On the intermittency and crystallization mechanisms of sub-seafloor magma chambers

C. R. B. Lister School of Oceanography and Geophysics Program, University of Washington, WB-10, Seattle, WA 98195, USA

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Summary. The conventional view of a magma chamber being an essentially permanent feature of a fast-spreading ridge is not compatible with the physics of either oceanic crustal structure or internal magma convection. Cumulates form a large volumetric fraction of many ophiolites and are deposited at the bottom of magma chambers, so they are not available as an insulating layer between magma and seawater. Extrusives and dykes are cracked and highly permeable to hydrothermal convection, so they also do not offer much thermal resistance to the cooling of magma. Only the layer of ‘plated’ or ‘isotropic’ gabbro, often observed between cumulates and dykes, is limited to conductive heat transport; a 0.5 km thickness implies a chamber lifetime of no more than 10 kyr km$^{-1}$ of magma. The intermittency of the chamber on such a short time-scale requires that the plates move apart to make room much more rapidly than they could at the steady spreading rate. Fluctuating magma pressure can achieve such intermittent movement by stress-change diffusion through an elastic lithosphere overlying a viscous asthenosphere. However, the space needed implies stress diffusion almost all around the Earth, and this, in turn, requires that the intermittent spreading be substantially synchronized all along a major segment of ridge.

The pressure-dependent slopes of equilibrium temperatures between crystal phases and liquid silicate magmas exceed the adiabatic gradients of the magmas themselves by about $10^6$ C kb$^{-1}$. If this temperature difference is considered the convective drive, a 3 km high chamber reaches a Rayleigh number of $3 \times 10^{15}$, and a Nusselt number of about 8000. The plated layer should grow until the heat loss rate decreases to that supplied by cumulate-depositing convection, implying a plated layer about 0.5 km thick, in agreement with observations. The boundary-layer/turbulent core structure of convection at high Rayleigh and Reynolds numbers is consistent with the formation of banded cumulates when a new type of fluid circulation is taken into account. If a suspended crystal phase is present in the bulk fluid, a ‘slow-convection’ mode is possible, where flow velocities are restricted by the equilibration rate between crystals and fluid, and the temperature profile is determined by the thermodynamics of crystal-fluid equilibria.
Introduction

The temperature and physical state of the rocks near a plate-tectonic spreading centre are determined by the temperature and heat content of upwelling material, and by cooling losses through the upper surface. An upper bound to the temperatures can be obtained by considering only thermal-conductive losses, and the flow geometry is amenable to straightforward calculation if the injection is imagined to occur on an infinitesimally thin zone at the mathematical origin. Such ‘thermal balance’ calculations have been done for a uniform unmelted material (Davis & Lister 1974), a molten ‘crustal’ layer overlying unmelted material (Sleep 1975) and a solidification crust forming on a fully molten base (Parker & Oldenburg 1973). The results are all similar, in that the time-averaged mean temperature at the divergence origin approaches zero surface temperature in a smooth manner, and at a slope dependent upon the spreading rate. Since magma can only be injected at or above its solidus temperature, this means that the injection process must be intermittent, in agreement with geological observations of sheeted dyke complexes. A block-type intermittent injection process can in fact be modelled by somewhat more complex mathematics (Royden & Parsons 1982) and shown to agree with the continuous flow models when the block thickness is allowed to go to zero.

The geological evidence from ophiolite suites (e.g. Pallister & Hopson 1981) is that at least some ocean crust contains a significant thickness of cumulate gabbro that is deposited by crystallization from a magma chamber. The calculations of Sleep (1975) show that no magma chamber can be present on a steady-state basis at spreading rates below 1 cm yr\(^{-1}\), but that magma chambers of considerable size are possible at high spreading rates. It has therefore become accepted that most ophiolite suites were produced at high spreading rates, extremely high if the field data on the angle of the cumulate bedding planes is taken at face value (Pallister & Hopson 1981). For under steady-state conditions, the build-up of the cumulate layer must occur at the base of a magma chamber shaped like an inverted triangle (e.g. Lister 1977; Pallister & Hopson 1981). This might be an acceptable solution were it not for other field evidence that describes vigorous hydrothermal circulation occurring close to the divergence origin at seafloor spreading centres (Spiess \textit{et al.} 1980). Any hydrothermal circulation requires that heat losses be much greater than conductive; how much greater cannot be specified without a theory for geothermal phenomena. The extant theory of water penetration into hot rock (Lister 1974, 1982) makes it very clear that high-temperature geothermal systems can only exist when water is actively penetrating into a body of hot rock by thermal-contraction cracking. Surface lavas and the sub-adjacent sheeted dykes cool on a piecemeal basis to form their characteristic structures. Their heat content is dissipated locally in a time short compared to their emplacement interval. The only substantial body of hot rock in the crustal layer is the mass of cumulate gabbros themselves, and thus the presence of high-temperature geothermal systems right at the axis of spreading would indicate that the cumulates are cooled hydrothermally before moving away from the axis. This is incompatible with a magma chamber of the inverted-triangle variety, and means that magma can be present only intermittently at the spreading-centre axis.

This step of reasoning requires that the theory of hydrothermal penetration be accepted as valid; and while no contrary evidence has yet been found, it may be helpful to arrive at the same conclusion by another method. It is clear that the surface lavas and the sheeted dyke complex are pervasively cracked and therefore permeable (e.g. Johnson 1980), so they do not provide the insulating layer needed by the thermal-balance calculations. Only a moderate permeability is needed to allow vigorous convection and a heat transport (or Nusselt number) many times the conductive level at the same overall temperature difference. Thus the cold boundary is effectively transferred to the base of these layers. The cumulates
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themselves are deposited at the bottom of the magma chamber and provide no barrier to the loss of heat through the top. Fortunately, all complete ophiolite suites contain a layer of ‘isotropic gabbro’ between the sheeted dykes and the cumulates; this is accepted as plated on to the top boundary of the magma chamber, and is thus the insulating layer we are seeking. Maximum thicknesses of about 1 km have been observed, implying a thermal gradient of at least 1200°C km⁻¹ and a heat-loss rate of 72 μcal cm⁻² s⁻¹ (HFU) as a minimum (basalt conductivity 0.006 cal cm⁻¹ °C⁻¹ s⁻¹). The maximum crystallization life of a 3 km high magma chamber with no heat losses to the sides can be calculated by taking a representative heat of crystallization of 250 cal cm⁻³; it is only 30,000 yr. Even at 5 cm yr⁻¹ spreading rate this is only a 1.5 km half-width, invalidating the assumption of no lateral heat losses, and in any case requiring a steady-state cumulate deposition angle of 60°, which is an impossibility.

Thus, there is clear-cut thermal evidence that magma chambers at spreading centres can only be present intermittently. Yet the pervasiveness of cumulate gabbroic outcrops in ophiolite suites show that the chambers, when present, are of considerable lateral extent. The problem now is to discover how a ridge-crest can make room for a large magma chamber in a time short enough for it to develop without being crystallized before it can form.

The response of the lithosphere to axial stress changes

The difference between a quiescent ridge crest, devoid of active volcanism, and one with a magma chamber present, is largely in the pressure head applied to the partially molten material rising along the axis. After hydrothermal cooling, the crustal layer should be largely cracked and permeable, so that the lateral pressure on that part of the lithosphere should be close to hydrostatic, not lithostatic. The absence of volcanic activity means that the overall pressure in the magma storage zone is not sufficient to upwell into the cracked region. Such a low pressure can be caused by the frictional heat loss to material upwelling through some kind of conduit, as described by Lachenbruch (1976). As the magma continues to segregate from a partial melt, the quantity of low-density mobile fluid builds up, and the pressure near the base of the crust builds up to where magma can penetrate and move up to the surface in the form of a dyke. The pressure in the entire magma system must now be magma-lithostatic. For the purpose of constructing a simple model, one could consider that the pressure in the divergence zone has gone up by about the equivalent of 5 km of magma replacing water, and that this pressure change is applied to about 15 km of young lithosphere.

A schematic of the situation is sketched in Fig. 1. Without trying to solve the problem of how the lithosphere actually acquires all of its new material, it is reasonable to imagine that the pressure change of 800 bar is applied to the young lithosphere via the magma conduit system, and a change of 400 bar is applied to the crust. If ‘typical’ lithosphere can be taken as 70 km thick, the mean stress change some distance from the spreading centre would be about 200 bar, and the resulting compression would be 200/1.5 x 10⁶ = 1/7500 given a Young’s modulus for rocks of about 1.5 Mb. Thus, compression of a substantial lateral extent of lithosphere, oceanic and continental, could make enough room for a magma chamber of 1 km half-width. It should be noted that this reasoning applies only to the differential stress produced by intermittent magma upwelling. There is no need to know what force or forces cause the overall motions of the plates, provided they can be considered steady over the period of the magma cycle.

The next question is whether the lithosphere can respond rapidly enough to the stress change, since motion over the asthenosphere must occur, both to transmit the stress and to make room for the magma chamber. Since only an order of magnitude answer is needed, a very simple model can be used. A flat elastic lithosphere of uniform thickness is presumed.
Figure 1. (a) Quiescent phase of the ridge. Tectonic and/or hydrothermal cracks penetrate crustal layer; magma accumulates somewhere beneath. Lateral pressure on crust water-hydrostatic: \( P_i = \rho_{\text{w}}gh \); low magma pressure in reservoir. (b) Sheet-flow stage of ridge volcanism. Magma rises to surface giving rise to dyke swarm. Beginning of crustal magma chamber. Tectonic cracks in crust closed: \( P_i = \rho_{\text{M}}gh \); high magma pressure in reservoir due to extension to surface.

to overly an asthenosphere of constant Newtonian viscous properties (Fig. 2). Ignoring divergences in the asthenosphere allows its properties to be reduced to the generation of a shear stress in response to a sliding velocity. If it is 100 km thick and has a viscosity of \( 10^{19} \) poise, the response is \( \xi = 0.03 \) bar cm\(^{-1}\) yr in the usual units of seafloor spreading velocity, \( u \). If the stress is \( \sigma \) in a lithosphere of thickness \( l \), the mechanical equilibrium of a thin slice requires that

\[
\frac{l}{\frac{\partial \sigma}{\partial x}} = \xi u
\]

where \( x \) is a horizontal coordinate. The length must change in order to build up stress, so that

\[
\frac{\partial \sigma}{\partial t} = Y \frac{\partial u}{\partial x} = \frac{LY}{\xi} \frac{\partial^2 \sigma}{\partial x^2}.
\]

This is the familiar diffusion equation, with a diffusivity \( LY/\xi \) that is calculated readily. For the 'typical' lithosphere of 70 km elastic thickness, it is \( 3.5 \times 10^{14} \) cm\(^2\) yr\(^{-1} \).
Figure 2. (a) Schematic model of an elastic lithosphere overlying a viscous fluid asthenosphere. More realistic visco-elastic gradations would not affect the order-of-magnitude of the result. (b) Relations between transient stress $\sigma$, transient velocity $u$, and basal drag in the elastic layer.

A step change in stress produces the standard error function solution (Carslaw & Jaeger 1959) whose characteristic width increases according to the square root of time, as does its integral. The stress change spreads about 190 km in 1 yr, 1900 km in 100 yr, and essentially all around the Earth in 1000 yr.

The pulse of spreading at the ridge crest will also follow a square root law following a step change in stress, with a half-width of 29 m available after 1 yr, 290 m after 100 yr and a maximum of 2.9 km approached after 1000 yr. The situation is somewhat more complex on a real spheroidal planet, and synchronism of the stress pulse over a great length of ridge would be required to allow the opening to approach the maximum. The idea that magma chambers are formed intermittently does, however, survive this simple physical test.

Magma convection and the plated gabbros

Between the cumulate gabbros deposited from a magma chamber, and the sheeted dykes produced by the near-surface spreading process, there is a layer of 'isotropic' or 'plated' gabbro (e.g. Pallister & Hopson 1981). It is recognised by the terminology that this is the 'chilled margin' of the magma chamber, and forms the solid thermal barrier layer between the convecting magma in the chamber, and the convecting seawater in the permeable surface rocks. Thicknesses range up to 1 km, and it is another interesting physical test of this model of a cooling magma chamber to see if the cooling rate through such a layer is consistent with the natural convection rate in a large magma body.
If the heat loss through a ‘plated’ boundary layer is greater than the heat supplied by convection, the layer should grow by the further ‘plating’ of more solids. An equilibrium can be reached when the mean thermal flux through the convective boundary layer equals the thermal flux through the solid boundary layer. When this situation is approached, the magma chamber heat loss is supplied by the latent heat of crystallization of material deposited as cumulates on the floor of the chamber.

The first, simplified, view of the convective drive available in the magma chamber is to assume that the working fluid is just liquid magma, and that crystals are only present in the crystallizing bottom boundary layer. The working fluid is in contact with solid at the top and also at the bottom, so that the temperature difference is prescribed by the gradient of the liquidus temperature with pressure. Stirred fluid would approach an adiabatic gradient of temperature versus depth and pressure, so that the effective thermal drive is the liquidus-controlled temperature difference minus the adiabatic temperature difference.

According to Fujii & Kushiro (1977) the slope of the liquidus temperature versus pressure is about 3.7°C kb⁻¹ for a representative basaltic liquid. The adiabatic temperature gradient can be calculated from the standard thermodynamic relation
\[
\frac{\partial T}{\partial P} = \frac{\alpha V T}{C_p} \frac{1}{T}
\]
where \(C_p/V\) is the thermal capacity of the liquid, taken as 0.8 cal cm⁻³ °C⁻¹, \(T\) is the temperature, 1510 K, and \(\alpha\) is the thermal expansion coefficient of the liquid. Quoted values of this last range from 3.8 to 28 x 10⁻⁵ °C⁻¹ (Skinner 1966), implying an adiabatic temperature gradient from 1.7 to 12.6°C kb⁻¹. At the low end, this would imply a highly unstable, convecting magma chamber, while at the high end it would imply a stable, non-convecting chamber that would have to to crystallize from the top down.

Such a range of possible values of adiabatic temperature gradient is unsatisfactory, so further understanding must be sought. Fortunately, Bottinga & Weill (1970) have found that the densities of liquid silicate systems correspond closely to those calculated from the partial molar volumes of the oxide components and their densities. Thus the thermal expansion coefficient of a complex liquid mixture should be closely approximated by the volume-weighted average of the expansion coefficients of the components, as measured in a silicate melt. The appropriate expansion coefficient of silica itself is very low, that of alumina low, the divalent metal oxides high and those of the alkali metals very high. The values for water and ferric oxide are, understandably, not known but should not be out of line with their homologs, although the highest values quoted by Skinner (1966) do exceed the mean that would be calculated for even a highly potassic granite. The range for mafic magmas is indicated by the values for pure liquid Forsterite, 5.2°C kb⁻¹ (melting point gradient 5°C kb⁻¹) and that for pure liquid Enstatite (MgSiO₃) 3.6°C kb⁻¹ (melting point gradient 9°C kb⁻¹). The increase in the proportion of silica decreases the adiabatic gradient in the melt, but the effect of crystallizing mineral variations on the liquidus gradient of a complex mixture has not been related to a simple formalism and may well be inherently intractable. For the purposes of calculating a value for a representative oceanic basalt I use the composition quoted by Smewing (1981) for an ‘isotropic gabbro’ from the Semail ophiolite. The expansion coefficient comes out at 5.6 x 10⁻⁵ °C⁻¹, implying an adiabatic temperature gradient in the melt of 2.5°C kb⁻¹. This is not the same gabbro as that of Fujii & Kushiro (1977), but there is little choice other than to accept their value of the liquidus gradient of 3.7°C kb⁻¹. A magma chamber 3 km high develops a pressure difference of nearly 1 kb between top and bottom, so that the thermal drive for convection would appear to be about 1°C.

If the height, expansion coefficient, and temperature difference are known for a chamber of liquid, only the viscosity and thermal diffusivity are needed to calculate a Rayleigh number for the convection. Fujii & Kushiro (1977) quote values of about 100 poise for the viscosity of their molten basalt at the liquidus at a total pressure of 1 kb, the value representative for the middle of a sub-oceanic magma chamber. The estimation of a thermal diffu-
sivity is at once more of a problem and less of one; no reliable measurements appear to exist (even for solid basalt at high temperatures) but the value can hardly differ dramatically from 0.005 cm²s⁻¹. This has been arrived at from the trend for diabase glass (Clark 1966) and an assumed heat capacity \( C_p/V \) of 0.8 cal cm⁻³°C⁻¹. An order-of-magnitude estimate of the Rayleigh number, \( \alpha g \Delta T h^3/\nu \kappa v \) is therefore \( 3 \times 10^{15} \), extremely high. The extreme magnitude of Rayleigh numbers in magma chambers has been remarked upon already by Huppert & Sparks (1980). For free convection between plane parallel perfectly conductive boundaries, the heat transport, or Nusselt number, has been measured up to a Rayleigh number of \( 4 \times 10^9 \). The divergence of the actual conditions from the ideal probably has little effect on the heat transport (see further discussion) but the extrapolation required is substantial.

The literature on convection contains a number of different power-law relationships between the Nusselt and Rayleigh numbers. In the formula \( N = AR^x \) the values for \( A \) range from 0.062 to 0.76 and for \( x \) from 0.278 to 0.346, suggesting that there are problems with the data. Katsaros et al. (1977) have critically reviewed the prior work and compared it with their experiment, where heat losses to the sides of the apparatus were well-constrained and higher values of Rayleigh number were achieved than before. In the range of Rayleigh numbers between \( 10^6 \) and \( 10^9 \) where their experiment can be compared with the two best prior studies (Rossby 1969; Chu & Goldstein 1973), the actual values of Nusselt number agree quite well, but the slope is steeper and therefore the value of \( x \) is higher. This is in spite of their experiment having had a free evaporative boundary versus the rigid conductive boundary of the others. According to the theory of Howard (1966) the difference between the boundaries is 16 per cent in the coefficient \( A \), and the agreement suggests that the correct approach to adjustment for differing boundary conditions is to raise the ratio of the critical Rayleigh numbers to the power \( x \). Katsaros et al.'s results suggest that \( x \) is exactly \( 1/3 \), which is what would be expected from the theory if the convective structure consists of thin boundary layers and a substantially isothermal core.

Taking their values, for the rigid boundary case, of \( A = 0.062 \) and \( x = 0.33 \), the Nusselt number is 8000, and so the thermal gradient in the boundary layer is 8000 times the super-diabatic gradient in the magma. If the latter is about \( 1^\circ C \) (3 km)⁻¹, the boundary layer gradient is about \( 2700^\circ C \) km⁻¹, and, extrapolated from magma core temperature to a hydrothermal boundary at or near seawater temperature (\( 0^\circ C \)), the layer of 'plated gabbro' should be 0.5 km thick. This result is of the right order of magnitude to be consistent with the field data from ophiolite suites. Because of the \( -1/3 \) power-law dependence on Rayleigh number, the thickness should grow gradually during the life of the magma chamber, as the convective height decreases. In the final stages of crystallization, the residual liquid is highly differentiated and the figures used would no longer apply, but this phase produces the keratophyres and plagiogranites often observed in the ophiolite sequence.

Boundary layers and the crystallization process

So far, the magma has been treated as a simple working fluid confined between plane conductive boundaries. A real chamber may be approximately equidimensional, at least in one vertical plane, and does not have insulated walls. The effect of such geometric variations on vigorous convection is of the order of a factor of 2, and not critical in order of magnitude investigations. More important is the departure of the top and bottom boundaries from constant-temperature conductors. The top boundary presents little problem; it is a heat-flux boundary, rather than a constant temperature one, but since the temperature rises rapidly downward from the magma liquidus, the working fluid should be pure magma. The heat transport should be very similar to that for a conductive boundary because of the thermostat effect of the solid—liquid transition. An order of magnitude estimate for the boundary-layer
thickness is $h(R_c/R)^{1/3}$, where $R_c$ is an effective critical Rayleigh number, about $2 \times 10^3$ (Howard 1966). If $h = 3$ km and $R = 3 \times 10^{15}$, the boundary layer should be only 0.13 m thick. A similar boundary layer should be present along the sides of the chamber if they cool through a layer of plated gabbro like that at the top. Thus the bulk of the magma in the chamber should form the turbulent, statistically isothermal core, with boundary effects very small scale and fluctuations on a short time-scale.

The bottom boundary is more difficult and more interesting, because it is the site of deposition of the cumulates, the geologically observable products of the convection. The heat source is the latent heat of crystallization of the magma, and appears wherever crystals form at a significant rate — not necessarily at the boundary. In the first, simple, case let us assume that a basic magma is producing olivine and pyroxene crystals, both substantially denser than the magma. The settling rate of a 4 mm sphere of density contrast $0.5 \text{ g cm}^{-3}$ is only 0.5 mm s$^{-1}$ in a fluid of viscosity 100 poise. On the other hand, the accumulation rate of material with $253 \text{ cal cm}^{-3}$ of latent heat is only $6 \times 10^{-6}$ mm s$^{-1}$ if the conductive heat loss through the top of the chamber is $1.6 \times 10^{-4}$ cal cm$^{-1}$ s$^{-1}$ (160 HFU), corresponding to a plated gabbro layer 0.5 km thick. Neither a high density of crystals, nor a thick settling layer are needed to achieve this rate, which is just as well, since the maximum superadiabatic temperature difference is only about $1^\circ$C. A fluid starting at the liquidus, but free of crystals, at the top of the chamber, would, if in crystal equilibrium, heat up above the adiabat by separation of crystals and release of latent heat. The crystal content would be of the order of $\Delta T C_p/\rho L$ or $3 \times 10^{-3}$ by weight, when in equilibrium near the bottom.

The settling of dense crystals has another important effect: the residual fluid has a substantially lower density than the mixture had before, and should convect vigorously away from the boundary. The drive can be compared to the thermal expansion drive in normal convection experiments by comparing the density difference produced by the settling of $3 \times 10^{-3}$ by weight of crystals, to the density difference produced by $1^\circ$C applied to the thermal expansion coefficient of pure magma. The settling drive ($\Delta \rho = 0.5 \text{ g cm}^{-3}$) is $1.5 \times 10^{-3}$ g cm$^{-3}$, while the thermal expansion of basaltic liquid with a density of $2.7 \text{ g cm}^{-3}$ produces only $1.5 \times 10^{-4}$ g cm$^{-3}$. Thus the basal convective drive due to crystal settling is 10 times larger than the equivalent thermal drive would be, leading to a bottom boundary layer even thinner than that at the top of the chamber, and mixing so vigorous that the temperature gradient may be closer to liquid-adiabatic than crystal-equilibrium. The picture of a chamber depositing only dense crystals, such as olivine and pyroxene, may therefore correspond to Fig. 3, where the main superadiabatic temperature drop occurs in the top boundary layer, and there are zones near the bottom where crystals are forming and redissolving.

The problem of plagioclase

The real situation in most cumulates is complicated by the fact that there is a common crystalline phase, feldspar (plagioclase), whose density is very close to that of the liquid magma, and possibly, under some circumstances, even lighter. This is different from saying that there is no volume contraction upon crystallization, since the density of pure molten plagioclase is much lower than that of a normal basaltic liquid. Bottinga & Weill (1970) have found that the density contribution of each oxide to a silicate liquid is essentially independent of overall composition. It is therefore possible to prepare a table of the common gabbroic minerals, their densities near magma liquidus temperature, and the density of their melts. This has been done in Table 1, using Robie et al. (1966), Skinner (1966) and Deer, Howie & Zussman (1963) as sources for the mineral data. In all cases, the crystals are
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Figure 3. (a) Simplified physical picture of a magma chamber depositing dense cumulates. Important features are the plated insulating layer top and sides; thin, turbulent convective boundary layers; and well-mixed adiabatic core. (b) Vertical temperature section. The core of the fluid follows the liquid-adiabatic gradient $T_{ad}$ except for the thermal boundary layer at the top (greatly exaggerated in thickness). The liquidus gradient for the first appearance of a dense phase is shown as $T_{m}$, and intersects $T_{ad}$ at some distance off the bottom, dependent on the crystal formation rate. A very thin boundary layer forms at the bottom due to the strong convective drive from the settling of dense crystals, but should not perturb the temperature gradient. The magnitude of the temperature changes is indicated by the degree scale at the bottom. The gradient in the top insulating layer cannot be shown on this scale; it approximates a constant-flux boundary.

Substantially denser than their source components in the magma, so that crystallization always increases the density of a mixture, no matter what the component. If the latent heats of solution of these minerals into a basaltic melt were available (perhaps also substantially independent of melt composition), it would be possible to calculate Clausius–Clapeyron slopes for all the possible crystallizing phases. Bottinga & Richet (1978) have begun work on this problem, but as yet there are no such data, and I shall assume that there are no startling

Table 1. Basaltic crystals and their magma contributions.

<table>
<thead>
<tr>
<th>Name</th>
<th>Formula</th>
<th>System</th>
<th>Density (RT)</th>
<th>Density (1300°C)</th>
<th>Density (liquid)</th>
<th>$\alpha$(liquid) x10$^4$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Orthoclase</td>
<td>KAlSi$_2$O$_8$</td>
<td>Feldspar</td>
<td>2.55</td>
<td>2.46</td>
<td>2.27</td>
<td>6.2</td>
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<tr>
<td>Albite</td>
<td>NaAlSi$_2$O$_8$</td>
<td>Feldspar</td>
<td>2.62</td>
<td>2.52</td>
<td>2.32</td>
<td>4.1</td>
</tr>
<tr>
<td>Anorthite</td>
<td>CaAl$_2$Si$_2$O$_8$</td>
<td>Feldspar</td>
<td>2.76</td>
<td>2.71</td>
<td>2.58</td>
<td>4.1</td>
</tr>
<tr>
<td>Plagioclase (An75)</td>
<td>Mixture</td>
<td>Feldspar</td>
<td>2.72</td>
<td>2.67</td>
<td>2.52</td>
<td>4.1</td>
</tr>
<tr>
<td>Forsterite</td>
<td>Mg$_2$SiO$_4$</td>
<td>Olivine</td>
<td>3.21</td>
<td>3.04</td>
<td>2.80</td>
<td>11.6</td>
</tr>
<tr>
<td>Fayalite</td>
<td>Fe$_2$SiO$_4$</td>
<td>Olivine</td>
<td>4.39</td>
<td>4.22</td>
<td>4.10</td>
<td>8.2</td>
</tr>
<tr>
<td>Enstatite</td>
<td>MgSiO$_3$</td>
<td>Orthopyroxene</td>
<td>3.20</td>
<td>3.08</td>
<td>2.60</td>
<td>7.9</td>
</tr>
<tr>
<td>Clino-enstatite</td>
<td>Mg$_2$SiO$_4$</td>
<td>Clinopyroxene</td>
<td>3.19</td>
<td>3.06</td>
<td>2.60</td>
<td>7.9</td>
</tr>
<tr>
<td>Diopside</td>
<td>CaMgSi$_2$O$_6$</td>
<td>Clinopyroxene</td>
<td>3.28</td>
<td>3.15</td>
<td>2.64</td>
<td>7.6</td>
</tr>
<tr>
<td>Hedenbergite</td>
<td>CaFe$_2$Si$_3$O$_6$</td>
<td>Clinopyroxene</td>
<td>3.55</td>
<td>3.42</td>
<td>2.99</td>
<td>6.6</td>
</tr>
<tr>
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<td>Clinopyroxene</td>
<td>3.31</td>
<td>3.18</td>
<td>2.32</td>
<td>5.1</td>
</tr>
</tbody>
</table>

Reference magma density: semial 'isotropic gabbro': 2.60 5.6
stillwater complex, average: 2.57 5.2
anomalies, and that all the equilibrium slopes of temperature versus pressure are substantially similar to the liquidus slope of an arbitrary basalt (Fujii & Kushira 1977). I do not accept the interpretation of data from Green & Ringwood (1967) that implies a negative latent heat for plagioclase, since a conclusion so unreasonable physically is not warranted by the precision of the data.

Beside the possible complications due to differing latent heats among the various minerals, there are several aspects of convection in the presence of a suspendable phase that are bizarre enough. When a dense phase crystallizes and settles out, the residual liquid is much lighter than neighbouring fluid with suspended crystals, but is also compositionally similar to, or lighter than, the parent fluid before crystallization. Not only does it convect away from the bottom boundary, but, containing no crystals to absorb latent heat on remelting, it is buoyant relative to parent liquid all the way to the top of the chamber, thus leading to the uniformly vigorous convection of Fig. 3. If plagioclase is slightly denser than the magma, it can also settle out on the bottom boundary, but, unless another phase settles out to maintain a cotectic, the residual fluid is chemically denser than the parent magma. Like the settling drive calculated above, this chemical density increase is much greater than the thermal expansion of the latent heat applied to the residual fluid, and thus this fluid cannot rise to the top of the magma chamber. If the crystallization on the bottom were direct, due to the magma being denser initially than the plagioclase crystals, the residual fluid cannot leave the boundary at all. Since thermal diffusion is always faster than chemical diffusion, the excess temperature of the layer would be removed before it could mix, and it would be trapped at the boundary until it crystallized by heat leakage downward through the cumulates.

A bizarre process such as this could, conceivably, be called upon to explain the centimetre-scale layering often observed in cumulates. Even at the low suspended-crystal density of $3 \times 10^{-3}$, to an order of magnitude less, a centimetre layer of pyroxene can only have settled from 3 to 30 m of fluid; in no way can it represent a fluctuation in the composition of a magma chamber kilometres high. Moreover, there are commonly layers of essentially pure plagioclase in cumulates, from centimetres to hundreds of metres thick (e.g. McCallum, Raedeke & Mathez 1980). In my opinion these cannot form by the direct crystallization of a phase lighter than the magma, but must be able to settle out, even if very slowly. As long as convection is vigorous due to the regular settling of a dense phase, plagioclase can only settle out at a fraction of the rate of the dense phase - it does appear to be denser than at least the initial basaltic magma (Table 1), but only just. This means that it must accumulate in the fluid, as suspended crystals mostly in disequilibrium with a fluid-adiabatic temperature gradient, but always trying to cool the upper regions of the magma chamber while heating the lower part, in the process of moving toward crystal/liquid equilibrium.

If vigorous stirring should cease for any reason, the magma column with suspended plagioclase will approach the crystal-equilibrium temperature profile. This is steeper than the liquid-adiabatic profile, but not convectively unstable. The extra temperature rise is attained by the separation of more crystals with increasing depth, and this increases the density of the mixture (Table 1). Fluid from any stratum is stable against rapid vertical motion, since it would follow the liquid-only adiabat, whose differential effect on the liquid thermal expansion is less than the density effect of the different proportion of crystals. However, the column, though density stratified, is adiabatically mixable at rates slow enough for the crystals to maintain equilibrium. This can be seen by comparing the adiabatic relationship

$$\left( \frac{\partial T}{\partial P} \right)_S = \left( \frac{\partial V}{\partial S} \right)_P = \frac{\Delta V \times T}{L}$$

for melting of crystals, with the Clausius--Clapeyron relation $(\partial T/\partial P)_{	ext{eq}} = (\Delta V \times T)/L$; as

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must be expected, the gradients are the same. Suppose this peculiar 'slow convection' regime is in contact with a bottom boundary, and the densities of liquid and plagioclase are such that some slow settling can take place. Fluid that has lost some plagioclase can rise in the chamber until its remaining plagioclase has dissolved; above this point the liquid component becomes denser than the surrounding liquid (which continues to melt more plagioclase) and ceases to rise. Because this takes place in the bulk fluid, and the stratification height is dependent on just how much plagioclase was lost, this upwelling material should mix into the mass, as long as thermal-convective stirring from the top boundary is present. Here is a process that can produce thick plagioclase, or anorthosite, layers at the bottom of a magma chamber, even if settling velocities are extremely small. The simplified thermodynamic relationships are sketched in Fig. 4; the deposition of pure plagioclase requires that the temperature curve remain above the cotectic(s) where denser phases can reappear.

The process of deposition of an anorthosite accomplishes two things: the total quantity of excess plagioclase in suspension declines, and the continued loss of heat at the top of the chamber forces the residual liquid toward a cotectic. Eventually the temperature near the base of the chamber crosses the cotectic line and a dense phase can again be crystallized. Since it has not been present in the chamber for some time, a degree of supersaturation should be required for crystals to appear; once present and able to settle, the residual fluid is both thermally and chemically lighter than before and can rise directly to the top of the magma chamber, albeit at the 'slow convection' rate dictated by the suspended plagioclase content. The re-invigorated convection removes all fluid that has crystallized the dense phase from the basal region, and the nucleation process must begin again for the next batch of fluid. It seems to me that this is the right sort of process to produce the rhythmic banding observed in cumulates: capable of a periodicity, yet not constrained to be precisely regular.

The above scenarios are already quite complex, but assume a very simplified thermodynamics, a single cotectic versus pressure and excess plagioclase lines substantially parallel to it. In fact the Clausius–Clapeyron $\Delta V$s of both plagioclase and olivine are relatively small, while those for the pyroxenes are much larger. Unless the latent heat differences are

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**Figure 4.** Temperature section of a magma chamber depositing a pure anorthosite. The region of normal, liquid-adiabatic convection is now limited to the top of the magma chamber by the thick layer of suspended plagioclase. Weak convective drive in the lower zone allows the fluid to reach equilibrium with the suspended plagioclase crystals. It can still mix vertically at rates compatible with crystal equilibration.
proportional, the equilibrium slopes for the pyroxenes will be steeper (in a $T$ versus $P$ sense), and so pyroxenes are more likely to appear low in the chamber than plagioclase. Where the crossover (if any) in the loci of first appearance will occur depends on magma composition, including the content of minor elements and volatiles. A quasi-equilibrium situation is sketched in Fig. 5, where a pyroxene can be deposited on the bottom intermittently, but would be leached from the top unless that were coated with a protective layer of plagioclase.

Unfortunately, a complicated result, like the cumulate sequence in the Stillwater complex (McCallum et al. 1980) requires a complex set of processes to produce it. A preliminary exploration of the physics of the magma chamber should end by discussing briefly the well-known problem of increasing magma density with time, due to the concentration of iron into the liquid phase. It is true that volatiles are also concentrated into the liquid and must reduce its density. To decide whether the magma ever becomes too dense to deposit plagioclase during the formation of cumulates, it is instructive to simply think through the consequences of such a change.

If plagioclase deposition begins to lag behind the inexorable cooling rate for the chamber, crystals must begin to accumulate in the magma. This should raise the level of transition from normal to 'slow' convection (see Fig. 5) and reduces the temperature drop available for the upper thermal boundary layer, so that the boundary temperature must in fact drop below the liquidus of plagioclase. Hence the phase should crystallize on the top boundary, continuing vigorous convective drive for the chamber as a whole by leaving behind a chemically denser fluid. This type of convection has been found to be similar in form to purely thermally driven convection, especially at high Rayleigh numbers (Turner 1973). The phenomenon of increasing plume size with time after the onset of convection (e.g. Green & Foster 1975) is limited to modest Rayleigh numbers and systems where inertial turbulence is suppressed. At the high Rayleigh numbers induced by the scale of the magma chamber, there should be no change in the boundary-layer and mixed-bulk structure of the chamber as a whole. The chamber also remains plagioclase-rich, since the pyroxene equilibrium slope

Figure 5. Temperature section through a magma chamber depositing a banded zone. Lines of first appearance of plagioclase, olivine and pyroxene are shown in schematic fashion; the pressure equilibrium slopes are not known. The possibilities for instability are indicated by the liquid adiabat projected through the slow-convection zone; deposition of enough of a heavy phase could re-invigorate the convection with drive from the base. Return to overall fast, liquid-adiabatic convection could release large amounts of the heavy phases from solution and push the magma composition strongly into the plagioclase domain.
is so much steeper, but, unless these slopes can change significantly, the composition of the magma is buffered by pyroxene crystallizing near the bottom, and cannot be returned to a composition whose liquid fraction is light enough to allow plagioclase to settle efficiently. In fact, as the magma height decreases due to the deposition of the cumulates, the total plagioclase excess declines, leaving the fluid component near the bottom buffered by the appearance of a heavy phase. As the system is so well buffered, the main effect of increasing fluid density is to split the crystallization between the cumulates and the plated layer, with less and less plagioclase appearing on the bottom. As long as the crystal-equilibrium 'slow convection' dominates, fluid supersaturated in plagioclase should not reach the bottom, and direct crystallization there is unlikely. Possibly the pyroxene could nucleate on existing plagioclase crystals and drag them down.

It must be noted that convection driven by the settling of a dense phase is vigorous, and, if it reaches a certain threshold, could force the chamber back to fast, fluid-adiabatic convection. Perusal of Fig. 5 shows that return to a fluid-adiabatic temperature profile would liberate a great deal of crystallizable pyroxene and even olivine, in the process perhaps melting off some plagioclase at the top. The system might, under some circumstances, be self-triggering, but the reinjection of a hot, primitive magma would certainly trigger a return to the type of convection of Fig. 3. The effects of such a reinjection on an evolved magma chamber are thus much more dramatic than demonstrated by the laboratory experiments of Huppert, Turner & Sparks (1982). On the other hand, it is not necessary to postulate a reinjection for every olivine-containing layer, but merely for the major olivine-bearing zones (e.g. McCallum et al. 1980). This, again, seems to me much more likely than a 'special creation' version of events.

Conclusions

This paper has concentrated on applying simple physics to the processes involved in forming and crystallizing magma chambers, with sub-seafloor spreading principally in mind. Some of the results may not turn out to be correct, but at least the picture produced is self-consistent in all its aspects, and follows what is known about convection in large-scale systems, where both Rayleigh and Reynolds numbers are extremely high. The one consistent result of all experiments on such systems is that the body of the fluid is mixed and statistically uniform, with all the gradients being concentrated into extremely thin boundary layers. The thermodynamics is still complex enough to generate the layering observed in large continental igneous bodies, without resorting to unrealistic physics or special pleading. Perhaps the best way of summarizing the potentially controversial conclusions is to list them, starting inside the magma chamber and ending up with the plate tectonic aspects.

(1) Plagioclase cumulates are formed by slow settling of crystals, not direct crystallization of a lighter phase on the bottom of a magma chamber.

(2) The proportions of suspended dense phases in a crystallizing magma are extremely small, much less than 0.003 by weight; only plagioclase can accumulate to any significant degree in the fluid and should normally be limited to about 0.003 by weight.

(3) Presence of a suspended crystal phase can induce a peculiar form of convection whose motions are controlled by the rate at which the crystals can equilibrate with the fluid. This is named 'slow convection' and the temperature profile closely approximates crystal equilibrium with the changing pressure and composition of the liquid phase as a function of depth in the magma chamber.

(4) Rhythmic banding in cumulates is a natural result of the fluid dynamics, thermodynamics and crystallization dynamics in an evolved magma chamber bearing plagioclase in suspension. It does not require continued re-injection of primitive magma.
(5) The super-adiabatic convective drive is exceedingly small \( \frac{1}{3} \text{C km}^{-1} \) but convection is still much more vigorous than in the largest-scale laboratory convection experiments. Boundary layers are thin (less than 1 m), and the bulk fluid is well-stirred.

(6) The heat transport in a convecting magma chamber of any size is great enough to maintain the thermal gradient across a relatively thin insulating 'plated' layer (0.5 km) on the top and sides of the chamber.

(7) The crystallization lifetime of such a chamber is short \( (10^4 \text{ yr km height}^{-1}) \) and hence a magma chamber cannot be a steady-state feature on a submarine spreading ridge.

(8) The stress change in the global lithosphere, caused by the increase in local pressure as magma rises to the surface, is both large enough, and rapidly enough spread, to make room for a magma chamber in the early phases of its lifetime.

(9) Major ocean ridges spread intermittently and are substantially synchronized over their entire length.

Perhaps the last is the most startling prediction/conclusion, and warrants a few sentences of discussion. The Krafla volcanic system on Iceland contains a single, continuously fed, magma chamber that periodically injects dykes laterally along its segment of the Mid-Atlantic Ridge (Björnsson et al. 1979). This solves the thermal balance problem for a slow spreading ridge, but produces a crust consisting almost entirely of dykes, since cumulates can only form in a magma chamber, and that represents only a small part of its ridge segment. The Semiall ophiolite contains pervasive cumulates and therefore must have been formed from pervasive magma chambers. Fast-spreading ridges should therefore behave as suggested here — sudden spreading leading to a magma chamber all along the ridge, followed by active hydrothermal systems, followed by a long quiescent phase.

A supply-side limitation on the length of the quiescent phase may cause the different style of slow-spreading ridges. There is evidence that these have a periodicity too: from volcano-geological evidence in the long term (Ballard et al. 1979), and seismic evidence in the short term (Limon & Recq 1981), where a pulse of earthquake activity is seen to migrate along a spreading centre in a time very short compared to the magmatic cycle. The coordination of motion along a plate margin makes good sense mechanically, since any lack of synchronism among ridge segments would require movement along dormant fracture zones, and even these are not always well aligned with modern spreading directions.

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References


Crystallization mechanisms of sub-seafloor magma chambers


