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MOTION OF CRYSTALS, SOLUTE, AND HEAT IN LAYERED INTRUSIONS

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ABSTRACT

Crystals accumulate and grow at the floors of magma sheets, so the floor is a wet blanket of crystals plus liquid locked to the equilibrium temperature. Because this blanket impedes cooling through the floor, most cooling occurs at the roof. Both the removal of heat and growth of crystals are promoted by Grout's classic two-phase convection of crystal-bearing packets or plumes. Such suspension flows can carry crystals from roof to floor on a fast time scale, and may dominate the fluid dynamical regime. A region of crystals plus liquid that is compressed by sinking tends to become more supersaturated in the mafic phase. The growth of mafic minerals in cumulates releases a light, feldspar-enriched rejected solute (RS) which may facilitate adcumulus growth by rising in the gravity field. This light RS from mafic cumulates is hot but evolved and therefore poisons the overlying magma against nucleation. Feldspar-cotectic cumulates release dense RS that tends to drain away and stagnate at floors, but is evolved and so may dissolve its floor and rise, or be mixed by large-scale convection. When cumulus Fe–Ti oxides form, the RS becomes light, as in rare calc-alkaline or late-stage Fenner-trend layered intrusions, so oxide-bearing, feldspar-cotectic cumulates should have lower residual porosity than oxide-free ones. Compositional convection can be the agency of a wide variety of interesting phenomena in magmas. Planetary crusts may arise from collection of unmixed light RS released from growing mafic crystals in a magma ocean. At Rhum, the pore liquid transported contaminants with light feldspathic RS and yielded local Sr-isotope disequilibrium. The heat needed to melt the Rhum fingers was supplied by the crystallization of basal olivine and was transported through the magma by the heat-pumping action of compositional convection. Lateral migration of solute in porous cumulates may account for lateral growth of modal layering. Abundant xenoliths may cause rapid cooling and local storage of evolved magma. Both two-phase and compositional convection are important determinants of magmatic dynamic patterns.

Keywords: two-phase convection, compositional convection, cumulates, rejected solute, layered intrusions, magma.

SOMMAIRE

Les cristaux s'accumulent et croissent à la base des horizons de magma; on peut donc considérer cette zone comme nappe à assemblage cristaux + bain fondu conservé à la température de l'équilibre. Vu les propriétés isolantes de cette nappe, la perte de chaleur est plus importante de la partie supérieure de la nappe que de son plancher. La perte de chaleur et la croissance des cristaux sont favorisées par le modèle classique de Grout, c'est-à-dire, convection à deux phases des panaches ou fractions porteuses de cristaux. De telles épanchements en suspension peuvent effectuer un transfert très efficace de cristaux de la paroi supérieure vers le plancher, et pourrait prédominer comme mécanisme dans un régime de dynamique des fluides. Une zone de cristaux + liquide qui est comprimée par enfouissement a tendance à devenir plus sursaturée par rapport à la phase mafique. La croissance des cristaux mafiques dans les cumulats mène au rejet d'un soluté léger enrichi en composant feldspathique, ce qui pourrait favoriser une croissance adcumulus au cours d'une ascension dans le champs de gravité. Ce soluté issu des cumulats mafiques est chaud mais en même temps évolué, et peut donc entraver la nucléation dans le magma. Des cumulats cotectiques à plagioclase, par contre, libèrent un soluté plus dense qui a tendance à drainer vers le bas de la séquence, et à y stagner; toutefois, le soluté est évolué, et il peut donc dissoudre la base de la séquence et remonter, ou bien il pourrait se mélanger par convection importante. Quand il y a formation de cumulats d'oxydes Fe–Ti, le soluté rejeté devient léger, comme on en trouve (assez rarement) dans les complexes stratiformes calco-alcalins ou qui montrent une évolution de type Fenner. Les cumulats cotectiques à feldspath qui produisent des oxydes auraient donc une porosité résiduelle inférieure à ceux qui sont sans oxydes. La convection dite compositionnelle pourrait expliquer un grand nombre de phénomènes intéressants impliquant des magmas. Les croûtes des planètes pourraient résulter de la collection d'un soluté léger rejeté pendant la croissance des cristaux mafiques dans un océan magmatique. Dans le cas de Rhum, le liquide interstitiel transporte les contaminants avec un soluté rejeté léger et feldspathique, et produit un déséquilibre local en isotopes de strontium. La chaleur requise pour dissoudre les interdigitations à Rhum a été produite par la cristallisation de l'olivine à la base et transportée à travers la colonne magmatique par l'action de pompe à chaleur du mécanisme de convection compositionnelle. Une migration latérale du soluté dans les cumulats poreux pourrait expliquer la croissance latérale d'une stratification modale. Une abondance d'enclaves pourrait tremper le système et causer un stockage local du magma évolué. Les deux genres de convection, à deux phases et compositionnelle, exerceraient d'importantes influences sur l'évolution dynamique des magmas.

(Traduit par la Rédaction)

Mots-clés: convection à deux phases, convection compositionnelle, cumulats, soluté rejeté, complexe intrusif stratifié, magma.

INTRODUCTION

Types of convection

Three types of convection are of particular importance in the magma bodies that become layered intrusions. These types are thermal convection, two-phase convection (Grout 1918), and compositional convection due to the rejection of solute by growing cumulates (Morse 1969). Each of these processes is limited by the nature of the principal heat sink, generally conduction through the roof or, more rarely, melting of the roof (Irvine 1970). Two-phase convection involves crystal + liquid suspensions that can move much faster than small individual crystals can settle. The crystals, in turn, can nucleate and grow only as fast as the requisite supercooling can be obtained in the boundary-layer near them. The standard causal cycle is: heat loss through the roof, nucleation, two-phase convection, transport of the two-phase packet or plume to the floor with its attendant supercooling, growth of crystals at the floor, release of light or dense solute, and either compositional convection or stagnation, modified according to the slope of the floor. Other regimes and scenarios are considered by Brandeis & Jaupart (1986), but the one just described corresponds in major respects to their case for common mafic magmas.

Recent work and scope of the present study

Some of the effects of two-phase and compositional convection in relation to adcumulus growth are explored elsewhere (Morse 1986a,b). In these studies and another by McBirney (1985), it was shown that double-diffusive convection (Huppert & Sparks 1984) cannot arise from the adcumulus growth of floor cumulates. The reason for this is that the compositional contributions to the buoyancy of magma near the cumulate interface are too large to be balanced by the heat effects due to crystallization. In the case of ultramafic cumulates, the rejected solute is light and rises unimpeded through the lower thermal boundary layer and the overlying main magma, so no double diffusion can arise for any ultramafic cumulate. In the case of felsic cumulates at the floor, the dense rejected solute tends to pond and no plausible thermal gradient can strongly oppose this tendency. Adcumulus growth of such cumulates requires diffusion or megascopic convective stirring to renew the composition of the boundary layer; hence, felsic adcumulates ordinarily betoken low accumulation rates, as long before supposed.

Infiltration metasomatism does not produce adcumulates as proposed by Irvine (1980), because it is intrinsically an orthocumulus (*i.e.*, liquid-trapping) rather than an adcumulus process. The rising rejected solute from ultramafic cumulates is trapped in a thermal minimum within the newly added layer, where it reacts with the local cumulus minerals. The type examples at Muskox are in fact mesocumulates, testifying to unusually high rates of accumulation that prohibited full adcumulus growth even in the most favorable case of ultramafic floor cumulates.

Compaction (Sparks *et al.* 1985) was probably not a major agent in reducing porosity in the cumulates of most layered intrusions, because the depth of crystal mush is rarely indicated to have been great enough (1–10 meters) to make compaction an interesting possibility. For example, the olivine load-structures emplaced into plagioclase mush described at Fongen–Hyllingen by Thy & Wilson (1980) suggest a mush depth of less than 10 cm. The scale of similar loading features at Rhum (Butcher *et al.* 1985) suggests mush depths commonly less than 1 m. Both intrusions were multiply replenished and hence should have offered favorable chances for rapid deposition of thick cumulates, but evidence for these is not observed. Muskox, where the accumulation rate may have been one of the highest, does not show mafic adcumulates, so compaction was evidently not an important agent of adcumulus growth there.

The earlier results are extended here to a consideration of the large-scale flow fields in mafic magmas, the effects of compression on phase equilibria and nucleation, the non-local behavior of buoyant and dense rejected solutes, the influence of quasi-crystalline melt structure, and the mechanism of magmatic heat-pumping.

THERMAL CONVECTION

Geometry and cooling behavior

Magma bodies having a sufficiently large ratio of diameter to height (d/Z) cool like sheets and have marginal border zones that are minor relative to their layered groups. The critical value of d/Z for sheet cooling can be evaluated from known layered intrusions, when their shapes are approximated by the nearest cylinder of equivalent volume and surface area. For a cooling sheet, the magma depth (r) should decrease from the initial depth (r_0) according to the square root of time, $t^{1/2}$ (Carslaw & Jaeger 1959, Irvine 1970, Norton & Taylor 1979). The original geometry of the Kiglapait intrusion is amenable to reconstruction (Morse 1969, 1979a) and has been subdivided and contoured in levels of percent solidified (PCS) representing the upward advance of the cumulate interface through time. For the lowermost 99.8 PCS of the intrusion, it is found (Fig. 1) that $r = r_0 F_L^{1/2}$, where $F_L = 1-(PCS/100)$. From this it is inferred that the body cooled like a sheet during this time interval, and that $F_L \sim t$. After 99.8 PCS, the slope of the relation evidently steepens from ½ to greater than 1.0, consistent with a change to volume cooling rather than sheet cooling (Morse 1979a).

For the geometry of the Kiglapait magma chamber (Morse 1969) the mean depth of magma is 6.3 km, and the aspect ratio d/Z is about 4. The aspect ratio of the Skaergaard Intrusion (Fig. 6 of Norton & Taylor 1979), for which sheet cooling is also calculated by these authors, is about 2. Since the marginal border zones of this body are proportionately among the largest encountered, the critical lower limit of d/Z for sheet cooling may be about 2. Any magma body likely to be described as flatter than subequant will probably undergo sheet cooling, to a first approximation at least. For such bodies, the geometry will have little effect on the overall cooling history or the flow regime (Marsh 1985).

The details of sheet cooling will vary with d/Z. Single-cell systems are more likely to occur near the smaller values of d/Z, whereas multi-cell or Rayleigh–Benard systems are more likely at higher values of d/Z. Brandeis & Jaupart (1986) limit their discussion to the case of a flat floor and consider that the flow field is dominated by down-going plumes with no return flow, but it is not clear that this will be the case for natural magmas when all the vagaries of compositional convection and chamber geometry are considered. In particular, their assumption of a strictly stagnant basal boundary-layer is difficult to reconcile with observed mineral orientations consistent with confocal basal flow (Belkin 1983) and with the requirements of local nucleation and adcumulus growth (Morse 1986b).

Major differences arise in the behavior of highly mafic or ultramafic magmas as opposed to felsic ones. Such differences include the higher temperatures and lower viscosities of mafic magmas. The latter are due not so much to the high temperature as to the simpler melt structure of mafic magmas, which are rich in network-modifying as opposed to network-forming components. The differences in cooling behavior center on whether or not an ultramafic cumulate is laid down at the floor, in which case vigorous two-phase convection may occur in company with vigorous stirring by compositional convection. Such an effect will promote melting at the roof and rapid crystallization, as in the Muskox Intrusion (Irvine 1970). More felsic bodies, on the other hand, will tend to form a felsic or cotectic upper border zone and thus tend toward slow crystallization. Such bodies occur in abundance in anorthosite massifs such as Nain (Morse 1983a).

Onset of convection

As long ago noted by Grout (1918), thermal con-

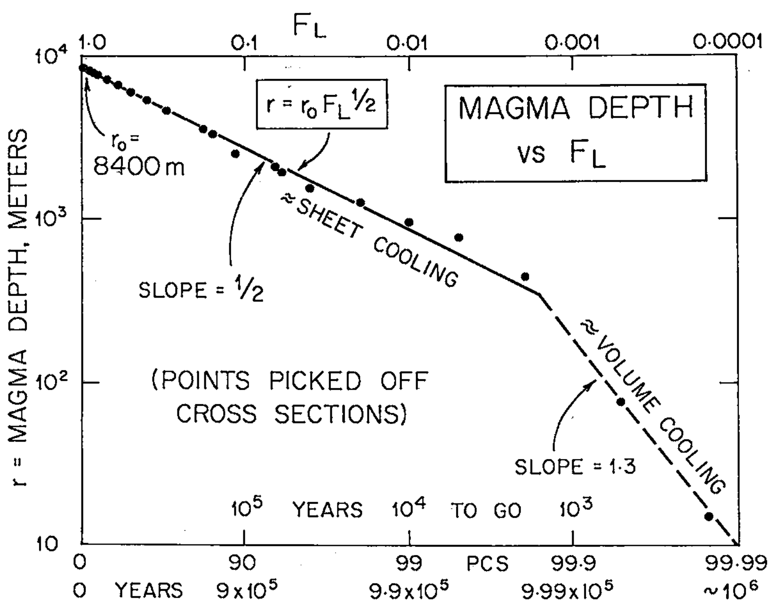


FIG. 1. Relation between magma depth and percent solidified (PCS) in the Kiglapait Intrusion. The upper horizontal scale is the fraction liquid $F_L = 1 - (PCS/100)$. The inference that most of the cooling history is like that of a sheet derives from the expectation that for a sheet, the magma depth should vary as the square root of time, $t^{1/2}$, hence the observed relation $r = r_0 F_L^{1/2}$ implies that $F_L \sim t$. The crystallization progress and age are indicated on the lower horizontal scales, based on the estimated crystallization time of 10^6 yr (see Morse 1979a).

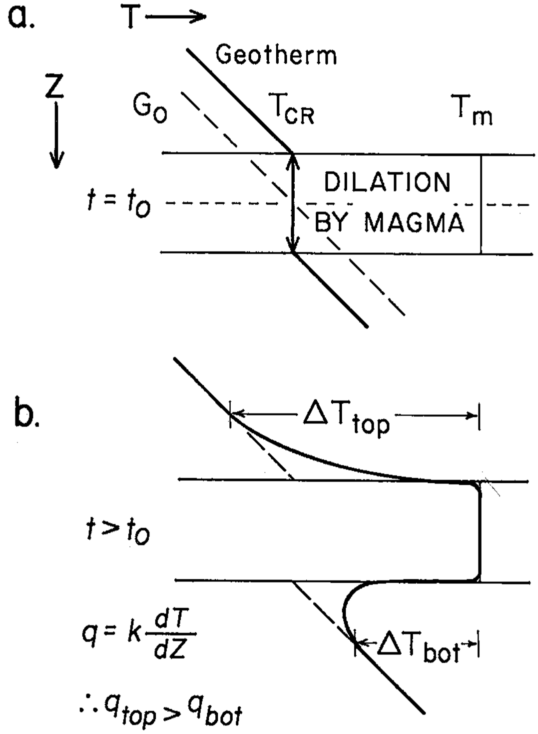


FIG. 2. First megascopic basis for convection. The crust is dilated by emplacement of magma at time $t = t_0$ and temperature $T = T_m$, and the geotherm is thereby offset from G_0. Thereafter, the thermal gradient and hence the heat flux at the top progressively exceed those at the bottom. Roof cooling occurs and becomes dominant.

vection begins when the magma splits the geotherm (Fig. 2a). At a given reference distance some time later (Fig. 2b), the thermal contrast is larger above a cooling sheet than below, and hence the heat flux is greater through the roof than through the floor. To this effect may be added that of a choked feeder (acting as a stove) and a blanket of floor cumulate (Irvine 1970). This must remain at the equilibrium (XL + LIQ) temperature as long as trapped liquid remains, so a high porosity impedes further abstraction of heat from the magma to the floor rocks. Orthocumulates and mesocumulates are common near the bases of layered intrusions (*e.g.*, Raedeke & McCallum 1984). When such floor cumulates containing trapped liquid reach thicknesses of $\sim 10^2$ m it can be shown that they have sufficiently long lifetimes to make floor cooling of small importance (*e.g.*, Hess 1972). The term *substrate cooling* is usefully applied to cooling through the floor (for floor cumulates) or through the roof (for roof cumulates).

A contrasting pathway for cooling is the indirect route from the floor via the magma and out through the roof, or in the case of roof cumulates, out through another part of the roof or walls. This pathway has been called *remote cooling* (Morse 1982). Adcumulus growth requires remote cooling (Morse 1969, 1982, 1986b) rather than substrate cooling, and therefore requires a local thermal profile like that in Figure 3C. Here the liquidus or cotectic temperature (T_e) joins the adiabat near the floor, and the adiabat is perturbed near the roof to rejoin T_e there. Thermal boundary layers occur at both levels. The upper boundary-layer represents the substrate cooling of the magma through the roof. This layer is maintained relatively thin by crystal nucleation and intermittent two-phase flow of denser packets of magma away from the roof area (Grout 1918, Wager & Deer 1939, Hess 1960, Brandeis & Jaupart 1986). These packets need not be rich in crystals in order to detach and flow, so they may retain some of their initial supercooling (as well as growing crystals). As they sink and compress, they tend to gain supercooling as the adiabat departs ever more from the equilibrium temperature. If the rate of descent is large enough relative to the rate of crystal growth, supercooling will be delivered to the floor, generating the deeper boundary-layer shown in the detail of Figure 3C. These thermal structures are discussed further by Morse (1982, 1986a,b). Their most critical feature is the presence of a "latent heat hump" at the floor, where the population density of growing crystals is greatest. This thermal maximum at the cumulate interface is a barrier to floor cooling of the main body of magma.

A thermal minimum (the supercooled boundary-layer) occurs for all felsic cumulates at the floor, but may not always be present above ultramafic cumulates whose latent heat is being swiftly conveyed by compositional convection upward to a heat absorber (Morse 1986c). Such a transport mechanism will tend to keep the magma at the adiabat.

Nucleation

A critical feature of crystal nucleation is that appreciable supercooling (typically 5–10°C for minerals in basic magmas) is needed in order for nucleation to occur, whereas finite growth occurs at infinitesimal supercoolings (Dowty 1980, Kirkpatrick 1983, Brandeis *et al.* 1984). A formal distinction is drawn between homogeneous nucleation, in which nuclei are formed independently of each other or of any internal surface, and heterogeneous nucleation, involving pre-existing objects as substrates. Such objects may be as obvious as wallrocks or older crystals, or as subtle as shear planes in the liquid. The heterogeneous process can be treated theoretically like the homogeneous process by allowing for a smaller critical value of supercooling needed for the onset of nucleation (Dowty 1980, Brandeis *et al.* 1984). Homogeneous nucleation is difficult if not impossible to achieve in the laboratory because a substrate is generally present. However, it is observed that heterogeneous nucleation is less frequent when the volume of melt is large compared to the surface area of contact (Dowty 1980). Criteria for judging the relative importance of the two mechanisms of nucleation are not well developed for plutonic magmas. Dowty (1980) cites the need for textural studies and grain-size analysis in this connection, some of which are supplied by Brandeis *et al.* (1984). These authors found a good correspondence among theory, experiment, and natural crystal sizes when the theory was based on homogeneous nucleation combined with minimum supercoolings taken from laboratory studies, so as to use parameters adequate for heterogeneous nucleation. If true homogeneous nucleation occurs instead, the critical supercooling needed for nucleation will be larger.

The conclusions reached in this paper are not materially affected by the exact choice of critical supercooling used. Despite this practical convenience, there is some textural evidence in layered intrusions that may bear on whether nucleation is largely homogeneous or heterogeneous. The common presence of independently zoned grains and other textural features of igneous cumulates suggests the dominance of homogeneous nucleation in large magma bodies. For example, the fine-grained, granular texture of the Upper Border Zone (UBZ) of the Kiglapait Intrusion (Morse 1969) is accompanied by abundant normal zoning of feldspars and suggests independent nucleation of randomly dispersed crystals. By contrast, crescumulate growth (Wager & Brown 1967) attests to at least one form of heterogeneous nucleation, and perhaps to the conditions favoring heterogeneous nucleation in general.

The population density of crystal nuclei formed depends upon the amount of supercooling beyond the critical value needed for any nucleation. At a very small increment of supercooling, the population density is very low, and that is the condition applying to large, aging magma bodies averaged over time and volume. For example, the average population of crystals in the Kiglapait Intrusion is estimated to have been about one crystal per cubic meter or ~ 1 ppm (Morse 1979a). The density of crystal nuclei in cooling magma rises dramatically as supercooling increases, however, to a maximum near 25,000/cm^3/hr at 30°C supercooling (Brandeis *et al.* 1984). Such supercoolings are likely to occur, at least locally, near the roof of a magma body. They may also occur intermittently if portions of the roof occasionally founder, as they appear to have done in the Kiglapait Intrusion. Evidence for such foundering is given by units with the fine-grained UBZ

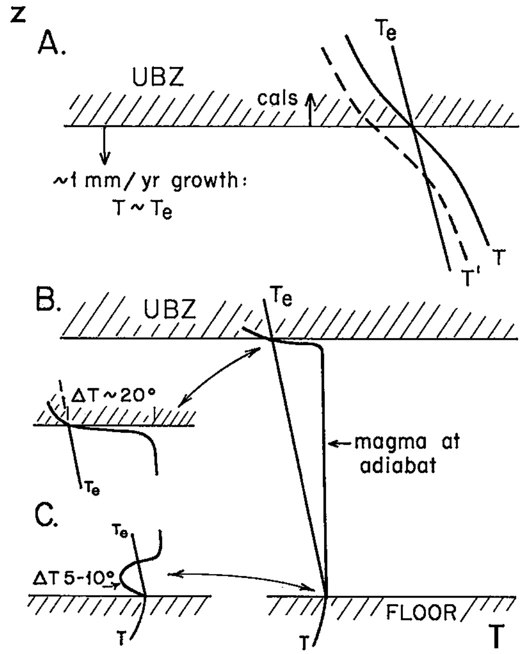


FIG. 3. Thermal regimes in a large layered intrusion behaving as a sheet with an upper border zone (UBZ). A. Thermal profile near the UBZ interface; T is the steady-state average profile with no supercooling at the interface, and T' is a perturbed profile leading to chilling. The growth rate of the UBZ is that estimated from the Kiglapait Intrusion (Morse 1979a). B. Whole thickness of magma with cotectic gradient T_e and the magma temperature at the adiabat except for a thermal boundary layer near the UBZ. C. Closeup of thermal boundary layers at roof and floor, showing the strong gradient near the roof and the undercooling near the floor. The latent heat hump is the thermal maximum at the top of the floor.

texture interlayered with floor cumulates (Morse 1969), and by rare UBZ rocks showing quench textures (Morse 1982). When sufficient supercooling does occur, the large number of nuclei produced can increase the two-phase density rapidly even though individual crystals remain very small, and hence intermittent two-phase convection is likely to play an important role in the patterns of heat and mass transfer (Morse 1986a).

An important feature of common mafic magmas is that mafic minerals whose structures are based on network-modifying units (such as independent tetrahedra) nucleate with relative ease, whereas those richer in network-forming units nucleate with difficulty. The *barrier to nucleation* decreases in the order: plagioclase >> pyroxene > olivine > spinel (Wager 1959, Morse 1986a and references therein). This means that mafic cumulates may originate with relatively little supercooling, whereas the complementary felsic cumulates may be inhibited from forming. Furthermore, after nucleation the growth of olivine may drive the local liquid metastably into the primary phase field of pyroxene or plagioclase.

PT cycles

Any periodically circulating magma undergoes cyclical excursions in pressure and temperature, or *PT* cycles. Depending on the nucleation, growth, detachment, entrainment, or resorbtion of crystals, such cycles may become very complicated. Nevertheless, an idealized history is worth stating, for the insights it may bring. The simplified *PT* cycle of a package of mafic magma is shown in Figure 4 as an irreversible stepwise process. Magma is cooled along the roof (1), transferred downward adiabatically to (2), yields mafic crystals, and locally absorbs heat from the growing crystals to reach the cotectic tem-

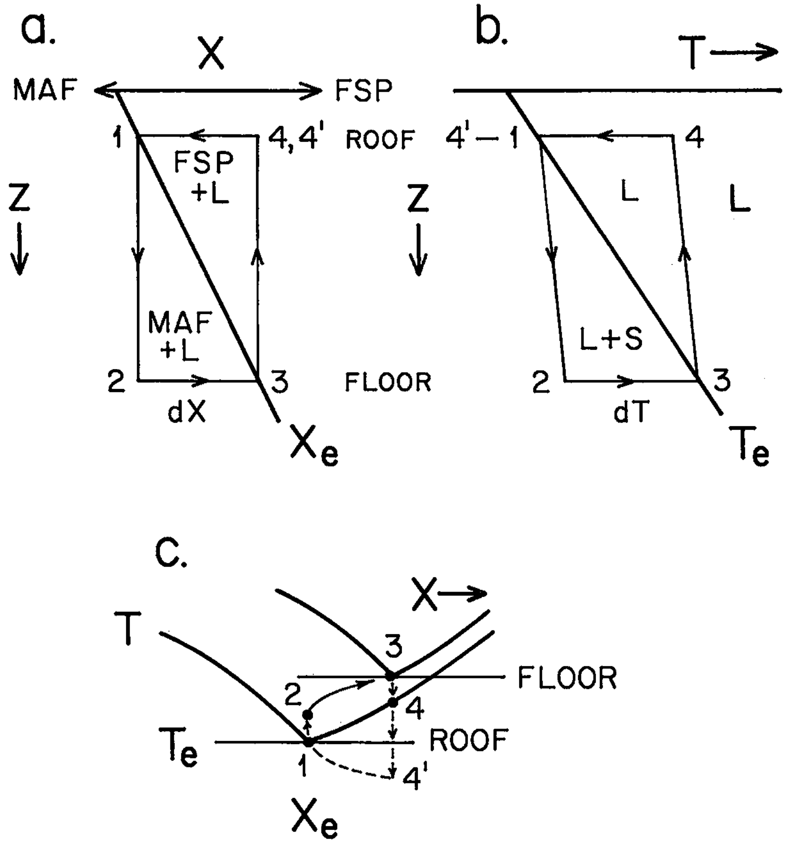


FIG. 4. Composition (a) and temperature (b) *versus* magma depth Z. In (a) the locus X_e is the trace of the generalized feldspar–mafic cotectic shown as points 1 and 3 in (c). In (b), the locus T_e is the trace of the equilibrium temperature at the cotectic. The history of a packet of magma is illustrated by an idealized *PT* cycle starting at (1). The steps in this cycle are (1–2) isocompositional adiabatic compression in the field of mafics; (2–3) nucleation and growth of mafics with evolution of latent heat returning the liquid to the saturation curve; (3–4) isocompositional adiabatic expansion in the field of feldspar; (4–1) strong roof cooling returning the liquid to the cosaturation temperature and eventual nucleation and growth of feldspar. The *T-X* diagram (c) at two different pressures shows that step (4–1) is likely to overshoot to 4′ because the feldspar liquidus temperature changes relatively little in the pressure range of the earth's upper crust, and greater supersaturation is needed for the nucleation of plagioclase than for mafics because of its high barrier to nucleation. Although the mafic and felsic crystallization events are restricted to floor and roof, respectively, in the *PT* cycle, it is expected that in nature, mafic crystallization deforms the path (1–2) and growth of mafics at the floor induces nucleation of plagioclase also at the floor.

perature (3); if it rises it does so into the field of feldspar (4). The feldspar may fail to nucleate, or if it does, the supercooling is likely to be so great that mafic crystals nucleate also, causing the entire packet of magma to sink. The bottom part of the Figure shows the T–X relationships corresponding to the upper diagrams, and emphasizes the possibility that considerable undercooling will ordinarily be required for the nucleation of plagioclase. The Figure can be studied further with the aid of the caption.

The T–X details of nucleation and growth in relation to compression are amplified in Figure 5, which shows the combined thermal–compositional trajectories resulting when the latent heat of crystallization is retained within the local packet of magma in which crystals nucleate and grow. The effects of pressure on melting and melt compositions are summarized in Morse (1980); values of the latent heat of fusion are taken from Ghiorso & Carmichael (1980). In the illustrated case a cotectic parcel of troctolitic magma at low pressure has been adiabatically compressed to (2) while the equilibrium cotectic and its attached field boundaries have migrated to (1) by compression. Nucleation and growth of olivine from liquid (2) will heat the liquid along (2–3), a trajectory whose slope (calculated with the use of the Lever Rule to reckon the compositional distance travelled) is fortuitously indistinguishable from that of the observed plagioclase liquidus. Because of this near-parallelism, the growth of olivine in a thermally isolated system will not appreciably increase the supersaturation of plagioclase represented by distance below the liquidus at constant pressure. This is so even when the liquid has reached the limit of olivine growth (3) at the metastable olivine liquidus. Supposing by contrast that plagioclase were to nucleate and grow at either (2) or (3), it would heat the liquid along such a low slope that rapid supersaturation and nucleation of olivine would ensue, and the limit of plagioclase growth would undoubtedly never be reached. Further compression of the magma will increase the distance between (1) and (2) along a steep trajectory of 13°C for each increment of 1 wt.%. The *in situ* nucleation of plagioclase seems best able to account for the presence of this potentially buoyant mineral in floor cumulates (Maaløe 1978, Morse 1979a). In order for the nucleation of plagioclase to occur, a substantial supersaturation must be present. In principle, this can be attained by a compositional shift of the liquid toward plagioclase rather than by supercooling. The compositional change can occur by local storing of the rejected solute from the growth of olivine. In view of the preceding analysis, however, this compositional effect would not effectively increase the local supersaturation in plagioclase unless the latent heat of fusion released by the growing olivine is extracted from the local system, as shown by the horizontal

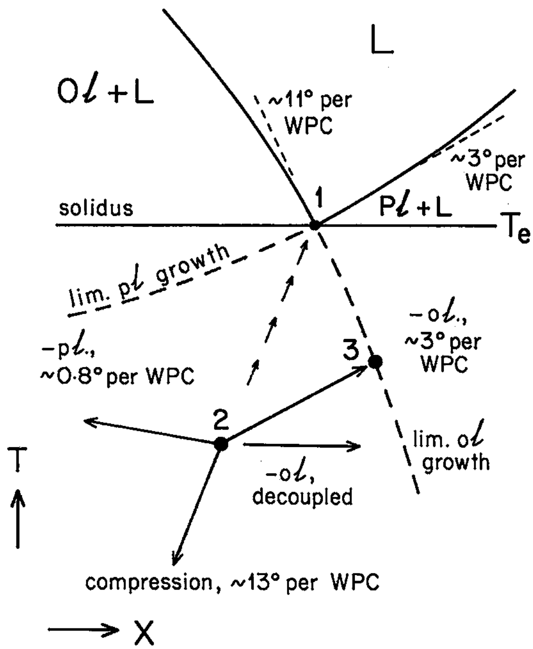


FIG. 5. Relative T–X motion of liquid and cotectic (olivine ~ 0.3, plagioclase ~ 0.7 by weight) with pressure and crystal growth. At the cotectic, the slopes of the olivine and plagioclase liquidi are about 11 and 3°C per weight percent composition, respectively (based on the system Fo–An; Osborn & Tait 1952). Their subsolidus extensions define the limits of crystal growth (*e.g.*, Kirkpatrick 1983). A low-pressure cotectic liquid (2) devoid of nuclei is compressed while the equilibrium cotectic (1) moves away along the small arrows; this amounts to an arbitrary supercooling and composition change along path (1–2). Olivine may then nucleate. Growth of 1 gram of olivine releases heat sufficient to raise 100 g liquid by about 3°C for every weight percent change of the liquid composition, using the Lever Rule (path 2–3). Little change occurs in the relative supersaturation of plagioclase because the heating curve is subparallel to the plagioclase liquidus. The limit of olivine growth is reached at (3). If nucleation of plagioclase occurred instead, the resulting heating curve leftward from (2) would have slope ~0.8°C per weight percent change in composition, thereby increasing the supersaturation with respect to the olivine liquidus. Further compression separates the liquid (2) from the cotectic along the steeper path shown, as the cotectic moves in the direction of the small arrows. If heat is continually removed during the growth of olivine, plagioclase supersaturation can be increased (horizontal arrow), as the latent heat is decoupled from the compositional effect.

"decoupled" path in Figure 5. This cooling is permitted by the ratio of thermal to mass diffusivities ($k_T/k_c \sim 10^5$; Huppert & Sparks 1980) and by the ratio of thermal to mass *fluxes* ($q_T/q_c \sim 300$) during adcumulus growth (Morse 1986b). This horizontal path turns out to be that of the quasi-isothermal supersaturation of plagioclase induced by the growth of olivine contemplated by Morse (1979a), except here more correctly removed downward into the supercooled region below T_e (Maaløe 1978).

By mentally translating (2) in Figure 5 to the right, one obtains the case of a feldspar-enriched liquid such as (4) in Figure 4, which can be brought into the growth field of olivine as the metastable liquidus rises past it upon compression. Thereafter the liquid can be chased back into the plagioclase field by olivine extraction.

Accelerations

Two-phase convection, combined with crystal growth due to compression, can be a significant accelerator of magma flow. Such action will augment any intrinsic tendency for three-dimensional convection to have an axisymmetric plume combined with downward sheet-flow (Loper 1985). This is the situation depicted in Figure 6 as a possible case for a large class of layered intrusions having d/Z perhaps between 10 and 2. The configuration is one of toroidal convection, and its stability would be enhanced by convergent basal flow and nucleation induced near the distal ends of the roof (positions 3–4), augmented by growth during compression along walls and sloping floor.

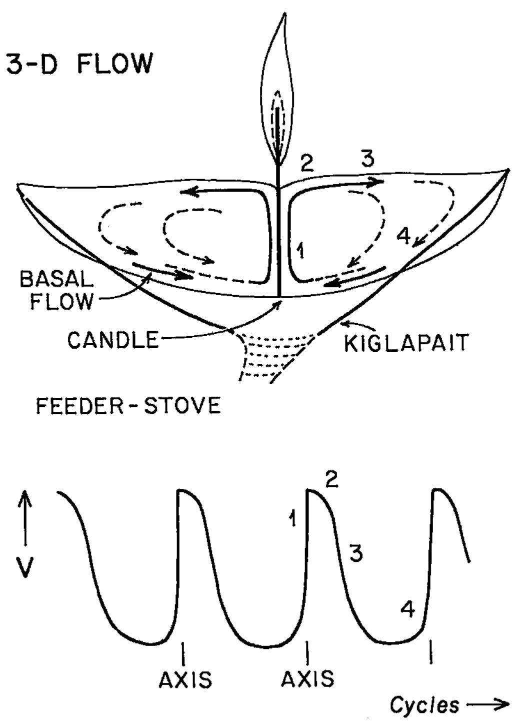


FIG. 6. The candle analogy illustrates accelerations in three-dimensional (toroidal) convection. The schematic velocity (V) spikes show strong accelerations near the wick (central axis of the toroidal flow) and decelerations during the outward radial transit of observable suspended particles. In the magmatic analog the driving force is the distal sinking of crystal-bearing magma packets and basal downslope flow rather than the axial heater, but the two effects are geometrically equivalent in that the axis is hot and light relative to the sheath.

Any such toroidal flow will tend to be maintained by downward accelerations. For a given steady state the stream discharge (Q) is that of the central column, and for the convergent basal flow at constant Q, there must either be a large acceleration or else a constant cross-sectional area (A) maintained by a large increase in current depth, since Q = AV. A helpful example of toroidal accelerations may be found in the behavior of small particles of soot entrained in molten candle wax (Fig. 6). The particles are nearly immobile near the distal edge of the pool (at 4), then begin to drift lazily toward the wick along the convergent floor, then suddenly whip up along the wick (1–2) and flip outward along the top, rapidly decelerating (2–3). Magma bodies lack wicks, but they may have axisymmetric plumes and downward sheet-flow along the margins that will cause analogous behavior. Given such relative accelerations in magma, the transit (2–3) may be so rapid that little cooling occurs and the major cooling should occur in the more distal portions (*e.g.*, along path 3–4 in Fig. 6) of the radially divergent flow. Most nucleation should therefore occur in an annulus.

McBirney & Noyes (1979) suggested that the sinking of particles in toroidal flow would be symmetrical in terms of radial distance travelled in floor and roof regions. This would be so only for particles of constant size, originating and depositing at places equidistant from the central axis. It appears instead that the nucleation of crystals and the consequent two-phase sinking will be highly asymmetrical with respect to the central axis, for example, depositing crystals along the path 4–1 that do not ordinarily exist in the roof counterpart, path 2–3, of the flow (Fig. 6).

A whole-body convective flow would presumably have the property of a graded stream in being just able to carry its load, hence energetically buffered with respect to central holes or piles of crystals. If a hole develops, the flow accelerates and more particles are carried in; if a pile develops, the flow decelerates and particles are dropped earlier. Examples of approximately convergent basal flow are well documented from Skaergaard as trough bands (Wager & Deer 1939, Irvine 1980) and as oriented olivine crystals (Brothers 1964). In the less vigorous Kiglapait, they are implied by the downslope lineation of subhedral olivine crystals observed in the plane of lamination (Belkin 1983). Base flow in density currents has also been suggested as a mechanism for transporting and depositing buoyant plagioclase (Irvine 1978).

Compositional Convection

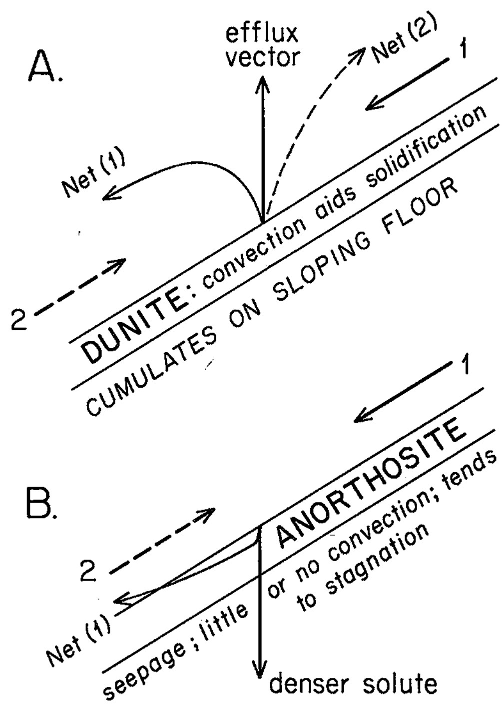
The limiting cases of dunite and anorthosite layers (Fig. 7) serve well to illustrate the principles of convection due to compositional change (Morse 1969, Tait *et al.* 1984, Kerr & Tait 1986). Ultramafic cumulates such as dunites, whose rejected solute is lighter than the main magma, are aided in their adcumulus solidification by the buoyant rise of the felsic component counterbalanced by an influx of the refractory components of olivine. Felsic layers may solidify more slowly because their rejected solute is dense and tends to stagnate and fractionate in place.

Mafic cumulates

Mafic and ultramafic cumulates tend to occur at the bases of layered intrusions and at the bases of cyclic units where replenishment of the magma chamber has occurred. The light rejected solute from these cumulates has approximately the temperature of the latent heat hump at the cumulate interface, but it is evolved, most notably in its content of feldspar component and incompatible elements. It therefore convects freely (Kerr & Tait 1986) through the overlying magma and adds to its heat content, carrying away some of the latent heat produced by the mafic cumulate. If mixed into the main magma, it causes the latter to evolve without cooling appreciably. If a mafic crystal is dropped into this magma, it will dissolve. The magma is therefore poisoned against spontaneous nucleation of mafic crystals or their survival if they are carried in. Therefore, the interior of a thick layer of picritic magma is most likely to be free of any crystals rather than charged with them, as assumed by Huppert & Sparks (1980).

Reduced cotectic cumulates

The average (cotectic) troctolite of the Kiglapait Intrusion is relatively felsic and would reject a dense solute (Morse 1969) leading to stagnation on a flat floor, or downward flow on a sloping floor. If a dense solute is rejected from average troctolite, the residual magma must increase in density. That this does indeed happen is guaranteed by the presence in the magma of the components of dense phases such as augite, Fe–Ti oxides, and sulfides, which eventually appear in the crystalline products. The well-known Fenner trend of iron enrichment shown by many basaltic magmas in a closed system also contributes to the rise in density. The magma density calculated for the Kiglapait Intrusion (Morse 1979a) increased throughout the deposition of the Lower Zone and lower parts of the Upper Zone, and reached a maximum only at the Main Ore Band (93.5 PCS; Morse 1980). The SiO_2 content of the calculated liquid (Morse 1981) is minimum (43.33%) nearby at 92.5 PCS. In the Skaergaard Intrusion (Wager & Brown 1967), the maximum in normative oxide minerals occurs for the MZ rock, which is also near the minimum silica content of the estimated liquid. The calculations of Bottinga & Weill (1970) show a progressive increase in density through the MZ liquid. From these two examples it can be concluded that the average rejected solute is dense over most of the history (>90%) of Fenner-trend layered intrusions and reaches a maximum in magma den-



(B is the general case for cotectic gabbros, norites, and troctolites.)

FIG. 7. Density contrast of rejected solute compared to main magma in the adcumulus solidification of (A) dunite and (B) anorthosite layers, as described by Morse (1969). Solidification of dunite from troctolitic magma releases buoyant feldspar-rich material tending to rise vertically, as shown by the arrow labelled "efflux vector". Downflowing currents (1) will bend this vector downslope, whereas upwelling currents (2) would bend it upslope. The buoyancy of the rejected solute favors rapid solidification of dunite because diffusion is greatly aided by convection. In the case of anorthosite (B), the rejected solute is dense and can escape only downward; if it seeps through crystal mush it will be impeded, depending upon the angle of repose and the permeability.

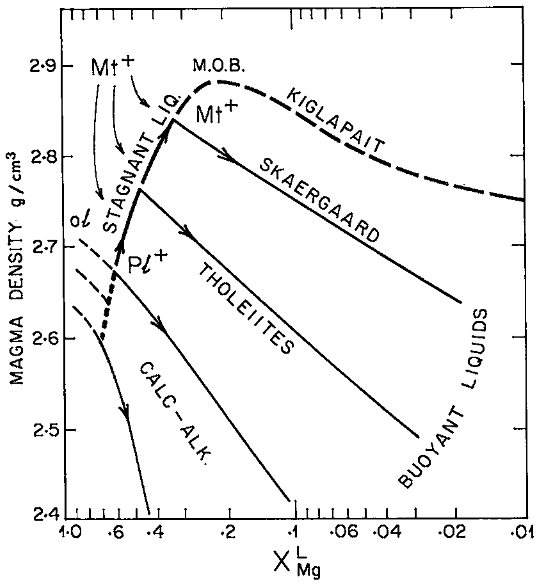


FIG. 8. Behavior of magmatic-liquid density with magnesium ratio for magma series ranging from most reduced (Kiglapait) to highly oxidized (Shasta calc-alkaline trend). The two limiting trends and the Skaergaard trend are plotted as calculated (Morse 1979a, Bottinga & Weill 1970); the others are schematic. A trend of increasing density begins, for reduced magmas, when plagioclase appears, and ends at a density maximum when magnetite appears. The calc-alkaline trends have magnetite on the liquidus early in their history, and therefore begin at the density maximum. The density trends are directly related to the FMA trends of Jakeš & White (1972), as shown in Figure 9. Key: Mt magnetite, Ol olivine, Pl plagioclase, MOB main ore band.

sity once cumulus Fe–Ti oxide minerals occur in abundance. Examination of the cotectic ratios and ferrodioritic daughter products of norites (Ranson 1981, Morse 1982, Wiebe 1979, 1984) leads to a similar conclusion for these rocks. Therefore, the average cumulates of all members of the anorthosite–norite–troctolite (ANT) suite will have a rejected solute that is dense.

Gabbroic liquids saturated with augite and plagioclase fractionate toward saturation with olivine or orthopyroxene and, therefore, move toward denser compositions. The Hat Creek basalt (Anderson *et al.* 1982) reaches a maximum femic index (FI = 100 CIPW *fem* = 47) at 23 PCS and a maximum in CIPW (*mt* + *il*) of 11.3 at 76 PCS, indicating that a density maximum probably occurs somewhere in the range 23–76 PCS. It therefore appears that gabbroic cumulates in general reject a dense solute unless they are oxidized enough to precipitate magnetite.

Solute behavior related to magma type

Liquids that belong to the calc-alkaline trend appear to originate at a density maximum (Bottinga & Weill 1970). The major reason for this is presumably the early crystallization of Fe–Ti oxides. A plutonic example may occur at Lake Owens, Wyoming, where magnetite occurs at a high Mg ratio in the pyroxene (Ridgley 1972, Patchen & Myers 1987). It is probable that the rejected solute (RS) from all average assemblages of calc-alkaline cumulates would be buoyant. The same conclusions have been reached from a somewhat different approach by McBirney (1985). The findings from both studies are generalized and summarized in Figure 8, based on the density of the Kiglapait magma plotted against the Mg ratio of the liquid. Three stages occur for Fenner-trend liquids. Fractionation of olivine causes a decrease im magma density and X_{Mg} until a cusp (Morse 1986b) is reached when plagioclase crystallizes. The rejected solute is then dense until maximum liquid density occurs. After the maximum, at or near the cumulus arrival of magnetite, the RS is intrinsically buoyant and will rise through the overlying magma. For progressively more oxidized magmas, the arrival of magnetite, and with it the density maximum, occur progressively sooner. The density maximum is more or less accurately mapped in Figure 8 for the Skaergaard Intrusion and generalized for the tholeiitic and calc-alkaline series as defined by Jakeš & White (1972). For the extreme calc-alkaline case, no stagnancy occurs, and the density maximum vanishes.

The behavior of magma density and rejected solute is further tied to the Jakeš–White classification in the FMA plot of Figure 9. Here, it is emphasized that retention of olivine at low oxidation states yields the maximum Fenner trend of iron enrichment (Osborn 1959) owing to an "olivine X_{Mg} drive", whereas increased oxidation and silication (Morse 1980) lead to the calc-alkaline trend caused by the drive away from magnetite and other mafic phases in company with the X_{An} drive of plagioclase. Although subject to local circumstances, the inferred relation between the behavior of rejected solute and the oxidation state of magma is probably correct enough for the generalizations of interest here.

Effect of water

The effect of dissolved water will be to lessen the magma density. The addition of water may also oxidize the magma by the preferential escape of hydrogen (Osborn 1959), indirectly causing an earlier density maximum. Bottinga & Weill (1970) found that the Skaergaard liquids continued to increase in density as water was added, but this conclusion was artificially dependent on the implicit assumption that

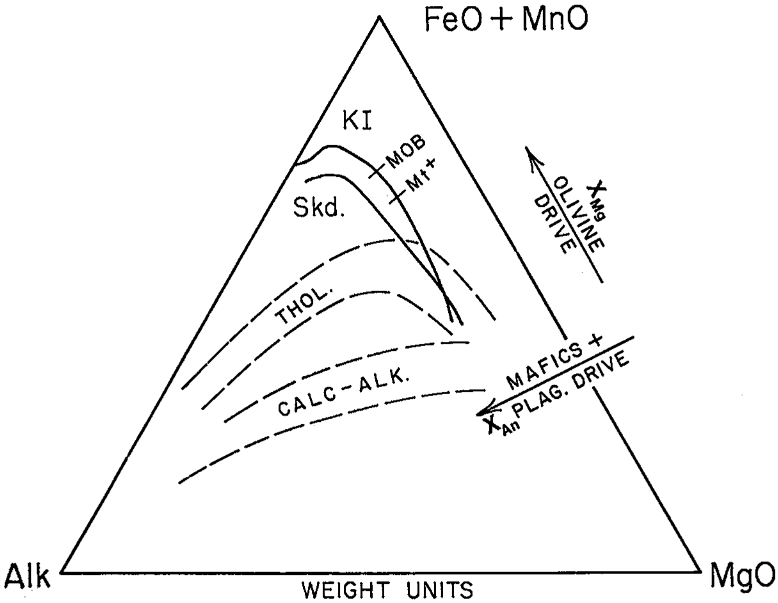


FIG. 9. FMA trends of lava series and Fenner-trend liquids, showing the variation from the most reduced (KI Kiglapait, Skd Skaergaard; Morse 1981) to the most oxidized (calc-alkaline; Jakeš & White 1972). The strongest iron enrichment, hence density increase, results from the fractionation of olivine ("Olivine X_{Mg} Drive" in the Figure). The strongest alkali enrichment, hence density decrease, results from the crystallization of total mafics (especially Fe–Ti oxide minerals) and plagioclase X_{An}, as shown by the "Mafics + Plag Drive" arrow in the Figure. The maximum in the density trend comes latest (at lowest X_{Mg}) for the most reduced magmas. However, an earlier maximum may exist at high Mg ratio for ultramafic magmas. Key: Alk (Na_2O + K_2O), Mt magnetite, MOB main ore band.

no change occurred in the oxidation state or the arrival of cumulus magnetite. The uncertainties make it difficult to generalize about the effect of water in realistic geological settings. It is possible that small amounts of water, as present in the Hyllingen magma (Wilson & Larsen 1985) may have caused an early density maximum, but the extreme iron enrichment and the low abundance of late cumulus amphibole in that intrusion suggest that the effect of water was not by itself large enough to cause a light rejected solute, at least during the early stages of crystallization.

Further Effects of Compositional Convection

Crystal settling

A general case for crystal settling has recently been made by Cox (1985), and the process is in principle augmented by two-phase convection. In addition, the following remarks are pertinent to layered intrusions.

The buoyancy of rejected solute from the growth of olivine should have a moderate effect on the settling of olivine crystals. The crystal occupies a thin film at the lower end of a rising trail of buoyant solute. If the film is driven to the composition of pure plagioclase, it would have a density approaching 2.5 g/cm^3, and the density contrast for olivine may be increased from ~0.8 to nearly 1.0 g/cm^3, enhancing the Stokes settling rate by twenty percent.

The case of plagioclase is opposite to that of olivine in that the rejected solute is dense and will tend to drain downward, tending to elevate the grain. However, the effect is much smaller than for olivine since the composition of the cotectic liquid lies near plagioclase, and supersaturation in the olivine component is likely to be minor. Clearly, however, any suspension of plagioclase crystals should tend to achieve adcumulus growth by drainage of dense rejected solute and the ensuing, intergranular, compositional convection.

The question of crystal settling has been somewhat clouded by the issue of yield strength (McBirney & Noyes 1979), which is irrelevant for small crystal populations. The time-averaged population of crystals in the Kiglapait Intrusion is about 1 ppm by volume (Morse 1979a,b). In the outer sheath where nucleation and growth of crystals occurred, the average crystal population was estimated at less than 3000 ppm, still well within the range of Newtonian behavior. The eventual nucleation and rapid growth of plagioclase near the floor would cause the formation of a protolayer. Within this protolayer, the yield strength may become significant, due to the local increase in crystallinity. A change in the rheological behavior would result, and crystal motions would be impeded (Morse 1979a). The high supercoolings investigated by Murase & McBirney (1973) would be more appropriate to the new protolayer after a burst of nucleation than to the acquisition of new crystals at the interface. Moreover, the experiments of Murase & McBirney were evidently made in air, generating a great excess of network-forming (Mysen *et al.* 1980) ferric oxide in the melt relative to the likely natural conditions. These experiments cannot, therefore, serve as useful guides to processes in most plutonic magmas having few entrained crystals.

Ponding of dense magma

Large-scale convective stirring will tend to inhibit any tendency to ponding, but suppose for the moment that the megascopic, magma-wide flow is weak or nonexistent. Consider a downslope flow initiated by rejected solute that is denser and more evolved than the average magma, as calculated for cotectic and felsic cumulates. In such a case the denser material will tend to pond at the center of the intrusion (Sparks & Huppert 1984). The downward motion is accompanied by an adiabatic compression, but since the melting gradient dT_m/dP is less steep than the adiabat, the magma remains at or below the cosaturation temperature. But the cosaturation composition for feldspar and mafic phases becomes more felsic with higher pressure, so that the dense magma must shed mafic minerals and become lighter. Ponding will not occur if the floor is relatively steep, so that the compression of magma is great enough to cause shedding of mafics.

If the floor dips gently, hence if the compression and consequent extraction of mafic crystals are negligible, and if the denser ejected solute hugs the floor closely, ponding may occur. The ponded liquid is evolved, and so it will react with its floor. The slow cooling of large bodies of magma means that any layer of ponded magma will initially be thin. It may therefore be heated by conduction from the overlying main magma, upon which event it will dissolve the floor, acquire buoyancy, and ascend. This may be a significant process by which the ponding of dense magma is inhibited on shallow-dipping floors.

Two-phase convection may also interfere with ponding, and could perhaps even give rise to potholes filled with ultramafic rock. Consider the liquid that has just become buoyant by dissolving a bit of its floor. Suppose a fresh batch of dense liquid takes its place; the heating and dissolving process is repeated, with the net result that the hole in the floor is deepened. By a long succession of such processes, a substantial hole or well in the central part of the cumulate floor might develop. But a hole in a cumulate is unstable, and sooner or later it will be filled by the one fluid dense enough to flow under the dense rejected solute: a two-phase flow laden with mafic crystals. This might resemble a mudflow caused by the sudden release of a local excess of mafic crystals in a slurry. The resultant hole-filling would be a pipe of dunite or other mafic rock, in which the mafic crystals would be more evolved than the cumulus crystal compositions in the pipe walls. This process is a conceivable alternative to a different sort of melt corrosion suggested by Irvine *et al.* (1983), but in order to be important in a slowly cooled body the catchment area would have to be considerable.

Planetary crusts

If the rejected solute from mafic cumulates can be collected in the flow field and prevented from complete mixing, it may rise to the roof and make plagioclase rafts when it cools. Anorthosite masses inferred to represent such rafts are found in many intrusions, for example Pigeon Point (Grout 1928) and Kiglapait (Morse 1969). By extension of this concept, one may account for the earliest planetary crusts by such a mechanism of concentrating the evolved rejected solute from the growth of ultramafic cumulates (Morse 1987).

Heat pumping

A further consequence of the rising rejected solute from mafic cumulates is that it may aid in melting the roof. Such action appears to have occurred on a local scale in the Rhum intrusion, where finger structures of mafic material protruding upward into allivalite are interpreted as melt features (Butcher *et al.* 1985, Morse 1986c, Morse *et al.* 1987). On a much larger scale, the light RS from mafic cumulates was apparently responsible for melting the roof of the Muskox Intrusion (Irvine 1970). In both cases, latent heat is transferred from growing floor cumulates to melting roof by rapid compositional convection. If such rapid transfer is a prerequisite to roof melting, as seems likely, then the roofs of felsic intrusions should be much less susceptible to melting than those of mafic intrusions. This effect may answer the question raised by Brandeis & Jaupart (1986) as to why some intrusions (here seen as the felsic ones, such as the Kiglapait Intrusion) have upper border zones, whereas others (the more mafic ones, such as

Muskox) lack them, having melted their roofs. Of course, if the roof is refractory it will not melt in any case, hence an upper border zone will be expected to form.

Upward transport of contaminants

The variation in Sr isotopic ratio within many layered intrusions may bear strongly on the motion and structure of intercumulus liquid. In a fine-scale study of Unit 10 at Rhum, Palacz & Tait (1985) showed that radiogenic Sr correlates with total Sr, which in this case resides almost totally in the plagioclase component of the bimodal olivine–plagioclase cumulates. The enrichment of ^{144}Nd common to crustal contamination also follows the felsic component of the cumulate. These relationships are plotted in Figure 10, with the isotopically "enriched" felsic component on the left and the "depleted" component on the right. The Rhum authors ascribe the isotopic variations to magma mixing, but these variations may be ascribed better to rejected felsic solute that retains the contaminant. The idea of a second magma fares badly because the highest degree of crustal contamination is then associated with the cooler felsic magma rather than the hotter mafic magma. It is more likely that a hot mafic magma did most of the assimilation, and that the main contaminant in terms of Sr and Nd isotopes resided in the feldspar component of the magma. The modal relations at Rhum suggest that the plagioclase itself will be found to carry the contaminant signature. A more definitive statement awaits analysis of mineral separates, but the evidence suggests that the contaminant resides in a feldspar-like structural component of the magma without becoming thoroughly equilibrated during the time between assimilation and crystallization (Morse 1983b). In such a case, the isotopic signatures become valuable tracers of the motion and structure of intercumulus liquid.

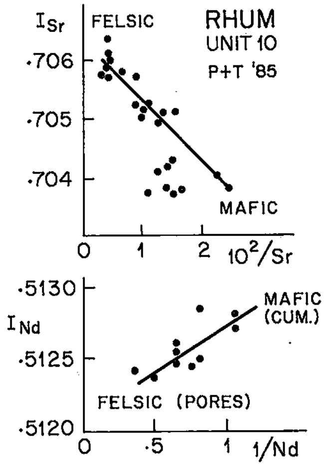


FIG. 10. Initial Sr and Nd isotope ratios of Unit 10 at Rhum, replotted from Palacz & Tait 1985. The abscissae are the reciprocals of concentration, which in this particular case correlate with color index, so that the left-hand end of the scale corresponds to a pure felsic rock (anorthosite). Because the rocks are mafic cumulates grading into felsic cumulates (troctolites), the principal component of trapped liquid is plagioclase. The diagrams can be interpreted as representing mixtures of magmas (as done by the original authors) only if it is supposed that a cooler, felsic magma was more highly contaminated by crustal melts than a hotter, mafic magma. The alternative interpretation, favored here, is that the felsic contaminant travels with the felsic component of the intercumulus liquid, which retains its structural and isotopic integrity over the time scale of assimilation and crystallization.

Lateral migration of solute

Several phenomena imply the lateral migration of intercumulus liquid in partly consolidated crystal mushes. One of these is the megablock in the Skaergaard Intrusion reported by McBirney & Noyes (1979) to display modal layering continuous with that in the surrounding cumulates. On a much smaller scale, lateral migration can be inferred from the study of Rhum cumulates by Young & Donaldson (1985). The likelihood of lateral migration is high if several conditions are satisfied: the porosity and permeability must be sufficiently high in a confined layer; the rejected solute must be of the appropriate density to move along the permeable layer, that is, dense enough to move downslope, light enough to move upslope within the confines of the layer, or neutral to move horizontally; heat loss must be effected either by conduction through the confining strata or by convection to some unidentified heat sink; and crystal growth in the permeable layer must be sufficiently active to impel the migration of solute. It is assumed here that such metasomatic lateral migration would occur chiefly in response to crystal growth in the porous layer, and that the easily nucleated and rapidly grown mafic crystals would be the dominant agents of the process. Such a mechanism would amount to a sort of megascopic overgrowth from heterogeneous centers, somewhat in the nature of oikocrystic growth.

Effect of abundant xenoliths

Any xenolith acts initially as a cold finger, *i.e.*, as a local heat sink, in a mafic magma. One effect of such an object is to supercool the surrounding liquid and cause crystallization, either of the currently stable phases or of lower temperature phases, as shown by the pyroxene rinds on xenoliths in the

Kiglapait Intrusion (Owens 1986). Where xenoliths are abundant, they may cause large-scale cooperative effects in the regional magma. This appears to have happened at Hyllingen (Wilson & Larsen 1985), where the abundance of metabasaltic xenoliths locally approaches 25% or more. The xenoliths are closely associated in space with evolved mineral compositions in the layered rocks, and these become extremely evolved along strike toward the southern margin. An upper border zone is lacking, and the latest syenitic differentiates of the intrusion are in sharp intrusive contact with country-rock amphibolites at the roof. This means that heat extraction occurred to an important degree through the south sidewall and into the xenoliths. The rejected solute was probably light, for the following reasons: the magma was slightly hydrous, the evolved rocks are volatile-bearing syenites (presumably of low color index), and the earlier layered rocks may be more mafic than the cotectic composition owing to the chilling effects of the xenoliths and the nearby sidewall. In any event, it is clear that the rejected solute from the crystallization of a large volume of layered-series rocks became trapped near the south sidewall, causing the extreme lateral variations seen in the mineral chemistry and in the appearance of evolved minerals. It is probable that this migration and trapping would not have occurred without the agency of the xenoliths. If the rejected solute was indeed light, it would represent a rare case of buoyancy despite strong iron enrichment, and an exception to the general case represented here by Figure 8.

Concluding Remarks

Compositional convection is evidently an important process affecting many aspects of the evolution of a magma body into a set of layered cumulate rocks. The behavior of rejected solute may affect the motion and growth of crystals, the flow of magma, the metastable differentiation of magma into parts richer in network-forming, felsic components among parts more mafic, and whether or not a roof may melt or a cumulate floor dissolve. The local isotopic disequilibrium found between felsic and mafic parts of some cumulates offers a chance to monitor the integrity and motion of structural–compositional units in the magma itself.

Phase equilibria, nucleation and growth kinetics, and two-phase convection furnish important boundary conditions in considerations of how magmas and crystals behave. The most important boundary conditions are imposed by the rocks themselves, for they tell us where the crystals are. The inversion from the evidence in the rocks to the processes that produced them, suitably constrained by appropriate boundary conditions, is a necessary and fruitful approach to understanding. In such an exercise, one repeatedly learns that layered intrusions are not volcanoes and that many plutonic systems are unlike vented systems. It is, therefore, not prudent to reason backwards from volcanoes to explain layered intrusions, for then everything begins to look like a volcano. The main premise of the study of layered intrusions has presumably been that they teach us how magmas crystallize and evolve within the crust. There is much left to learn from the exhumed rocks, as well as from theory and experiment cannily used.

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