.g., Maaloe and Hansen 1982; Maaloe et al. 1988). That is, adjacent crystals may have markedly different compositions. Yet when any single eruption is plotted on a CaO-MgO variation diagram, well-defined olivine-control lines appear denoting the effective fractionating olivine to be of a composition of Fo_{87.5}. Curious enough, this same fictive composition appears worldwide wherever high modal amounts of olivine are erupted. A bag of olivine beach sand from Rum, for example, gives a similar overall composition (R. Hunter, personal communication, 1995). This particular feature of magmas to carry a heterogeneous load of crystals is also not restricted to olivine-rich magmas as has been repeatedly shown down to the smallest scales by Davidson and co-workers (e.g., Davidson et al. 2005a, b, 2007) who aptly refer to this as the ‘Crystal Cargo.’

There are several aspects of the presence of these olivine crystals that are distinctive to understanding the building of

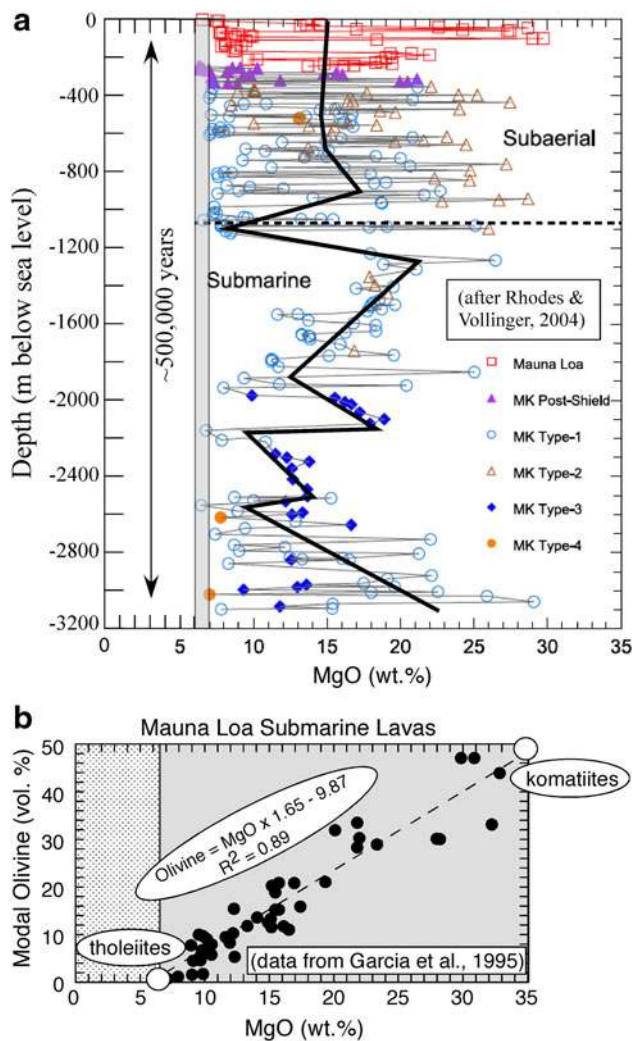


Fig. 8 **a** The variation of MgO in lavas encountered in the 3 km Hilo drill hole covering $\sim 500,000$ years of eruptive history (after Rhodes and Vollinger 2004). The strong variation in MgO reflects the modal abundance of primocrystic olivine, which varies in accord to the magnitude of the eruptive flux. There are no lavas with MgO contents less than ~ 7 wt% MgO, which is the residual composition after loss of all olivine. **b** The modal variation of olivine in submarine Mauna Loa lavas as a direct indication of the bulk MgO content (lower axis). This exemplifies the dynamic link between the eruptive power of the magma to entrain olivine and the bulk magma composition. The slowest moving, most passive magmas contain the lowest amount of olivine and are tholeiitic; those with the most power of entrainment contain the highest modal amount of olivine and approach komatiites in composition

magmatic systems. First, given the rapid rate at which olivine can equilibrate by diffusion with a melt, these crystals, depending on size, have probably been in this melt for no more than a few months. Second, the most obvious source of these crystals is the peridotitic mantle wall rock lining the underlying magmatic mush column. If so, where then are all the other peridotitic phases—orthopyroxene, clinopyroxene, and/or spinel, plagioclase, and garnet? The

simple answer is that, individually being of lesser modal mass and more out of equilibrium than olivine, they have all dissolved in the melt. Third, if all these crystals are stripped from the magma, the lava composition is routinely reduced to ~ 7 wt% MgO. All chemical variation diagrams end abruptly at this end point, and this point is reached as soon as the slurry slows enough in ascent to not be able to transport the load of crystals. In fact, no composition more siliceous than 51.5 wt% SiO₂ and less magnesian than 6.8 wt% has ever been erupted from the Kilauea summit. This is also abundantly clear in the Hilo deep hole especially when the bulk MgO content is plotted against depth; in all the lavas penetrated covering 500,000 years of eruptions, none have less than ~ 7 wt% MgO (see Fig. 8). This feature at the Kilauea summit particularly impressed Murata and Richter (1966): ‘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.’ This composition defines the leading edge of the solidification front once the magma has come to rest and lost its load of crystals and thus reflects the existence of a passive batch of magma staged for eruption (Marsh 2007).

The dynamic interface between physics and chemistry

This general characteristic bulk composition (~ 7 wt% MgO) is seen worldwide, even at ocean ridges. The details of the bulk composition certainly vary considerably, depending on the exact nature of the mantle source rock, but in olivine-dominated systems, this broad composition predominates. In the lava lakes, where active ongoing crystallization has been studied in detail through drilling, this composition is found to define the leading edge of the basaltic solidification front. All compositions more refined than this reside as interstitial melt within a physically active unit, the solidification front, and are inaccessible to differentiation by normal crystal fractionation. As the magma comes to rest in ponding, it is stripped of the large crystals; the crystals drop under gravity to the floor. This is *Punctuated Differentiation*, and further crystallization proceeds solely within the enveloping solidification fronts, the leading edge of which is characterized by tiny, newly nucleated crystals that are unable to escape the advancing front. Once they are large enough individually to settle, they are so far back in the solidification front that crystallinity has advanced to the point where individual settling is impossible. Should small crystals escape the leading edge of the front, they enter hotter magma and are re-sorbed, thusly stabilizing the front edge (Jaeger and Joplin 1955; Mangan and Marsh 1992). Out in the middle of the magmatic body, where settling is possible, there are no crystals. Bowen’s usual reaction series resides and thus

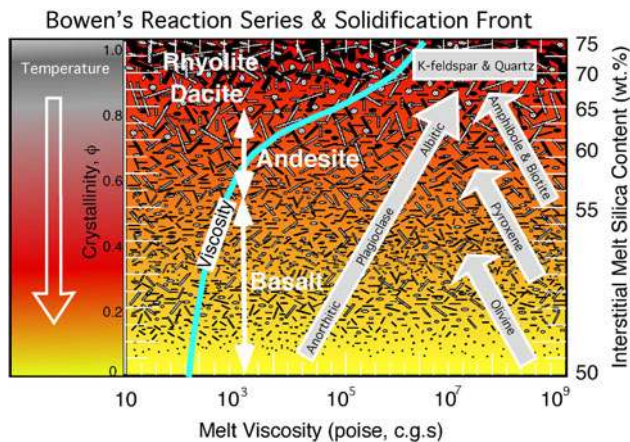


Fig. 9 Bowen's reaction series in relation to its spatial position within a magmatic solidification front. Temperature increases downward whereas melt composition becomes increasingly silicic upwards in the front. Significant increase in silica (~ 5 wt%) only occurs after ~ 50 vol% crystallization (for a tholeiitic basalt) at which point the crystalline matrix is at maximum packing, is welded together, and is unruptible

takes place within the solidification front, which encapsulates and stifles the 'liquid line of descent,' truncating the approach to granitic, continent-like compositions (see Fig. 9).

Once a magmatic body reaches its final resting place, the possibilities of any significant differentiation vanish. Differentiation is intimately linked with the dynamics of transport. Marked differentiation occurs by the repeated combination of rapid transport and local ponding; rapid transport allows entrainment of crystals—essentially any crystals with which the melt can react—and cessation of transport or ponding allows the magma to drop the load of crystals, producing rapid fractionation. The ultimate origin of the entrained crystals, in terms of producing fractionation, is almost immaterial. The two ready sources of crystals are the local wall rock and earlier-formed disrupted solidification fronts. The walls of magmatic plumbing systems are structurally highly unstable on many scales and in collapsing disrupt growing solidification fronts and also dump primitive wall rock materials into the magmatic system. In large, long used magmatic mush columns, these materials are bathed in magma, and the magma and crystal, regardless of origin, are continually reacting in an attempt to reach equilibrium. The span of compositions up to ~ 30 wt% MgO in systems like Hawaii reflects the maximum addition of roughly 40–50 vol% forsteritic olivine, which approaches the maximum amount, in terms of critical crystallinity and packing, that can be transported without stalling the magma.

That stalled or ponded magmas do not undergo significant differentiation is best exemplified by the vast collections of dolerite dikes found in ophiolite sequences. These

dikes form in passive response to sudden spreading of the rift overlying the sill-like axial magma chambers beneath ocean ridges. Although in the long term, spreading is steady, in the short term it is episodic and each event splits a previous dike, opens the underlying solidification front, and extracts an aliquot of magma from the magma chamber. This random sampling gives a compositional snapshot view of the chemical and physical state of the chamber. That these dikes are notoriously featureless and of a uniform composition approaching ~ 7 wt% MgO indicates the chemically stagnant state of the underlying magma chamber. This extraction takes place rapidly, and the flux rate may be large, but the most easily accessible melt is that which is most fluid near and beyond the leading edge of the solidification front, near the liquidus. If this high flux rate was sustained long enough magma containing an increasing concentration of large crystals might appear—the ones that do appear enter the chamber base to form cumulate layers. But at ridges, sustained eruptions to reach this state are unlikely. At large basaltic volcanic centers high fluxes must be sustained simply to produce an eruption, and the root of the eruption, as reflected in the seismic record, may be many kms deep, so there is ample opportunity for crystal entrainment (e.g., Wright and Klein 2013). The fact that the MgO content of Hawaiian lavas can vary so strongly in successive lavas reflects a similar strong variation in eruptive flux. Low flux or small volume eruptions evict nearby ponded magma, which has lost its crystal load and is akin to ocean ridge eruptions. High-flux eruptions are not only voluminous but carry large quantities of entrained crystals.

Something of the actual quantitative linkage between eruptive flux, modal abundance, and crystal composition can be seen in studies of the Kilauea summit eruptive history (Garcia et al. 2003) and from nearby Pu'u 'O 'o (e.g., Marske et al. 2008). The eruptions at Kilauea show a pronounced correlation between flux, bulk MgO content and modal olivine as shown in Fig. 10. The highest estimated eruptive fluxes (~ 0.10 km³/year) along with those of Kilauea Iki (3.3 km³/year), yielding the high modal abundances of olivine, give some indication of the actual fluxes involved in transporting and erupting the high-MgO lavas sampled in the deep drill hole (Fig. 8). Figure 11 shows the olivine core compositions for these Kilauea summit eruptions as a function of the bulk Mg number ($Mg\# = (Mg/(Mg + Fe^{2+}))$). The stippled band is the equilibrium composition expected for this bulk composition as based on Roeder and Emslie's (1970) method. The bars for each eruption show the full range of reported olivine compositions for phenocrysts, xenocrysts, and micro-phenocrysts. The spread in compositions, reflecting less and less overall equilibrium, increases with increasing Mg#, which is a reflection itself of the abundance of olivine

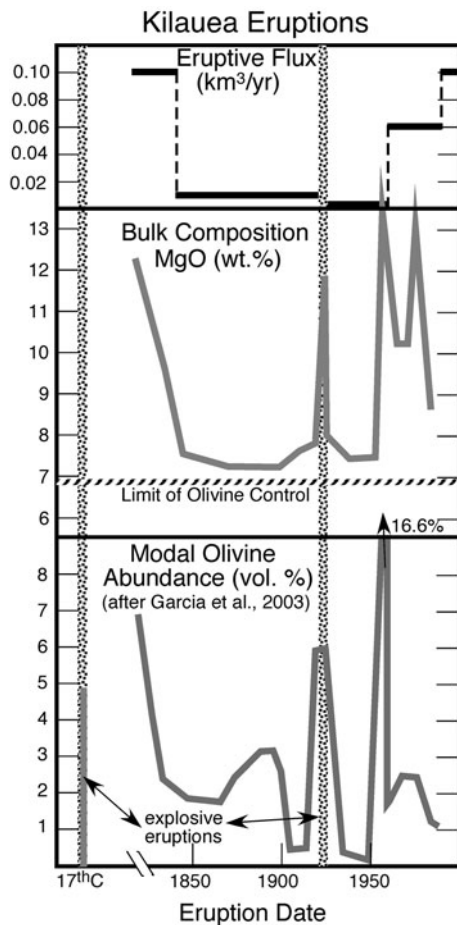


Fig. 10 An historical record of the summit eruptions of Kilauea volcano as determined by Garcia et al. (2003) showing the correspondence over time (*lower axis*) between modal olivine content (*lower*), bulk composition (*middle*) and eruptive flux. Notice the correlation between eruptive flux and olivine abundance

in the lava. For any single eruption, the lava commonly contains new crystals, micro-phenocrysts, that are in equilibrium along with phenocrysts of a wide (6–8 mol%) range of compositions the width of which increases with Mg#. The xenocrysts are identified as showing signs of deformation or resorption. The average size of crystals, as might be expected, also increases with the magnitude of the eruptive flux (Murata and Richter 1966), which is also apparent in the Kilauea Iki eruptive sequence. Broadly similar overall olivine characteristics are also found in kimberlites (Arndt et al. 2010).

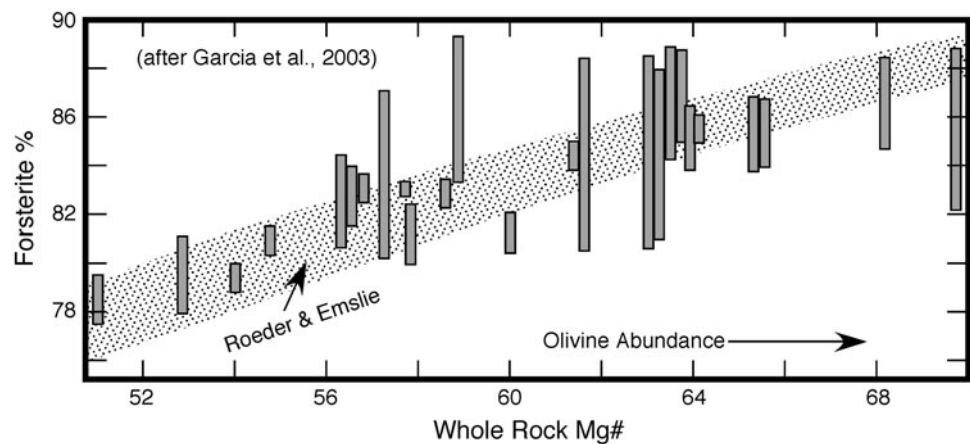
Garcia et al. (2003) also find that all lavas with Mg# > 53.5 contain two populations of crystals: a high-Fo (86–89) group that shows normal zoning and a lower Fo-group (83–85) that shows reverse zoning near the rims. Given the rapid response of olivine to its environment (e.g., Gaetani and Watson 2000), the approach to equilibrium reflects the age of the crystal with small newly formed crystals most closely reflecting the present, local melt

composition. That the more Fo-rich crystals are normally zoned more than likely identifies this group as the larger mass of crystals that are controlling the ongoing change in melt composition. These crystals could be identified with an earlier, more Mg-rich phase of the host magma, or they may simply be a common population of crystals found throughout the plumbing system, the ultimate provenance of which could be varied and complex. The key general feature here is that any given magma if examined during transport will likely contain a batch of crystals only weakly in equilibrium with the present magma. Once transport is halted, this batch of crystals can quickly separate from the magma (to be perhaps entrained by a later magma), and ongoing crystallization can produce an equilibrium assemblage; in most tholeiitic magmas, this happens when MgO falls to ~6.8 % and plagioclase and clinopyroxene join olivine in the crystallization sequence within the enveloping solidification fronts.

Carrier-fluid magma concept

Crystal-rich magmatic systems that have cooled fast enough, regardless of size, to capture a natural magmatic state show a slurry-based magma with crystals of diverse provenance. And, even when large crystals seem to be in bulk chemical equilibrium with the melt, adjoining crystals may have a distinct isotopic identity. The load of crystals carried, and the melt carrying the crystals, defines magma. Hence, instead of thinking of magmas that always contain indigenous crystals that become modified by additions and losses to the magma, it is perhaps more accurate to think of a magmatic system as consisting of a silicate melt or fluid that begins life free of crystals and gains and loses crystals as the local physical environment allows. The ascending fluid, Carrier Magma, is bathed in crystals, some of which may have been grown in earlier fluids and some of which may be xenocrystic. The mass of crystals and fluid itself are always, *sensu stricto*, changing composition in response to one another. In essence, there is no, *sensu lato*, primary primitive (i.e., high MgO) magma from which all subsequent melt compositions derive via protracted crystallization, but, instead, a magma—probably of low MgO—whose identity is set by its dynamic state: being stagnant or in rapid transport. The physical state of the magma sets its present or local chemical evolution. Physically stagnant magmas are chemically stagnant. Physically active magmas entrain swarms of crystals react with them, lose them when they tarry, entrain more when are again on the move, and are constantly evolving chemically. In high volume, well-used systems active for long periods (100,000 s of years) of time, the system becomes chemically cleaner or less noisy due to ongoing reactions with debris in the

Fig. 11 Olivine composition as a function of bulk rock Mg number (Mg#) for Kilauea summit eruptions (after Garcia et al. 2003). As the lava bulk MgO content increases, the span of observed crystal compositions also broadens. The shaded field is the olivine composition expected from the Roeder and Emslie relation



system and the wall rock itself, but will, like Hawaii, nevertheless always show some characteristic noise, relative to small volume erratically used systems which will be inherently noisy. And basaltic magma, being much more able to shed a load of crystals upon being arrested, will show these characteristics much more than granitic melts that may, especially when they are crystal-rich, entrain and lose relatively little of their crystal load during transport and stagnation. That there are distinctive major basaltic magma types—tholeiitic, alkalic, and high-alumina—says that there are also chemically distinctive carrier magmas whose physical processes may be exceedingly similar even though the compositional outcome remains distinctive.

Slurries and layering

Slurries are concentrated, fluidized masses of particles and fluid. Soupy mortar, creamy cooked cereal, and sloppy mud are slurries. Any basaltic magmatic slurry consisting of an assemblage of primocrysts of varying size and density will undergo unmixing into layers during transport and deposition; it is an unavoidable process. The varying flux and crystalline nature of inputs or re-injections that go to building a large vat of magma can be assured to produce a highly layered rock record. Loading from the top, sides, or bottom will provide the requisite motions to induce layering. A stronger case can be made that all layering is due to slurry self-organization than the converse.

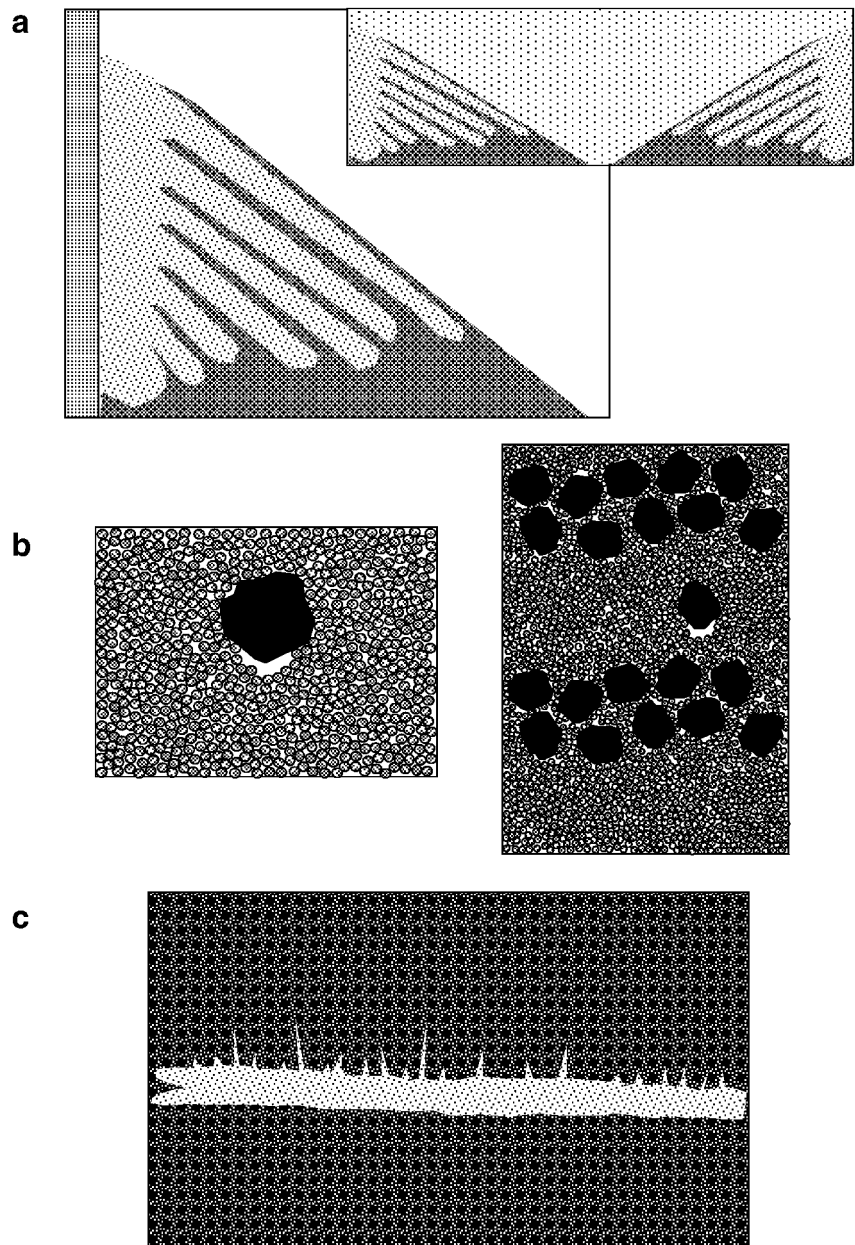
Granular processes

For bi- or multiply-dispersive particulate systems (i.e., two or more particle sizes or types), the natural physical equilibrium state is not a uniform mixture but an unmixed or segregated state arising from a wide spectrum of physical disturbances (Williams 1976; Jaeger and Nagel 1992). Any

motion at all of granular slurries will bring about unmixing or sorting into layers, pods, and/or pipes. The tendency to segregate or unmix arises due to contrasts in particle size, density, shape, or resilience. In building a chamber of magma from diverse inputs, these conditions are strongly met. This contrast in particle size, in addition to shape and density, leads to formation of several distinct types of crystal segregation. The first two below occur in both dry and fluid media. The basic physics of these and many other similar processes are discussed and reviewed by Aranson and Tsimring (2009; see also the review by Ottino and Khakhar 2000; see also Al-Naafa and Selim 1992; Davis and Acrivos 1985; Greenspan and Ungarish 1982; Jaeger et al. 1996; Kynch 1952; Pouliquen et al. 1997).

1. *Layering in Granular Avalanches*: If two species of particles, each possessing significantly different angles of repose and shape, are poured together to form a conical heap, a remarkable layering occurs (Meakin 1990; Makse et al. 1997). Coarse grains roll down a bed of fine particles much easier than fine particles roll down a bed of coarse particles. The fine particles sieve through and fill the interstices among the larger particles. As the pile grows, an avalanche eventually occurs and the grains reaching the base of the pile form a slope with a distinct break or kink in its profile, which runs back up slope to form a layer (Fig. 12a). Three layers are produced when three particle classes are used. The layering produced in this process is reversely graded, coarse over fine.
2. *Kinetic Sieving in Granular Flows*: In poorly sorted granular flows, particles become sorted by size not density, fining *downward* (Savage and Lun, 1988; Vallance, 1994). There are two fundamental processes. First, during avalanche transport of a well-mixed assemblage of small and large particles, small porosity fluctuations favor the downward inter-granular transport of small over large particles. The small particles

Fig. 12 Mechanisms of particle sorting in granular media and slurries. **a** Avalanche sorting due to differences in angle of repose related to particle size. **b** Upward migration of large particles in a finer medium due to kinetic sieving. **c** Interstitial sieving of small dense particles within a much coarser medium at maximum packing



sieve through the matrix of large particles, forming a basal layer. Once a high concentration of small particles forms, a concentration or layer of larger particles is left above, acting as a sieve to the further addition of small particles, hence the term kinetic sieving (Middleton 1970, 1993). Second, isolated large particles within a region of small particles are squeezed or lifted upwards by small grains rolling into tiny interstices beneath the large grains (Fig. 12b). A large steel ball bearing buried in fine sand, for example, will upon shaking migrate to the surface. This is also observed in simply vibrating or shaking a column of large and small particles (Jullien et al. 1992; Cooke et al. 1996), which leads to a form of granular

convection (Knight et al. 1993). In shear flows, both normally graded and inversely graded beds can be laid down due to flow timing and deposition.

3. *Layering and Segregation in Slurry Sedimentation:* A large contrast in size allows small particles to sieve through the matrix of larger grains, even when the larger grains are at maximum packing, as long as the ratio in particle size is <0.155 , which is the curvilinear triangular void space of close-packed spheres. Percolation velocity becomes large, and small grains plug the matrix and collect at specific horizons, spread horizontally, separating the larger grains and forming an extensive layer (Fig. 12c). It is literally impossible to prevent segregation and layering no matter how well

the assemblage is initially mixed. The higher the density contrast between the size classes, the better the layering or segregation. In experiments with metallic copper grains as the small particles and Lucite as the larger particles, the copper forms unusually sharp and distinct layers strikingly similar to the Dwars River chromite seams in Bushveld anorthosite.

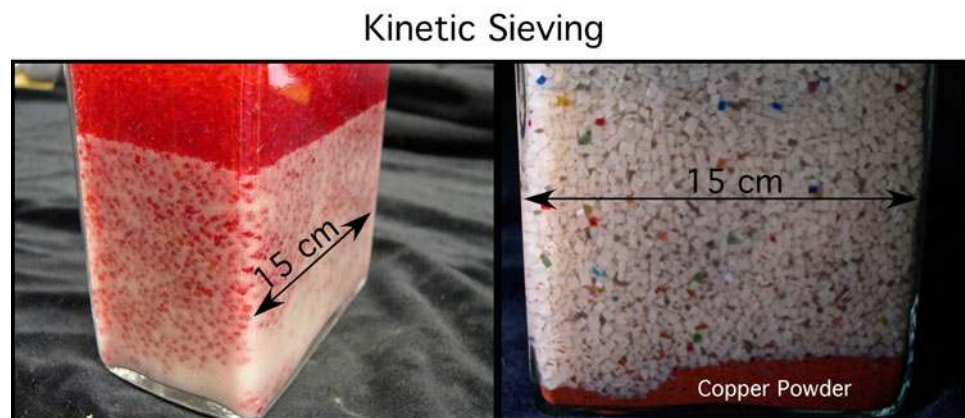
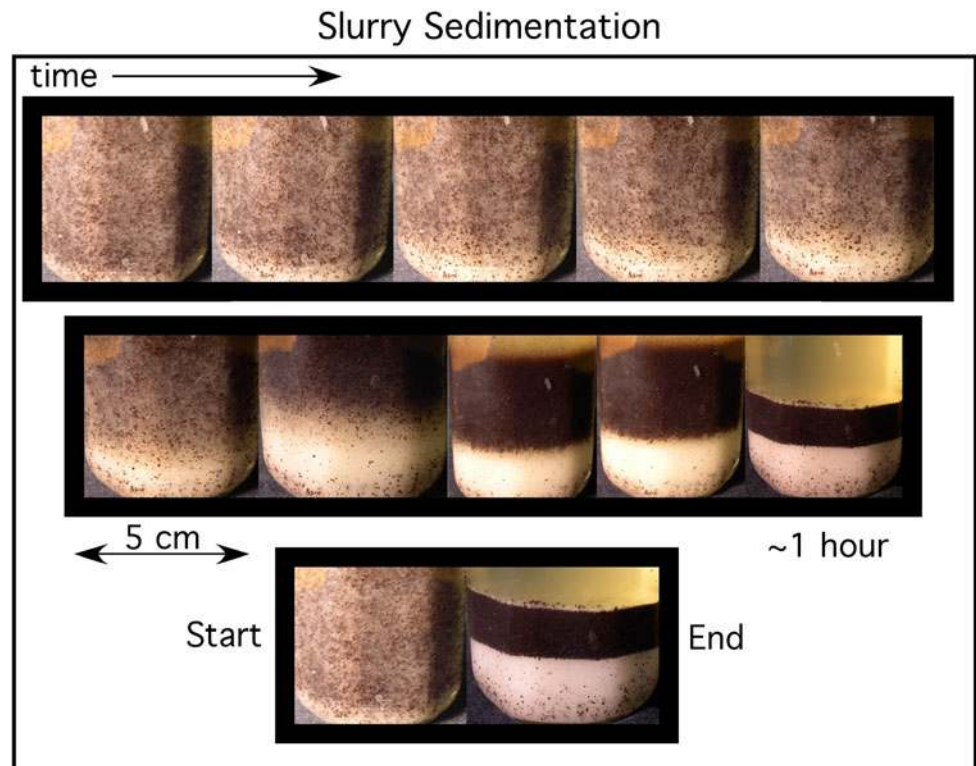
4. *Whitmore Instabilities in Hindered Poly-disperse Settling*: In batch settling of concentrated suspensions or slurries made up of many classes of particles (i.e., poly-dispersive), vertical columns or plumes of particles sometimes appear (Whitmore 1955; Weiland et al. 1984; Marsh 1988). This occurs with either a mixture of buoyant and dense or all dense particles and rapidly transports the entrained particles upward and the remaining particles downward thereby rapidly sorting the assemblage. Small (5×20 cm) vertical pipe-like concentrations of plagioclase are not uncommon in some coarse orthopyroxene cumulates (Bedard et al. 2007).
5. *Segregation in Granular Shear Zones*: The common occurrence of sharp narrow shear zones in Earth materials is a reflection of a granular medium. Granular materials do not shear homogeneously like fluids, but most often form rigid, locked regions separated by narrow shear or slip bands (e.g., German 1989; Nedderman 1992). At near maximum packing, shear causes local narrow dilation zones into which small grains selectively migrate forming distinct bands of segregated grains. If the large grains are loosely packed enough to also participate in the shear, the shear zone may grow significantly wider, perhaps almost without limit (Fenistein and van Hecke 2003). Granular shear zones can occur at any attitude from vertical to horizontal and are thus good indicators of flow direction in magmatic slurries. Vertical conical shear zones of this type are common in grain silos (Wojcik and Tejchman, 2009), and similar structures are sometimes found in feeder vents in primocryst-laden magmas like that supplying the Basement Sill of the McMurdo Dry Valleys (Fig. 14; Marsh 2004; Bedard et al. 2007).

Almost any conceivable slurry will undergo self-organization during transport and deposition. Figure 13 shows some simple experiments of slurry sedimentation and kinetic sieving. The upper sequence shows a series of snapshots over ~ 1 h of the unmixing of an initially well-mixed suspension of fine white sand (~ 0.5 mm) and larger grains of high-density paraffin (~ 1 g/cm³) in vegetable oil where the particle Reynolds No. ~ 1 . During sedimentation, there is an upward return fluid flow that cleanses the suspension of lighter particles forming in the end two clean layers. There is also a final uppermost fine layer of very

small, dusty sand grains that have been similarly cleansed from the assemblage by the upflow. Broadly, similar experiments using minerals were performed and analyzed by Coats (1936). The lower two examples show close-packed assemblages of Lucite particles in water with much smaller (<10 times smaller) intestinal grains of sand (left in red Lucite) and copper powder (right). In each case, vigorous shaking mixes the assemblage and upon resting the smaller grains sink through the interstices as in sieving and form concentrated layers. Here, most of the copper is at the base with a paucity of grains left in the matrix; in other experiments, the copper sometimes forms a layer up in the Lucite. Figure 14 gives some field examples of slurry processes from the massive (orthopyroxene) OPX Tongue of the Basement Sill. The upper figure shows fine ($<<1$ mm) plagioclase stringers or schlieren within massive coarse orthopyroxene (>10 times larger). The photomicrograph shows this process of plagioclase sieving downward between large (black, at extinction) orthopyroxene grains. The lower example shows a family of concentric and near vertical plagioclase stringers in the feeder vent area of Central Bull Pass; similar features have been found in a large feeder dike on Kerguelen Island (K. Nicolaysen, personal communication 2008). The high-density fine copper layering may be similar to the Dewars chromite seams in Bushveld, which may have been initially precipitated during reinjection at the interface between magmas of contrasting oxidation state, followed by annealing after deposition.

Although the above processes occur in high concentration slurries, broadly similar processes also occur in lower concentration suspensions of vertical pipe flows leading to flow differentiation as originally suggested by Baragar (1960) and discussed in detail in application to picritic sills by Simkin (1967). These shear flows sort and organize crystals horizontally and vertically, in each case fine to coarse from outside inward or downward, prior to emplacement, enabling slurry sedimentation to be even more effective in sorting the final assemblage. Because the marginal and leading edge magma is stripped of crystals, it is this melt that is most ‘differentiated,’ approaching ~ 7 wt% MgO, is the most mobile, and goes to form the chilled margins, which are commonly of unusually uniform composition for hundreds of km² about dolerite sills and plutons. These exact processes were envisioned by Bowen (1928; p. 148 etc.) when he accompanied Alfred Harker in the field on Skye in search of a true primary high MgO magma. Bowen found that the high Mg rock always had large concentrations of big olivine and that this olivine did not occur near the dike margins or leading tips. He saw that these crystals were sorted by the flowing magma and that the Mg content was due to sorting and not something fundamental to the magma composition. He had, without

Fig. 13 Examples of particle sorting due to simple sedimentation of a well-mixed slurry (*upper*) and kinetic sieving of small sand grains (*lower left*) through a coarser matrix and copper powder through a similar matrix (*lower right*). (See text for more explanation.)



knowing it, discovered ‘Tramp Crystals’ and the Simkin Effect.

Transport and sorting in igneous systems have long been recognized ever since the initial work on Shonkin Sag when Hurlbut (1939) realized that all the extensive crystal fractionation in this thin 70 m laccolith was due to massive settling of the 35 vol% phenocrysts present upon emplacement (e.g., Marsh et al. 1991; Marsh, 1996). Later work on many sills showed similar effects, but there has always been the over-riding urge to explain all such variations by the effects of layered precipitation through phase equilibria. This has produced a dichotomy in igneous systems between pure phase equilibria and physical transport and deposition. A major part of this dichotomy has been due the observed crystal sizes.

Body size and crystal size

One of the most perplexing aspects of field studies is recognizing the proper size of crystal belonging to any specific magmatic body. A chilled margin with a steady progressive inward increase in crystal size is nearly universally understood. And unusually large crystals in the centers of smallish dikes and sills are also easily recognized as crystals grown elsewhere at depth and having been carried in during emplacement. But as body size increases and the modal abundance and spatial coverage of such large crystals increases, it becomes much less clear from where these crystals came. Did they grow in situ or are they exotic? Large crystals in moderate bodies (e.g., 200–300 m sills) are sometimes loosely attributed to almost supernatural

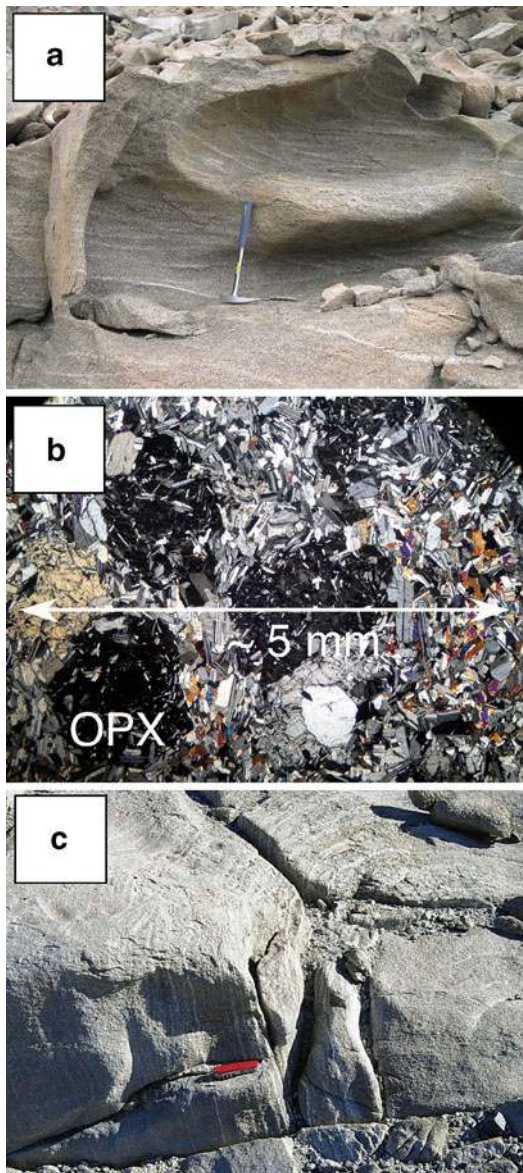


Fig. 14 Field examples of crystal sorting from the Dais Layered Intrusion of the Basement Sill, McMurdo Dry Valleys, Antarctica. **a** Plagioclase stringers or schlieren (*white bands*) in a much coarser grained orthopyroxenite; **b**) photomicrograph of small plagioclase grains sieving downward between large grains of orthopyroxene (*black grains*); **c**) near vertical sheets or stringers of small grains of plagioclase in a coarse grained orthopyroxenite reflecting upward flow

processes such as ‘gas fluxing,’ ‘halogen effects,’ ‘super-high growth rates,’ or even ‘gas cavity migration,’ which, if believed, go on to sidetrack any further cogent understanding of the evolution of the magmatic body. Just as clear signs of reinjection are lost or annealed away in bodies beyond a certain size, so too are clear signs of crystal provenance.

A standard gauge or standard state of crystal size relative to body size is available in the relation of CSD (crystal

size distribution) slope (S) and intercept (I) (e.g., Marsh 1988). CSD slope is a measure of the overall spread in crystal size; the lower the slope the larger the overall crystal size. CSD intercept is a measure of the population density (n°) of the smallest crystals. There is a natural tradeoff between intercept ($I = \ln(n^\circ(L = 0))$) and slope (S) found by Zieg and Marsh (2002):

$$\ln(n^\circ) = 4 \ln(S) - \ln(C) \quad (1)$$

The constant, C , is a factor accounting for the variations in modal abundance of the mineral being considered in any specific rock. That is, the population density of plagioclase in gabbro relative to peridotite will be different even though the variation in crystal size may be identical. In plagioclase phyric rocks, this is a small factor. All igneous CSDs so far measured match this relation exactly (see Fig. 15).

Given this relation, if either S or I is known the other is automatically given, which together gives the characteristics of the complete CSD. It is straightforward to show that S is given by the mean crystal size (Zieg and Marsh 2002). That is,

$$\bar{L} \equiv \frac{L_T}{N_T} = \frac{1}{S} \quad (2)$$

where L_T is the total length of crystals and N_T is the total number of crystals; each of these measures is directly available from the CSD. The mean length as defined here can also be intimately connected to the size of a body through a cooling model giving a time for local

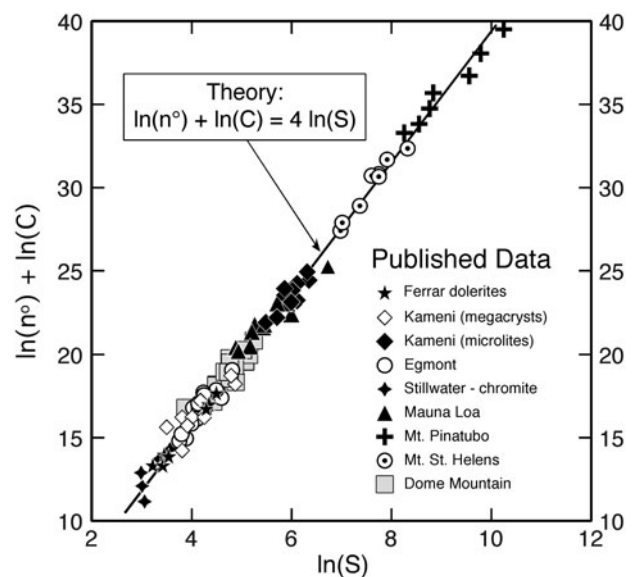


Fig. 15 The theoretical relationship between CSD Intercept (*left axis*) and Slope (*lower axis*) as shown by the solid line along with observed I and S values from a wide class of igneous rocks (after Zieg and Marsh 2002)

solidification and a prescribed growth rate. That is, any arbitrary growth law will connect the local cooling time Δt to body size. For example,

$$\bar{L} = G^\circ (\Delta t + \tau) \quad (3)$$

To be specific, the local cooling or solidification time (Δt) is the interval of time at any given location in the body for passage of the liquidus and solidus or solidification front, which measures the maximum amount of time any phase has to nucleate and grow at the rate G° . The factor τ accounts for the possibility of crystals of nonzero size appearing at the contact ($z = 0$). In deeper bodies, the contact temperature may approach or exceed the solidus temperature, which allows for phases nucleated at the contact to continue growing until the leading edge of the solidification front reaches the center of the body whereupon the prevailing contact isotherm itself begins propagating inward.

The time interval (Δt) for crystallization can come from a detailed numerical model or an appropriate analytical model, which is especially convenient to gauge the effect of body size on mean crystal size. The positions $z_i(t)$ of the solidus and liquidus each are given by an equation of the form

$$z_i(t) = b_i \sqrt{\kappa t_i} \quad (4)$$

where b_i is a Stefan-type constant containing the effects of latent heat of crystallization and the initial temperatures of the magma and country rock. The time at which the liquidus or solidus reaches any specific point within the body is found by rearranging (4) for t_i , finding the difference $\Delta t = t_S - t_L$, and substituting this into (3): namely

$$\bar{L} = G^\circ ((z^2/\kappa b^2) - \tau) \quad (5)$$

where now b^{-2} is a new constant representing ($b_S^{-2} - b_L^{-2}$).

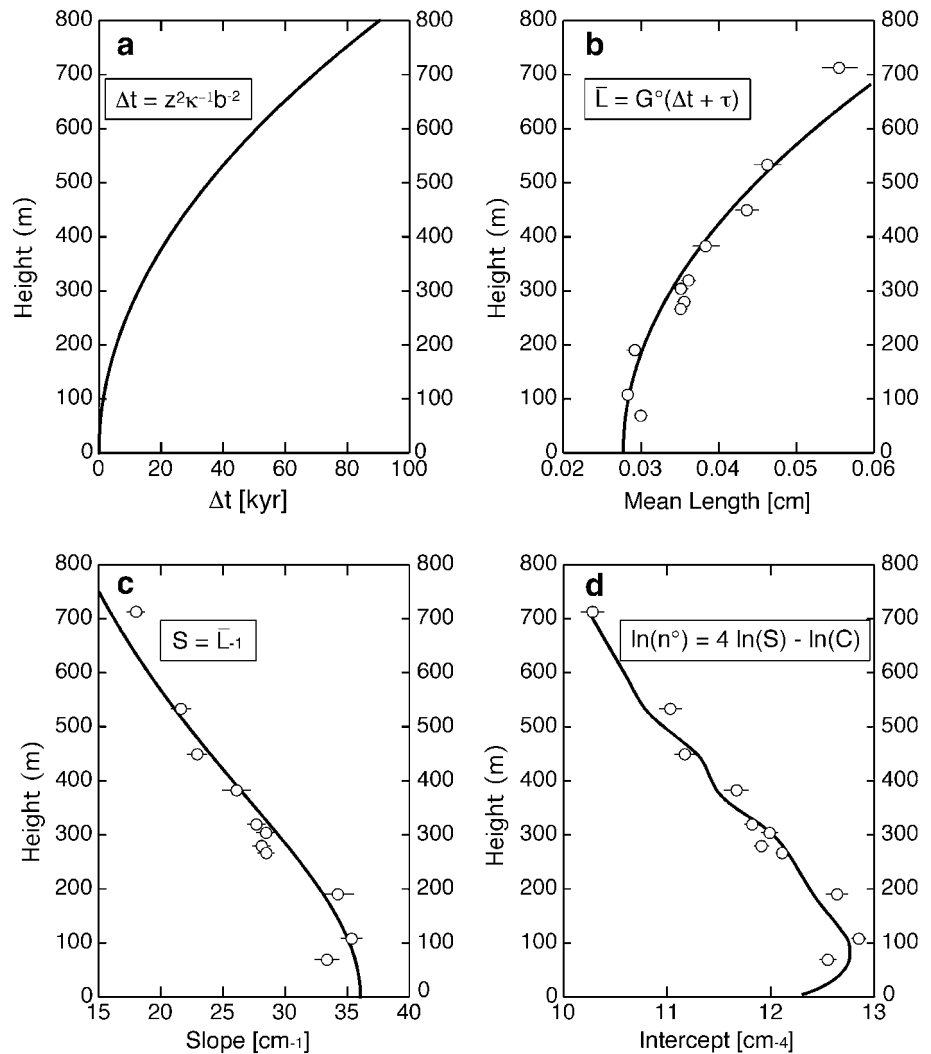
The CSD slope for this fresh crystallization event is given by the reciprocal of (5), and when used in (1), the intercept is also found along with the full CSD. Alternatively, on a plot of I versus S , the I and S values for any specific body will closely reflect the cooling time, and thus the overall size, of that body. That these relationships accurately predict the slope, intercept, and mean crystal size has been demonstrated in detail by measuring CSDs in drill cores through the norite of the Sudbury impact melt sheet (Zieg and Marsh 2002, 2005). These relationships are shown here as Fig. 16, where a growth rate of 1.55×10^{-14} cm/s has been used.

The distinct effect of body size and cooling time on crystal size is shown by a plot of I vs S (Fig. 17) for Sudbury and the Ferrar dolerites sills of the McMurdo Dry Valleys, Antarctica. Even though the bulk rock compositions are quite similar, the Ferrar sills (~ 300 m) being

much smaller than the Sudbury melt sheet (~ 3 km) plot at much larger values of I and S . And with increasing body size, I and S will progressively decrease along this curve. The effect of cooling rate from the sill chilled margin to the interior as well as from the basal contact at Sudbury upwards into the interior is also noted and is reflected in the crystal sizes. Going to even larger bodies, the CSDs for the Kiglapait intrusion (e.g., Morse 1979), which is ~ 8 km thick, have been determined by Higgins (2002), and the I – S relations for plagioclase, olivine, and clinopyroxene are shown by Fig. 18. These data adhere to the general I – S predicted trend, and at the larger slope values, which are at or near the upper or lower contacts, overlap with the deeper Sudbury data. The lowest slopes (coarsest crystals) and intercepts are in the interior, but when the S values for plagioclase and olivine are plotted against height an unusual pattern emerges (Fig. 19). That is, for almost the first basal kilometer, CSD slope follows that of Sudbury, which is expected for sheet solidification. Upwards beyond this level the CSD slopes progressively diminish with increasing height, strongly tending toward the values found at the upper and lower contacts; crystals get increasingly smaller. For progressive crystallization of a single thick 8 km sheet, a C-shaped variation in CSD slope is expected, as indicated on the Figure. Morse has long favored, as based on mass balance arguments through his PCS (percent solidified) index, that the intrusion crystallized from the bottom up with the last remaining melt existing at or near the roof. For a single 8-km sheet established all at once, such a model is hard to justify on heat transfer grounds, which favors symmetric cooling with the last melt to be more or less in the geometric center. But the CSD results generally imply (except for the single plagioclase point at 6.8 km) progressively faster crystallization upward through the body, possibly suggesting a long series of successive piecemeal emplacements as in a contiguous stack of large sills. In a hot mushy system, the internal contacts would be essentially erased by subsequent crystallization and annealing. More detailed CSD data are warranted to explore this possibility more thoroughly.

Some preliminary I – S data for Skaergaard (Lower Zone through Upper Zone) are also shown with the Kiglapait data in Fig. 18. Although there is some uncertainty in the true shape of Skaergaard (e.g., Norton et al. 1984; Nielsen, 2004), it has more or less the shape of a laccolithic loaf of bread with dimensions ~ 3 – 4 km thick, 5 – 8 km wide, and 9 – 11 km long, which makes it similar in thickness to Sudbury but much less expansive. That is, the Sudbury melt sheet was ~ 3 km thick and 200 – 250 km in diameter, giving it an aspect ratio similar to that of a compact disc with a depth of burial of ~ 3 km from fallback breccia. Given that Skaergaard is, in essence, an exceedingly small granule of the Sudbury sheet, and it is expected, with much

Fig. 16 Crystal abundance and sizes calculated for the basal norite of the Sudbury Impact Sheet as a function of distance above the base. **a** Shows the maximum time interval (Δt) for plagioclase crystallization with passage of the solidification front; **(b)** the mean crystal size for a constant growth rate and a linear growth law; **(c)** CSD intercept (I , solid line) along with the observed data (solid dots); **(d)** CSD slope (S , solid line) along with observed data (solid dots). After Zieg and Marsh (2002)



closer lateral walls and a broadly similar depth of burial, to have cooled faster than Sudbury, which solidified in about 10^5 years (Zieg and Marsh 2005). It is thus surprising in these preliminary data that the Skaergaard CSD slopes for plagioclase generally fall to the left of the Kiglapait data and far from the deepest Sudbury data indicating much longer growth times. To be specific, the Skaergaard plagioclase crystals are unusually large for a body of this size. Some difference in crystal size might be expected between Sudbury norites and Skaergaard gabbros, but this is excessive, especially relative to the Kiglapait results. The relative large size of the Skaergaard crystals is readily apparent if single thin sections are compared of similar rocks.

In summary, the general variation in crystal size within a body intimately reflects the size and thus solidification time of that body, which can be conveniently depicted on a CSD intercept-slope diagram (Fig. 20). Bodies containing crystals larger or smaller than those predicted by a representative cooling model will show up out of place on this

diagram. That is, lavas carrying slurries of phenocrysts/xenocrysts will appear as sills or plutons of a size appropriate to these crystal sizes. This also holds for plutons built by emplacements of crystal-laden inputs or slurries. Moreover, the spatial variation of CSD slopes and/or intercepts within the plutons is a clear indication of the pattern of solidification. These relations for Kiglapait and Skaergaard, for example, suggest that Kiglapait is made from a stack of successive emplacements with individual much shorter cooling times than that for a single ~ 8 km thick body; and Skaergaard crystals (plagioclase) are much larger than expected for a body of this size, clearly suggesting injection of slurries.

Discussion and conclusions

In basaltic magmatic systems where the conditions of formation are the most obvious, the final result seems almost obvious. Apparently regardless of size, systems

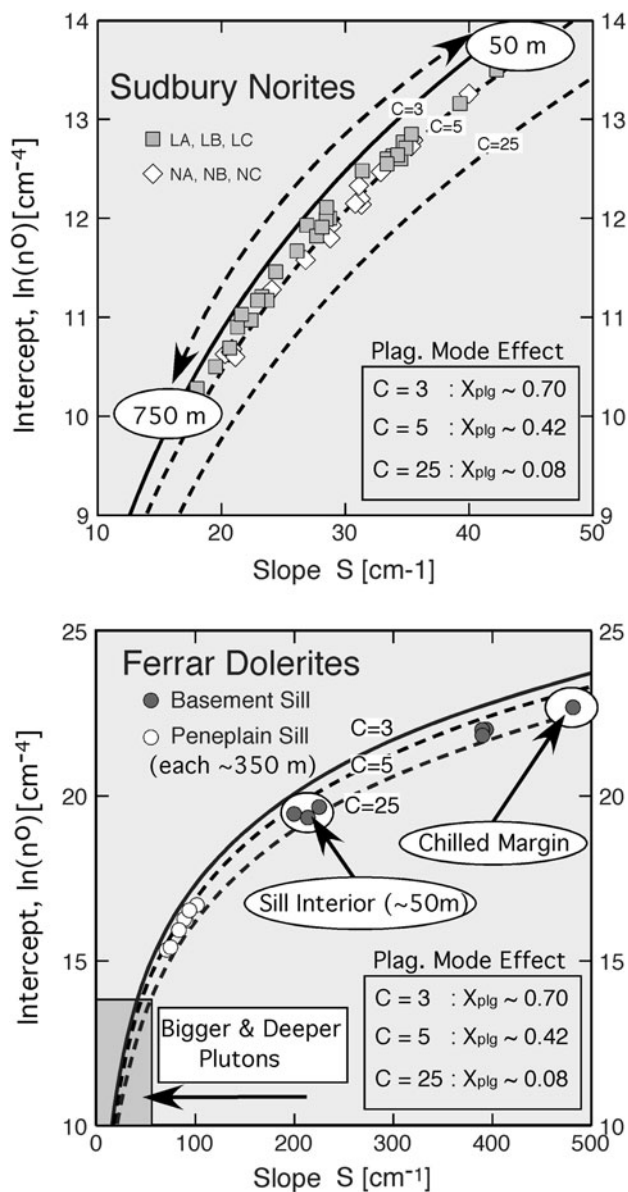


Fig. 17 CSD Intercept (I , left axis) and Slope (S , lower axis) relations for Sudbury norites (upper) and Ferrar dolerite sills (lower). The systematic variations along the curve show the effect of overall local cooling time on crystal size and abundance; crystal size increases systematically with depth into the body. After Zieg and Marsh (2002)

formed by magma initially free of large crystals produce nonlayered, uniformly varying sequences reflecting simple patterns of progressive cooling and solidification. Solidification fronts propagate inward from all margins containing all the freshly grown crystals, which are unavailable to fractionation. It is thus no surprise to find small batches of featureless primocryst-free magma; the conditions of transport and emplacement favor them. But to find large volumes of crystal-poor magma is much less common; the larger the mass of magma the more probable the mass will have initially contained high concentrations of large

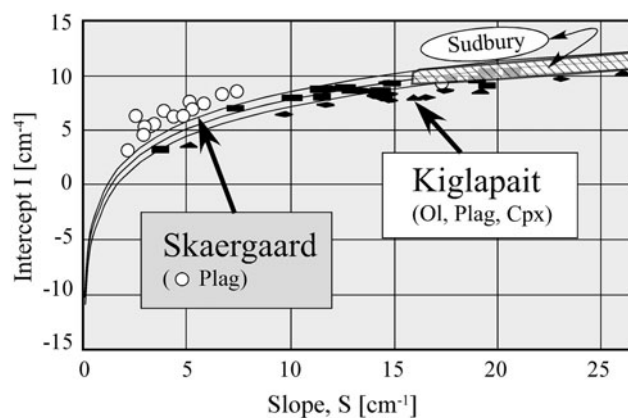


Fig. 18 CSD Intercept (left axis) and Slope (lower axis) for Kiglapait (data from Higgins 2002), Sudbury, and Skaergaard. Although Skaergaard is smaller than either Kiglapait or Sudbury, it has larger plagioclase

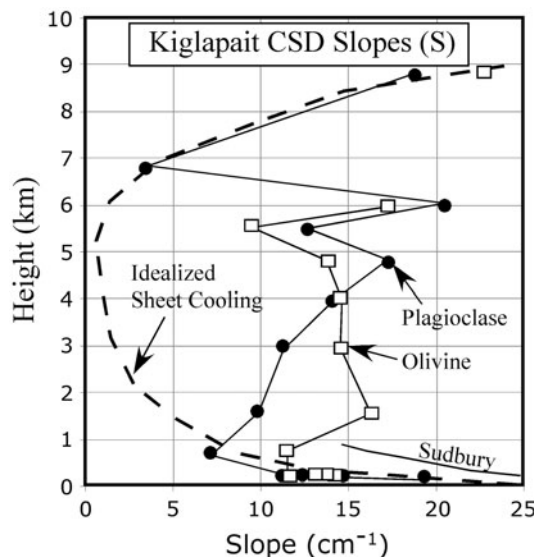


Fig. 19 The variation of CSD Slope (lower axis) with height above the base (left axis) for the Kiglapait Intrusion. The lowermost variations are similar to that of Sudbury, but upwards from there the CSD Slopes suggest a series of individual sheets cooling at much more rapid rates than for a body with the overall size of Kiglapait (~8 km)

crystals. Crystal-free magmas do exist in some flood basalt provinces, but more commonly at large volcanic centers, the issuing magma is highly varied in crystal content, reflecting the overall dynamics of transport throughout the system. For magma to differentiate by fractional crystallization crystals must be available to react with the melt, the ultimate origin of the crystals is almost immaterial. Crystal-rich magmas when examined in any detail are messy; the crystals may have a wide range of provenances. Yet as repeatedly seen in Hawaiian lavas and lava lakes, a heterogeneous collection of solids most often produces

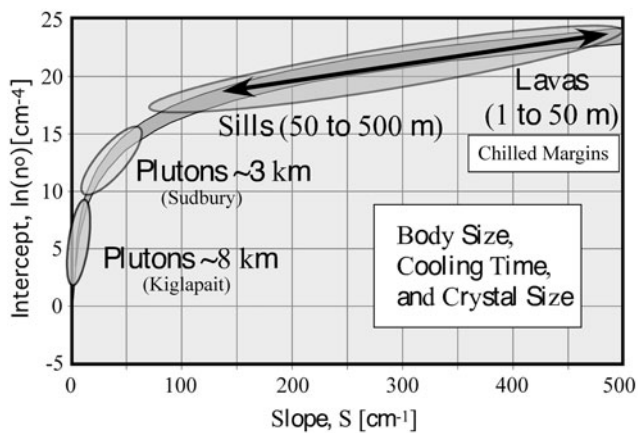


Fig. 20 The general relations between CSD Intercept (I) and Slope (S) expected for increasingly large igneous bodies from lavas to large plutons

exceedingly well-behaved patterns of chemical/crystal fractionation. And the problem is reduced simply to explaining chemical variations, from which oversimplified or unrealistic processes are assumed to govern the full evolution of the magma, chemical and physical alike. It is only when the crystals are chemically examined in great detail that the true complexities are revealed. Even more revealing, in terms of evaluating the full process of magma emplacement and the pattern of subsequent crystallization, is to carefully map the *spatial patterns* of crystal abundance, size, and bulk rock composition, of which the truly fundamental importance of each vanishes when considering only chemical variation diagrams. Sudden changes in primocryst abundance over short distances, near boundaries, or at variance with the thermal regime reveal the sequence of emplacement and the nature of the emplacement materials. These fundamentally important spatial variations are masked and lost in purely chemical scenarios.

Once the prevalence of magmatic slurries is recognized as central, fundamental initial features of magmatic systems, the spectrum of explaining all the richly varied forms, chemical and physical, of layered and nonlayered intrusions becomes attainable. Delivering alternating loads of crystal-rich and crystal-free inputs to chambers allows direct explanation of the observed features. These are certainly the crystal debris avalanches so vividly described and investigated by Irvine (e.g., 1980) for Skaergaard, and the varied cumulus textural developments so carefully portrayed by L. Wager and associates can easily come from slurry sedimentation. Random packing of deposited crystals provides a porosity of ~ 50 vol% at which point the magma is *en masse* immobile, but continued solidification allows for prolonged re-equilibration, extensive annealing, and textural refinement enhancing the initial template (e.g., Boudreau and McBirney 1997; McBirney and Nicolas

1997). And the wholesale flow of the remaining interstitial melt is, in essence, a catalyst to these processes, chemically connecting vast areas of the body never before connected. The final chemical outcome of this extensive, long-lived communication is that of homogenization, bringing closer to equilibrium what was initially a physically juxtaposed collection of solids and melt. The final rock record thus reads as one dominated more by chemistry than physics.

Magmatic systems, like all physical systems, end up pretty much as they start out. Assuming highly unrealistic initial conditions, namely instantaneously emplaced large batches of crystal-free magma, makes it literally impossible to reach the final rock record without invoking unrealistic processes. The crystals in Skaergaard, for example, are too large to have grown in a body of this size; and this is also reflected in the degree of compaction. Dumping a thick pile of large crystals on the floor of a magma chamber is the ideal condition for compaction to operate. The degree of compaction as exemplified by the deformed crystals in many Lower Zone Skaergaard rocks is remarkable relative to bodies of similar thickness, like Sudbury, where the crystals have grown in place. That is, in the Sudbury melt sheet, with no large crystals initially present, all the crystals grew in the solidification front under the prevailing stress field. Although there may have been some compaction, even at the base of 1 km of norite, it is not at all obvious. A sharp, even discontinuous change in spatial porosity upward in a stack of crystals provides ideal initial conditions for compaction to operate.

Slurries entering magma chambers are also exceedingly effective in crystal fractionation and thus strongly affect magma chemical evolution. This is because these crystals are inward and beyond the reach of lateral solidification fronts and cannot be captured. Punctuated differentiation takes place. A central feature of many chemical scenarios for magmatic evolution in initially crystal-free chambers is to produce crystals in the center of the chamber that can be fractionated to replicate the requisite chemical patterns. This is usually (of necessity) invoked by assuming some form of thermal convection to operate in the interior, which is taken to reflect wholesale cooling and crystal growth. But a strong case can be made based on all known experimental results in solidifying systems similar to silicates that thermal convection only occurs when the system contains superheat (Brandeis and Marsh 1989; Marsh 1989; Hort et al. 1999; Zieg and Marsh 2005). That is, thermal convection occurs only when the temperature is above the liquidus, which has never been observed on Earth.

Yet, there are no properly scaled convection experiments in actual silicates melts, which possess an unusually wide range in temperature between solidus and liquidus. Some field examples, however, may attest to the lack of importance of thermal convection. In Shonkin Sag

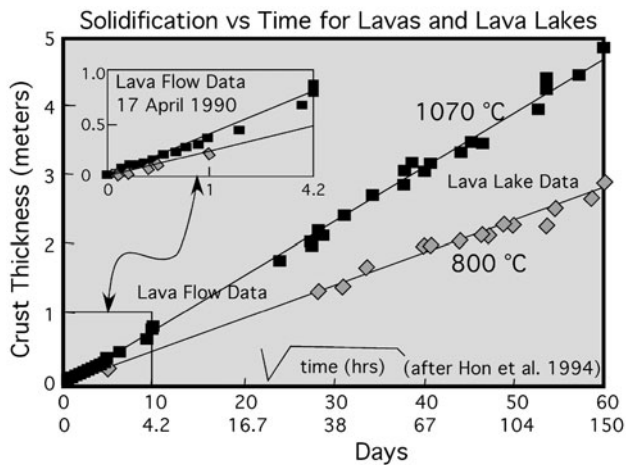


Fig. 21 The thickening of the upper crusts or solidification fronts of Kilauea lavas and lava lakes as defined by two isotherms (from Hon et al. 1994). That the crusts of thin lavas, where thermal convection is precluded, grow at the same rates as those of lava lakes suggests a similar lack of thermal convection in the much thicker lava lakes

laccolith, a central 15 m thick syenite layer resulted from the loss of the large crystals of primarily clinopyroxene in the initial magmatic slurry (Marsh et al. 1991). This low viscosity unit embedded below the upper solidification front was thick enough to have undergone thermal convection (i.e., the Rayleigh No., with even a tiny temperature gradient, would have been supercritical). During sedimentation and compaction of the underlying clinopyroxenite (Shonkinite), thin columns of interstitial melt were evicted upward in pipes to form symplectite pillows on the Shonkinite; at least one of these pipes rose upward traversing the full thickness of the syenite. That this pipe is straight and undeformed suggests the layer of syenite was stagnant. The second example is in the Isle au Haut bimodally layered sheet complex (Patwardhan and Marsh 2011). Here, layers of near liquidus basaltic magma formed a series of fairly thick (20 m) sheets within a more silicic, much more crystal-rich magma. The gabbroic layers were cooled from above and below, which did lead to some instability along the upper contacts, but here, again, pipe-like bodies emanating upward from the underlying layers are straight and undeformed, suggesting little convection. And last, in the measured rates of thickening of the solidification fronts or crusts of Kilauea lavas, which is a direct reflection of the mode of heat transfer, Hon et al. (1994) found the rates to be the same as that for much thicker lava lakes (see Fig. 21). That is, in thin lavas flows, the values of Ra are much too small to warrant any thermal convection as might be imagined in the much thicker lava lakes where Ra could be large. The similar rate of thickening thus precludes thermal convection. All present evidence thus points to no thermal convection in sheets of silicate magmas. The case of thermal convection along tall

vertical walls that may furnish limited stirring to an associated chamber remains to be experimentally investigated. Being that most larger gabbroic bodies are sheet-like, the overall effect of thermal convection driven by a wall, should it be found to operate, will be limited simply due to the amount of momentum available for stirring a large lateral mass.

Convection in response to density currents associated with the emplacement of crystal-laden slurries is clearly a highly viable and dynamic process. This occurs in response to uneven spatial variations in magma density, reflecting the entrainment and transport of crystal-rich magmas. This form of convection is not to be confused with purely thermal convection.

A sometimes-forwarded argument against assuming the dominance of slurries in forming layered intrusions is the lack of nearby sequences, unlike at Hawaii, of lavas showing high concentrations of large crystals. That this argument is specious is clear from the Ferrar Dolerites of the McMurdo Dry Valleys, Antarctica (Marsh 2004; Bedard et al. 2007; Marsh 2007). This system consists of four extensive (~100 km), thick (~350 m) sills capped by a thick section of Kirkpatrick basalt, equivalent to the Karoo basalts (Elliott and Larsen 1993; Fleming et al. 1995, 1997). The lowermost sill, Basement Sill, contains a vast ultramafic tongue of large orthopyroxene crystals that has developed a well-layered body, Dais Intrusion, in Wright Valley. The layering takes the form of up to 1 m layers of anorthosite inter-bedded in pyroxenite with a bulk composition of 20 wt% MgO. Upwards through the section, each sill shows less and less sign of any primocrysts of pyroxene. The thick overlying section of contiguous basaltic lavas shows no sign whatsoever of large mafic crystals. And this holds all across the Transantarctic Mountains, some 2,000 km, even though some underlying sills carry large crystals and the massive (~450,000 km³) layered Dufek Intrusion (e.g., Ford and Himmelberg 1991), with an overall composition almost identical to the Basement Sill, forms part of this same system.

Conclusions

For a more than a century, magmatists have been trying to explain rock records using untenable initial conditions. Crystal-free, instantaneously emplaced magmas form uniform predictable expanses of rock; exactly the opposite of what is often desired. Tenable initial conditions involve the periodic emplacement of crystal-laden slurries, the more varied, the more exotic the final rock record, and the stronger the process of fractional crystallization. Physical processes buttressed by chemistry govern magmatic processes.

Acknowledgments Deeply and sincerely appreciated are forty years of discussion and debate, beginning with Ian Carmichael and continuing with fellow magmatists and my own students—you know who you are, a century of rich petrology literature, and soul searching reviews by Edmund Mathez, Alan Boudreau, and Ryan Currier. Mike Rhodes directed me to the olivine modal data for Fig. 8b. This work is funded by the National Science Foundation (OPP 0440718).

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