The Role of Diffusion-controlled Oscillatory Nucleation in the Formation of Line Rock in Pegmatite–Aplite Dikes

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The George Ashley Block (GAB), located in the Pala Pegmatite District, San Diego County, California, is a composite pegmatite–aplite dike of 8 m thickness displaying striking mineralogical layering in the aplite portion of the dike, referred to as line rock. Rhythmic layering is characterized by garnet-rich bands alternating with albite–quartz–muscovite-rich bands. Cumulus textures are notably absent from the layered portion of the dike. Elongated quartz megacrysts are oriented perpendicular to the garnet-rich layers and poikilitically include garnet, albite, and muscovite. Calculated crystal-free magma viscosity with 3% H2O is 10^-6 Pa s and the calculated settling velocity for garnet is 0.51 cm/year. Conductive cooling calculations based on emplacement of a 630°C dike into 150°C fractured gabbroic country rock at 1.5 kbar, and accounting for latent heat of crystallization, demonstrate that the line rock portion of the dike cools to 550°C in about 1 year. Crystal size distribution studies also suggest very rapid nucleation and crystallization. Diffusion-controlled gel crystallization experiments yield textures virtually identical to those observed in the layered aplite, including rhythmic banding, colloform layering, and band discontinuities. Thus, observed textures and calculated magmatic parameters suggest that mineralogical layering in the GAB results from an in situ diffusion-controlled process of oscillatory nucleation and crystallization. We propose that any event that promotes strong undercooling has the potential to initiate rapid heterogeneous nucleation and oscillatory crystal growth, leading to the development of a layer of excluded components in front of the crystallization front, and the formation of line rock.

KEY WORDS: aplite; layering; line rock; oscillatory nucleation; pegmatite

INTRODUCTION

Modal or phase layering is a phenomenon that is seen in plutonic igneous rocks of all compositions from high-temperature, low-viscosity, mafic complexes to low-temperature, high-viscosity, felsic complexes. The sizes of such plutonic bodies can range from the large layered intrusive complexes such as the 3000 m thick Skaergaard complex down to dikes a few meters wide. There is, however, no consensus as to the origin of the layering. Modal layering that results from the gravitational settling of crystals from a magma is a model often invoked to explain the layering in mafic magmatic systems. This is more problematic in granitic systems because of the higher viscosities of the melts and the smaller density difference between crystals and melt.

This paper presents a temperature-crystallization model for the George Ashley composite pegmatite–aplite dike, which displays well-developed rhythmic layering (line rock) in the aplite portion of the dike. Kleck (1996) proposed that the layering in the George Ashley Block (GAB), a float slide block from this dike, could be explained by the gravitational settling of garnet and albite from the melt, and that many of the textural features which characterize the line rock, such as offsets in the layers, formed as a result of soft sediment deformation of a crystal–liquid mush. The purpose of this investigation was to evaluate crystal settling as a possible mechanism for generating the layering in the GAB, as well as other possible layering mechanisms, and to constrain a crystallization model for the GAB.
GEOLOGIC SETTING
The GAB is located in the Pala Pegmatite District in northwestern San Diego County, California (Fig. 1). The Pala district is famous for many gem-bearing pegmatites which have produced tourmaline, spodumene, and beryl (morganite). These pegmatites are situated in the western portion of the Peninsular Ranges province, which roughly parallels the Pacific coast, and are the youngest rocks of the batholith. They are chiefly subhorizontal tabular dikes enclosed in gabbroic host rocks and were emplaced along well-developed sets of fractures in the older gabbro (Foord et al., 1991). The fractures are thought to have resulted from regional cooling and contraction within the batholith as a whole, and fracture development postdated the main period of batholith emplacement (Jahns, 1979). Contacts between pegmatites and host rock are generally sharp and regular in detail. Host rock alteration adjacent to pegmatite bodies is not extensive (Foord et al., 1986). The GAB is located on the eastern slope of Hiriart Mountain (Fig. 1) and was emplaced into 118–120 Ma country rock, the San Marcos Gabbro-norite, which had been extensively fractured before pegmatite emplacement. Pala district pegmatites that have been radiometrically dated yield \(^{40}\)Ar/\(^{39}\)Ar emplacement ages of 99–100 Ma (Foord et al., 1991). The GAB is less evolved than other acclaimed gem-producing dikes with miarolitic cavities in the Pala District. However, rare tourmaline crystals up to 15 cm in length are found in the sparse pockets associated with the core of the GAB, indicating that very late in the crystallization history, boron concentrations were sufficient to form tourmaline. The GAB is characterized by a spectacular footwall quartz-albite aplite that is rhythmically layered with garnet-rich layers (Fig. 2a). Many of the pegmatites in San Diego County have layered aplites and were first described in detail by Jahns & Tuttle (1963). Locally, layered portions of pegmatite–aplite dikes are known as line rock.

FIELD AND MINERALOGICAL DESCRIPTION
The GAB is a tabular dike with an exposed thickness of 8 m and a width of 15 m. A schematic vertical section of the dike is illustrated in Fig. 3 along with the distinctive mineralogical and textural characteristics from bottom to top. The lower 150 cm of the dike is composed of granular pegmatite with crystals up to 5 mm in size of quartz, plagioclase, microcline, and muscovite (Fig. 3a and b). A few scattered aplite layers occur interspersed with the granular pegmatite for the next 100 cm. The distinctly layered line rock begins 245 cm above the base and extends for 70 cm (Fig. 3c). It is characterized by almandine–spessartine (solid solution) enriched layers alternating with layers enriched in albite, quartz, and muscovite with small amounts of microcline, biotite, and tourmaline. A 20 cm thick faintly layered aplite underlies the 85 cm thick core which extends from 335 to 420 cm above the base. The core consists of coarse quartz and perthitic microcline. Tourmaline crystals up to 15 cm in length occur in sparse pockets associated with the core. Granular pegmatite with blocky graphic microcline continues to 600 cm above the base (Fig. 3d). The remaining 210 cm of the pegmatite has a granular texture and contains microcline, but it is no longer graphically intergrown with quartz and the overall grain size decreases to the top of the dike (Fig. 3e).

PETROGRAPHY
Petrographic examination of the line rock samples reveals textures that suggest in situ crystallization and rapid growth. The overall texture is granular with anhedral to subhedral plagioclase, quartz, and muscovite, and euhedral garnet. Elongated quartz grains, which can be as long as 1.5 mm, are commonly oriented perpendicular to the garnet layers (Fig. 2c and d). In some cases, quartz poikilitically encloses plagioclase or garnet (Fig. 2c–f). Muscovite grains which can be as long as 1.5 mm are randomly oriented, showing no parallelism with garnet layers (Fig. 2c–f). Plagioclase is typically <1 mm in length and garnet is <0.5 mm in diameter (Fig. 2b).

WHOLE-ROCK CHEMISTRY
Fifteen samples were collected from the GAB at 60 cm intervals for X-ray fluorescence (XRF) major element whole-rock analysis, and inductively coupled plasma mass spectrometry (ICP-AES) trace-element analysis. Analytical data along with analytical uncertainties can be found in Table 1. The granitic samples are all peraluminous, with an average A/CNK ratio of 1.23. The average composition of the GAB falls close to the 1.5 kbar H₂O saturated minima in the haplogranite system (Fig. 4) as do the other granular pegmatite samples. The samples with an abundance of microcline fall closer to the Or apex, samples more enriched in quartz fall closer to the Q apex, and the line rock sample falls along the Q–Ab join with very little of the Or component. The microcline-rich core was not plotted. Overall, no fractionation trends are seen from the bottom to the top of the GAB. Principal chemical changes are mineralogically controlled and are most apparent in sample F from the line rock (300 cm above the base), and sample G from the core (360 cm above the base). The aplite line rock sample F is mineralogically enriched in garnet, muscovite, and albite, and has virtually no
Fig. 1. Location map of the George Ashley Block (GAB) near the town of Pala, San Diego County, California. The GAB is located on the southern slope of Hiriart Mountain.

microcline. Chemically, this sample has the highest concentrations of Fe₂O₃ (total iron), MnO and Na₂O, and the lowest concentration of K₂O. Sample G from the core has the highest concentration of K₂O, Al₂O₃, B and Li, and the lowest concentration of Na₂O, resulting from the abundance of perthitic microcline, the appearance of scattered, large, tourmaline crystals, and the lack of albite crystals.

MAGMATIC PARAMETERS
Total water and other volatiles content
There is evidence from petrological and fluid-inclusion studies to support the idea that pegmatites do not attain saturation in an aqueous phase until they approach their solidi (London, 1992). There is no evidence that the GAB magma was water saturated when it was emplaced, and saturation was probably only achieved in the final stages of core crystallization with the development of a few small pockets. Overall boron concentrations are low in the GAB and increase from both the top and the bottom of the pegmatite toward the core zone (Table 1). Fluorine abundance in muscovite determined from microprobe analyses (E. E. Foord, analyst, unpublished data, 1993) also increases from both the top and the bottom of the pegmatite toward the core zone. The most F-rich muscovites analyzed were from the line rock sample F below the core (0.4 wt %) and sample H from above the core (0.5 wt %). Muscovite from the remaining samples had no more than 0.1% F. Finally, metasomatic effects on the adjacent country rock, which would have been produced from an escaping vapor phase from the crystallizing pegmatite, are absent. Thus, we have chosen a H₂O content of 3% for calculation purposes, which would permit crystallization of at least 75% of the dike before vapor saturation occurred.

Temperature and pressure
The estimated temperature of emplacement for the GAB was 650°C, based on mineral equilibria of the assemblage quartz, plagioclase, potassium feldspar, muscovite, and garnet (Huang & Wyllie, 1981). As previously mentioned, the average composition of the GAB falls close to the 1.5 kbar H₂O-saturated minima in the haplogranite system (Fig. 4) corresponding to an emplacement depth of ~5 km. According to Brisbin (1986), the rheologic state of the crust at a depth of less than ~7 km will be brittle and intrusion at shallow crustal levels will be accompanied by brittle host-rock behavior. If the host rocks are characterized by fractures with random orientations and the lithostatic stress field is modified by horizontal tectonic
compressive stresses, then the only fractures that will serve as sites for pegmatite emplacement at these depths will be oriented horizontally. However, below a depth of ~7 km, the host rocks would behave in a more ductile fashion and the pegmatites would be emplaced along both vertical and horizontal fractures, and at greater depths would become increasingly more irregular in shape. The majority of the Pala District pegmatites are subhorizontal, tabular dikes that were emplaced along pre-existing fractures in the country rock. Thus, the
Fig. 3. Schematic vertical section of the GAB showing textural differences throughout the dike. Photomicrographs were printed from actual thin sections in polarized light and are therefore inverse images (garnets appear as bright white grains).
# Table 1: Whole-rock analyses of George Ashley block samples

| Sample: | A | B | C | D | E | F | G | H | I | J | K | L | M | N | O | Av. |
|---------|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|-----|
| cm above base: | 0 | 60 | 120 | 180 | 240 | 300 | 360 | 420 | 480 | 540 | 600 | 660 | 720 | 780 | 810 | Av. |
| wt % | | | | | | | | | | | | | | | | | |
| SiO$_2$ | 74.9 (2) | 73.7 | 74.1 | 74.6 | 76.7 | 68.3 | 65.2 | 77.8 | 73.4 | 74.3 | 74.6 | 74.7 | 76.2 | 74.9 | 74.8 | 73.9 |
| Al$_2$O$_3$ | 14.1 (2) | 14.5 | 14.4 | 14.2 | 13.7 | 15.7 | 18.4 | 13.6 | 14.6 | 14.0 | 14.0 | 14.6 | 13.5 | 14.3 | 14.4 | 14.5 |
| Fe$_2$O$_3$ | 0.78 (2) | 0.65 | 0.64 | 0.71 | 0.96 | 4.79 | 0.15 | 0.51 | 0.43 | 0.56 | 0.66 | 0.68 | 0.83 | 0.59 | 0.76 | 0.91 |
| MgO | 0.11 (2) | 0.12 | 0.10 | 0.10 | 0.13 | 0.15 | 0.03 | 0.10 | 0.06 | 0.10 | 0.13 | 0.11 | 0.14 | 0.09 | 0.09 | 0.10 |
| CaO | 0.39 (2) | 0.25 | 0.40 | 0.38 | 0.34 | 0.13 | 0.03 | 0.08 | 0.29 | 0.21 | 0.22 | 0.40 | 0.39 | 0.31 | 0.29 | 0.27 |
| Na$_2$O | 4.37 (4) | 3.53 | 4.26 | 4.56 | 4.33 | 5.56 | 2.90 | 4.11 | 4.28 | 3.21 | 3.33 | 4.32 | 4.16 | 4.38 | 4.19 | 4.10 |
| K$_2$O | 3.31 (2) | 5.56 | 4.02 | 3.31 | 1.92 | 0.75 | 11.77 | 2.04 | 5.19 | 6.23 | 5.33 | 3.41 | 3.05 | 3.66 | 3.66 | 4.21 |
| TiO$_2$ | 0.04 (1) | 0.04 | 0.03 | 0.03 | 0.04 | 0.03 | 0.03 | 0.03 | 0.03 | 0.03 | 0.04 | 0.04 | 0.05 | 0.04 | 0.04 | 0.04 |
| P$_2$O$_5$ | 0.08 (1) | 0.13 | 0.09 | 0.09 | 0.07 | 0.13 | 0.22 | 0.07 | 0.08 | 0.09 | 0.09 | 0.06 | 0.07 | 0.07 | 0.07 | 0.09 |
| MnO | 0.13 (2) | 0.05 | 0.06 | 0.08 | 0.27 | 3.82 | 0.02 | 0.04 | 0.03 | 0.02 | 0.02 | 0.07 | 0.04 | 0.07 | 0.20 | 0.33 |
| LOI | 0.67 (5) | 0.53 | 0.51 | 0.51 | 0.76 | 0.43 | 0.23 | 1.03 | 0.39 | 0.48 | 0.67 | 0.96 | 0.88 | 0.86 | 0.86 | 0.65 |
| Total | 98.88 | 99.04 | 98.64 | 98.47 | 99.30 | 99.74 | 98.97 | 99.40 | 98.83 | 99.23 | 99.10 | 99.34 | 99.25 | 99.24 | 99.32 | 99.12 |
| Tot H$_2$O | 0.60 (5) | 0.56 | 0.56 | 0.52 | 0.74 | 0.39 | 0.26 | 0.96 | 0.41 | 0.46 | 0.74 | 0.75 | 0.74 | 0.67 | 0.69 | 0.60 |
| B (ppm) | 10 (1) | 12 | 10 | 10 | 12 | 12 | 41 | 65 | 29 | 12 | 10 | 11 | 13 | 11 | 12 | 18 |
| Li (ppm) | 54 (1) | 88 | 77 | 98 | 100 | 68 | 210 | 160 | 99 | 87 | 78 | 52 | 61 | 59 | 50 | 89 |

**CIPW norm**

| Q | 36.7 | 32.0 | 33.8 | 35.4 | 44.1 | 31.8 | 3.8 | 46.7 | 28.6 | 31.8 | 35.0 | 36.2 | 40.1 | 35.4 | 36.4 | 34.0 |
| Or | 19.9 | 33.4 | 24.2 | 20.0 | 11.5 | 4.5 | 70.4 | 12.3 | 31.2 | 37.3 | 32.0 | 20.5 | 18.3 | 22.0 | 22.0 | 25.3 |
| Ab | 37.7 | 30.3 | 36.7 | 39.4 | 37.2 | 47.4 | 24.9 | 35.4 | 36.8 | 27.5 | 28.6 | 37.2 | 35.8 | 37.7 | 36.0 | 35.2 |
| An | 1.4 | 0.4 | 1.4 | 1.3 | 1.2 | 0.0 | 0.0 | 0.0 | 0.9 | 0.5 | 0.5 | 1.6 | 1.5 | 1.1 | 1.0 | 0.8 |
| C | 2.8 | 2.6 | 2.6 | 2.4 | 4.1 | 5.8 | 1.4 | 4.7 | 1.7 | 1.9 | 2.6 | 3.3 | 2.8 | 2.8 | 2.8 | 3.0 |
| Hy | 0.0 | 0.0 | 0.0 | 0.0 | 3.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 |
| Mt | 0.3 | 0.0 | 0.1 | 0.2 | 0.8 | 7.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.1 | 0.0 | 0.1 | 0.5 | 1.0 |
| Ilm | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.0 | 0.0 | 0.0 | 0.0 | 0.1 | 0.1 | 0.1 | 0.1 |
| He | 0.6 | 0.6 | 0.6 | 0.4 | 0.0 | 0.2 | 0.5 | 0.4 | 0.6 | 0.7 | 0.6 | 0.8 | 0.5 | 0.4 | 0.2 |
| Ap | 0.2 | 0.3 | 0.2 | 0.2 | 0.2 | 0.3 | 0.5 | 0.2 | 0.2 | 0.2 | 0.2 | 0.2 | 0.2 | 0.2 | 0.2 | 0.2 |

Major elements in wt % were determined by X-ray fluorescence (PW-1606 instrument), using fused Li-tetraborate disks, by the method of Taggart et al. (1987), analysts D. F. Siems and J. S. Mee. Total iron is expressed as Fe$_2$O$_3$. Errors for each oxide are shown in parentheses for sample A. LOI, loss on ignition at 925°C; Tot H$_2$O, total water, determined by Karl Fisher Titration, using the method of Jackson et al. (1987), analyst T. R. Peacock. B and Li in ppm were determined by ICP-AES, using the method of Lichte et al. (1987), analyst D. L. Fey. Errors are shown in parentheses for sample A. All analysts and equipment from US Geological Survey, Lakewood, CO.
log \eta = (0.0292649 T - 53.198903 X_{\text{tot H}_2\text{O}} - 0.0129207 T + 24.2973)

for \( X_{\text{tot H}_2\text{O}} > 0.25 \) the equation is

log \eta = -2.2977826 X_{\text{tot H}_2\text{O}} - 0.0095110 T + 15.6293

where viscosity is in pascal-seconds, and temperature, \( T \), is in kelvins. This model is applicable to granitic melts with total \( \text{H}_2\text{O} \) contents up to 12.3 wt % and yields melt viscosities much lower than those estimated using the method of Shaw (1972).

Calculated viscosities for the GAB are \( 10^{6.2} \) Pa s for 3.0 wt % \( \text{H}_2\text{O} \), \( 10^{6.0} \) Pa s for 4.0 wt % \( \text{H}_2\text{O} \), \( 10^{5.8} \) Pa s for 6.0 wt % \( \text{H}_2\text{O} \), and \( 10^{5.5} \) Pa s for 9.0 wt % \( \text{H}_2\text{O} \).

These calculated values demonstrate the small decrease in viscosity above 4% \( \text{H}_2\text{O} \), which can be attributed to the dominant hydrous species present in the melt. At 3 wt % or less total water the dominant melt species is \( \text{H}_2\text{O} \), which interacts weakly with Si-O-Si bonds and has little effect on melt viscosity (Baker, 1996).

**CRYSTALLIZATION AND COOLING MODEL**

**Importance of conduction vs convection**

To constrain a realistic cooling model, many parameters need to be evaluated, including those related to the magma and its intrusion (temperature and depth of emplacement, thermal diffusivity, latent heat of crystallization, temperature interval of crystallization, and intrusion size), as well as knowledge about the initial temperature, thermal diffusivity, and geothermal gradient of the host rock (Jaeger, 1968). Additionally, one must evaluate whether conductive or convective processes were acting upon the magma body during crystallization.

Convection and gravitational settling of crystals have long been accepted as two important processes governing the evolution and differentiation of magmas (Bowen, 1928; Wager & Brown, 1967; Sparks et al., 1993). However, the importance of convection and crystal settling in thin sills and dikes has been re-evaluated by a number of workers, including Brandeis & Jaupart (1987), Marsh (1988a, 1989), Spohn et al. (1988), Gibb & Henderson (1992), and Mangan & Marsh (1992). Although many of these models have predominantly evaluated mafic composition sheet-like magma bodies, they all conclude that either cooling in thin sills or dikes is entirely by conduction or that if convection is occurring, it will not be vigorous.
 Frequently, the velocity and vigor of convection has been evaluated from the magnitude of the dimensionless Rayleigh number \( Ra \), the ratio of thermally induced buoyancy to viscous drag:

\[
Ra = \frac{\alpha \Delta T L^3}{\nu \kappa}
\]  

where \( g \) is acceleration due to gravity, \( \alpha \) is the coefficient of thermal expansion, \( \Delta T \) is the temperature gradient, \( L \) is the thickness of the liquid layer, \( \nu \) is kinematic viscosity \((\eta/\rho)\) and \( \kappa \) is thermal diffusivity. A fluid cooled at the top or heated at the base should convect spontaneously when \( Ra \) exceeds a value of \( \sim 2000 \) (McBirney, 1993). The difficulty in determining an appropriate \( Ra \) value for any magma body lies in choosing meaningful \( \Delta T \) and \( L \) values. Jaupart & Tait (1995) suggested that the appropriate scales may be the thickness of the unstable layer that controls convection is a thin layer near the leading edge of the descending crystallization front and that convection will not be vigorous in thin dikes. He suggested that the vigor and style of convection within a sheet-like magma body should not be measured by the usual measure of \( Ra \) [equation (4)], which assumes heating from below, as it suggests rates of heat transfer so large that solidification times are unreasonably short. Instead, although there may be motion within sheets of magma, it is so slight that the thermal history is well approximated by heat conduction alone with inclusion of latent heat. Additionally, Brandeis & Jaupart (1987) suggested that conduction is the dominant mechanism of heat transport in dikes up to widths of 100 m. Therefore, we conclude that a conductive cooling model is appropriate for the 8 m wide GAB dike.

**Cooling parameters for the GAB**

As previously discussed, the GAB magma was intruded into the extensively fractured 118–120 Ma San Marcos Gabbro. As a result of pegmatite emplacement in the Pala District, the age of the country rock, as determined from both biotite and hornblende \(^{40}\)Ar/\(^{39}\)Ar geochronology, was reset to the time of pegmatite emplacement \( \sim 100 \) Ma (Foord et al., 1991). However, samples collected from the spatially associated Woodson Mountain Granodiorite at distances up to 15 km away from the pegmatites have not been affected by pegmatite intrusion. According to Foord et al. (1991), this granodiorite was emplaced before 113 Ma (hornblende age), and it cooled below 300°C (biotite argon-retention temperature) by 107 Ma (biotite apparent age), and had cooled to below 150°C (microcline argon-retention temperature) by 104 Ma (microcline apparent age) before pegmatite emplacement. This temperature of 150°C is consistent with what obtained using a geothermal gradient of 30°C/km for the GAB dike emplaced at a depth of 5 km. Based on phase equilibria (Huang & Wyllie, 1981), the GAB magmatic temperature of emplacement was \( \sim 650°C \). The temperature interval of magmatic crystallization thus began at 650°C and, for a melt of this composition, should have been essentially completed by the time the magma cooled to \( \sim 550°C \). The simplifying assumption that thermal diffusivity is the same for both magma and country rock was used for the cooling models. Thermal diffusivity can be defined as

\[
\kappa = \frac{K}{\rho C_p}
\]

where \( \kappa \) is thermal diffusivity in cm\(^2\) s\(^{-1}\), \( K \) is the coefficient of thermal conductivity measured in cal cm\(^{-1}\) s\(^{-1}\) C\(^{-1}\), \( \rho \) is density in g cm\(^{-3}\), and \( C_p \) is heat capacity in cal g\(^{-1}\) C\(^{-1}\). Heat capacity is defined by

\[
C_v = \frac{\Delta q}{\Delta T}
\]

where \( \Delta q \) is the amount of heat that must be added or removed to raise or lower the temperature of one gram by one degree centigrade. Heat of crystallization, \( \Delta q_c \), is the heat liberated by the crystallization of one gram of melt that is already at the temperature where the liquid and solid coexist. For most rocks, \( \Delta q_c \) is \( \sim 0.3 \) cal g\(^{-1}\) C\(^{-1}\), whereas typical values for \( \Delta q_c \) are \( \sim 60–100 \) cal g\(^{-1}\) (McBirney, 1993). This amounts to roughly the same amount of heat in the crystallization of a melt as in raising the temperature of the melt by 200–300°C, which will reduce the rate of cooling of the magma as it passes through the crystallization interval. The heat of crystallization \( \Delta q_c \) given up during crystallization through a temperature interval \( T_1-T_2 \), can be allowed for by adding the proportional heat per degree to the heat capacity to obtain an effective heat capacity, \( h \) (Jaeger, 1968):

\[
h = C_p + \frac{\Delta q_c}{T_1 - T_2}
\]

This can be substituted for \( C_p \) in equation (6) to obtain an appropriate value which accounts for the heat of crystallization.

Using a \( C_p \) of 0.3 cal g\(^{-1}\) C\(^{-1}\), a \( \Delta q_c \) of 75 cal g\(^{-1}\) (McBirney, 1993), and a temperature interval \( T_1-T_2 \) of 100°C (from 650°C to 550°C) the effective heat capacity \( h \) is 1.05 cal g\(^{-1}\) C\(^{-1}\). Then, using a thermal conductivity \( K \) of \( 4 \times 10^{-3} \) cal cm\(^{-1}\) s\(^{-1}\) C\(^{-1}\), the GAB calculated magma density of 2.31 g cm\(^{-3}\), results in a thermal diffusivity \( \kappa \) value of \( 1.66 \times 10^{-7} \) m\(^2\) s\(^{-1}\). The thermal diffusivity obtained by not taking latent heat of crystallization into account, and using the same magma
density, heat capacity and thermal conductivity, is 5.80 × 10⁻³ m² s⁻¹.

The conduction equation for a sheet of thickness 2a (Jaeger, 1968) that assumes the same thermal diffusivity, \( \kappa \), for both magma and country rock is as follows:

\[
\frac{T}{T_0} = \frac{1}{2} \left[ \text{erf} \left( \frac{x/a + 1}{2 \sqrt{k t/a^2}} \right) - \text{erf} \left( \frac{x/a - 1}{2 \sqrt{k t/a^2}} \right) \right]
\]

(8)

where

\[
\text{erf}(y) = \frac{2}{\sqrt{\pi}} \int_{0}^{y} \exp(-u^2) \, du
\]

(9)

\( T \) is the temperature at time \( t \) (in seconds) at distance \( x \) (in meters) from the mid-plane of the sheet minus the country rock temperature, \( T = T(x) - T_0 \), and \( T_0 = T_m - T_cr \), where \( T_m \) is equal to the initial magma temperature and \( T_cr \) is equal to the country rock temperature.

The \text{erf} function can be approximated (Philpotts, 1990) by

\[
\text{erf}(y) \approx 1 - \left( a_1 \Gamma + a_2 \Gamma^2 \right) \exp\left[ -\left( \frac{y^2}{2} \right) \right]
\]

(10)

where \( \Gamma = 1/(1 + 0.47047) \); \( a_1 = 0.34802; \) \( a_2 = -0.09587; \) \( a_3 = 0.74785 \).

The cooling history of the GAB was modeled in two ways. The first model assumed no latent heat of crystallization was generated during the crystallization of the dike, and used a thermal diffusivity of 5.80 × 10⁻³ m² s⁻¹. The line rock area of the dike, which is 240–300 cm above the base, cools to below 550°C in ~75–125 days after emplacement (Fig. 5a). The second model, which accounts for latent heat of crystallization by using a modified thermal diffusivity value of 1.66 × 10⁻³ m² s⁻¹, slows the rate of cooling, such that the top of the line rock cools to below 550°C in ~300 days (Fig. 5b).

**CRYSTAL SIZE DISTRIBUTION OF GAB GARNET LAYERS**

**Theory**

The theory of crystal size distribution (CSD) was initially developed by chemical engineers (Randolph & Larson, 1971) for evaluating the kinetics of crystallization, i.e. growth and nucleation rates, of industrial compounds. CSD is an empirical model of crystal nucleation and growth based on a population balance of crystals in the system. Crystals move into and out of a size range through physical movement in the system and growth into and out of the size range of interest. Marsh (1988a) developed the standard CSD theory for geologic systems, to determine the kinetics of crystallization by examining variations in grain size. On the basis of grain-size information it is then possible to evaluate nucleation rate, growth rate, growth or residence time and the degree of undercooling. Cashman & Marsh (1988) applied CSD theory to rocks from the Makaopuhi lava lake, and evaluated growth and nucleation rates for crystals as well as the degree of undercooling. The crystal population density \( n \) (number of crystals in a given size class per unit volume) is defined as

\[
n = \frac{\Delta N}{\Delta L} = \frac{dN}{dL}
\]

(11)

where \( L \) is the characteristic linear size, and \( N \) is equal to the cumulative number of crystals per unit volume less than or equal to \( L \). The population density \( n \) can be related to crystal size \( L \) at steady state by

\[
n = n^0 \exp \left( -\frac{L}{G} \right)
\]

(12)

where \( n^0 \) is nucleation population density at \( L=0 \), \( G \) is the average crystal growth rate, and \( \tau \) is the average crystal growth time (residence time). This yields a straight line with a slope of \(-1/G\) and a \( y \)-intercept of \( n^0 \). Thus average growth rate \( G \) and the average nucleation rate \( J \) of the system are related through equation (13):

\[
J = n^0 G.
\]

(13)

**Calculations**

In CSD studies, the crystal dimension measured is the longest horizontal chord. The number of crystals per unit volume \( N_V \) can be determined from the number of crystals per unit area \( N_A \) using the conversion

\[
N_V = (N_A)^{1/3}
\]

(14)

as suggested by Wager (1961) and Kirkpatrick (1977). Garnets in the GAB are euhedral and nearly spherical, and are thus ideal for CSD studies. Values of \( n \) were calculated numerically from measured values of \( N \) and \( L \) from BSE images of about 2500 garnets from 13 garnet layers, using image analysis. A cumulative CSD plot of \( ln n \) vs \( L \) for all 13 layers generates a linear trend with a slope of \(-1/G\) and an intercept of \( n^0 \), the nucleation density. A linear regression through the GAB garnet data points was used to calculate the slope and intercept and yields a \( G \) value of 5.52 × 10⁻³ cm and an \( n^0 \) value of 1.9 × 10⁻⁴ cm⁻¹.

Using this \( n^0 \) value and the relationship in equation (13), it is then possible to calculate either \( J \) or \( G \), if one of the two variables can be determined.

**Nucleation and crystal growth rates**

Brandeis & Jaupart (1987) used dimensional analysis of crystallization equations for conductive cooling together
with crystal size data from thin dikes to obtain values for nucleation rates $J$ and growth rates $G$. The rates for both variables can be expressed as a function of temperature and undercooling, following kinetic theory. Both the nucleation and growth rate functions have bell-shaped curves with respect to degree of undercooling (Brandeis et al., 1984). In geologic systems, the total range of possible nucleation rates is much larger than that of growth rates. Brandeis & Jaupart (1987) suggested that local growth rates in silicate systems in general range from $10^{-10}$ to $10^{-8}$ cm s$^{-1}$, and local nucleation rates are on the order of $10^{-7} - 10^{-5}$ cm$^3$s$^{-1}$. Their data from dike margins constrain maximum growth rates of close to $10^{-2}$ cm s$^{-1}$ and nucleation rates of $\sim 1$ cm$^3$s$^{-1}$. The characteristic cooling time, $\tau$, can be related to $G$ and $J$ by

$$\tau = (GJ)^{-\frac{1}{4}}$$

(15)

For calculation purposes we have chosen growth rates of $10^{-8}$ cm s$^{-1}$ and $10^{-7}$ cm s$^{-1}$, as these values yield characteristic cooling times ($\tau$) consistent with those of Brandeis & Jaupart (1987) of between $10^5$ and $10^7$ s for conductive cooling of dikes less than 10 m in width. A characteristic cooling time of $> 10^5$ s would result in equilibrium crystallization and grain size corresponding to a thicker dike. If a growth rate of $10^{-8}$ cm s$^{-1}$ is used along with the $n^0$ value obtained from the GAB garnet CSD study, then the nucleation rate $J$ is equal to $1.9 \times 10^{-5}$ cm$^3$s$^{-1}$ [from equation (13)], and the characteristic cooling time ($\tau$) is $2.69 \times 10^6$ s, or about 31 days. A growth rate of $10^{-7}$ cm s$^{-1}$ yields a characteristic cooling time ($\tau$) of $4.78 \times 10^5$ s, or about 5.5 days. These results underscore the rapid crystallization time for the line rock portion of the pegmatite–aplite dike compared with the rest of the dike based on our conductive cooling models.

**LAYERING MECHANISMS**

Many different mechanisms have been proposed for layering in igneous rocks including gravitational settling with or without convection (Wager & Brown, 1967;
Huppert & Sparks, 1980), oscillatory nucleation and crystal settling (Hort et al., 1993); and in situ crystallization from a static boundary layer (Jackson, 1961; Mc Birney & Noyes, 1979).

Modal layering that results from the gravitational settling of crystals from magma is an often invoked model to explain layering in mafic magmatic systems. A settling mechanism is more problematic in granitic systems because of the higher viscosities of the magmas and the smaller density difference between crystals and magma. Settling velocities ($V_s$) were calculated for garnet and albite for the line rock portion of the GAB using Stokes’ Law:

$$V_s = \frac{2g\rho_r^2\Delta \rho}{9\eta}$$

where g is acceleration due to gravity, $r$ is the radius of the crystal, $\Delta \rho$ is the density difference between the crystal and the magma, and $\eta$ is viscosity. Stokes’ Law is only applicable for spherical grains settling in a Newtonian, non-convectioning magma. Therefore, settling velocities were only calculated for garnet and albite as both muscovite and quartz are typically non-spherical, elongated grains. A viscosity of $10^6$ Pa s, a magma density of 2.31 g/cm$^3$, and mineral densities of 4.2 g/cm$^3$ for garnet and 2.62 g/cm$^3$ for albite were used for calculation purposes. Maximum crystal size diameters (Fig. 2b) of 0.5 mm for garnet and 1.0 mm for albite yield settling velocities ($V_s$) of 0.51 cm/yr and 0.34 cm/yr, respectively.

Jaupart & Tait (1995) emphasized the importance of the thermal regime of the magma on crystal settling. In the absence of thermal convection, crystals nucleate and grow only in boundary layers. In this case, the temperature of the magma away from the boundary layer is higher than that at the boundary layer. However, if thermal convection is operating in the magma, the whole body will be cooled and crystal nucleation and

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**Fig. 5.** (a) Cooling curves for the GAB calculated with no latent heat of crystallization. Position of line rock from center of dike (in meters) is shown. The contact between the country rock and the GAB is at 4 m. (b) Cooling curves for the GAB calculated with latent heat of crystallization. Position of line rock from center of dike (in meters) is shown. The contact between the country rock and the GAB is at 4 m.
growth can occur anywhere in the magma body. The implication of this is that the cumulus theory of gravitational settling relies on the existence of thermal convection. According to Marsh (1988b), crystallization in basaltic composition sheet-like magma chambers is inward progressing and takes place along a mush—
suspension interface. Near this interface the viscosity increases and thus acts as a capture front that overtakes and traps slowly settling crystals. Crystals nucleated and grown in the suspension zone can escape and settle only if the capture front slows to a critical rate attainable only in bodies thicker than 100 m. Marsh further suggested that sills of viscous silicic magma would have to be more than ~1 km thick before capture by the downward progressing solidification front could be averted. Gibb & Henderson (1992) evaluated convection and crystal settling in mafic sills, and concluded that the magmas had not undergone turbulent convection and that gravitational settling did not play a significant role in modifying the sills.

Hort et al. (1993) considered the role of oscillatory nucleation in the development of layering and cumulates, coupled with convection. Oscillatory nucleation is initiated by slight compositional variations of the liquid resulting from the crystallization of one phase over another, i.e. crystallization of one phase changes the melt composition to favor crystallization of a second phase. In their model, continued growth eventually allows settling, thus leading to phase layering on the floor of the chamber. However, Hort et al. suggested that relatively small bodies cool too fast to nucleate and grow crystals large enough to effectively settle and produce cumulates, and they concluded that layering is unlikely in sheet-like magmas thinner than ~100 m.

Model for the GAB

Based on the preceding arguments against settling and convection, we have concluded that no large-scale convection occurred in the GAB magma. Given the short time necessary for conductive cooling of the GAB coupled with the slow Stokes' Law settling velocities for garnet and albite, it is unlikely that crystals nucleating and growing at the boundary layer could escape a downward solidifying capture front to accumulate in layers in the line rock. Therefore, we conclude that gravitational settling cannot be the mechanism for generating the line rock layering. However, as rhythmic layering does exist in the line rock, we need a mechanism for rhythmic crystallization which does not result from gravitational settling. We propose that the line rock layering resulted from a process of diffusion-controlled oscillatory nucleation and crystallization that took place in situ at the boundary layer.

Evidence for in situ crystallization

The lack of preferred orientation of elongated quartz and muscovite grains parallel to the line rock garnet layers precludes the formation of the layering as the result of some type of flow process. Instead, textural evidence such as elongated quartz and muscovite grains oriented perpendicular to the garnet layers (Fig. 2), supports in situ crystallization. Elongated quartz grains poikilitically include plagioclase, garnet and muscovite. In some instances, plagioclase is smaller in the center of the quartz grain and is larger toward the edges. McBurney & Noyes (1979) interpreted plagioclase laths enclosed in poikilitic pyroxenes as evidence that the phases nucleated and grew together in situ. In the case of the GAB it appears that fewer quartz nuclei formed than did the other minerals, hence as the minerals grew together the quartz enclosed them, and thus stopped further growth.

Geochemically, the GAB shows little variation in whole-rock chemistry from bottom to top (with the exception of the garnet-rich layers). Because in situ crystallization does not change the overall magma chemistry, the geochemical implications are that fractional crystallization will not take place. In layered complexes where fractionation occurs by the removal of crystals, mineral chemical trends are smooth, i.e. olivines vary from Mg to Fe rich, plagioclase from a higher to lower An content. However, in the GAB, garnets do not change chemically in a systematic way from Fe-rich to Mn-rich compositions toward the center of the dike (E. E. Foord, analyst, unpublished data, 1993). This suggests to us that the chemistry of garnets in the GAB is controlled by the melt chemistry of the boundary layer.

Oscillatory nucleation and crystallization

Rapid nucleation and growth of albite and quartz in the line rock aided in the development of a boundary layer of excluded components (initially Mn and Fe) ahead of the crystallization front, which accumulated to the point of local saturation. This led to the nucleation and growth of garnet, which continued as long as the concentrations of Fe and Mn were high. When Fe and Mn concentrations decreased at the crystallization front, nucleation stopped and growth slowed until the concentrations built back up again to saturation levels. As garnet nucleation slowed, the system became dominated by the nucleation and crystallization of albite, quartz and muscovite. This rhythmic oscillation of garnet growth repeated itself throughout the line rock interval.

GEL EXPERIMENTS

To evaluate whether the textures produced in diffusion-controlled crystallization would be similar to the observed
line rock textures, we designed a series of gel crystallization experiments. We were particularly interested in trying to reproduce the colloform structures and layer discontinuities that are observable in the line rock (Fig. 2a). Rhythmic crystallization was originally recognized by Liesegang [cited by McBirney & Noyes (1979)], who suggested that banding in agates resulted from a diffusion-controlled process. He demonstrated this process of oscillatory crystallization by placing a small crystal of silver nitrate in a dilute solution of potassium dichromate in gelatin. Successive bands of silver chromate would nucleate and grow when supersaturation occurred. Between the bands were areas of no crystallization as a result of local depletion of the surrounding area in chromate ions.

For our experiments we dissolved potassium dichromate in gelatin to which we added a solution of silver nitrate and lead nitrate in 50/50 molar quantities. To try to simulate bands of crystals, instead of rings, we added the solution as a streak on top of the gel and let it diffuse. In some cases, the amount of solution was varied along the length of the streak. The gels were then dried in a desiccator, sliced into ribbons, carbon coated and then investigated by BSE imaging (Fig. 2g and h). The colloform-type structures seen in Fig. 2g were produced when concentration was varied along the streak. Breaks in bands and variation in band width are apparent, as well as the presence of a few scattered crystals between the bands.

The similarity between gel crystallization textures and those seen in line rock reinforce our belief that diffusion-controlled crystallization is an important process in the formation of the GAB. The most distinctive line rock texture in the GAB which was duplicated in the diffusion-controlled gel experiments was the colloform structures (Fig. 2a and g).

DISCUSSION

The textural relationships in the GAB dike are a reflection of the overall nucleation and growth history of the minerals crystallizing from the dike. In most mafic dikes crystal size increases only slightly from finer-grained margins to coarser-grained interiors, suggesting that nucleation and growth rates vary only slightly and regularly with time (Cashman & Marsh, 1988; Marsh, 1988a). It is reasonable to expect that similar kinetics of nucleation and growth should apply to granitic dikes. Such is not the case with the GAB. The abrupt change in nucleation rates coincident with the onset of line rock formation (Fig. 3) is strong evidence that the normal crystallization process was interrupted.

Upon emplacement, nucleation and crystallization in the GAB took place along boundary layers at both the top and bottom of the magma sheet. This is reflected by the granular texture observed along both margins (Fig. 3). As crystallization of the mainly anhydrous phases proceeded, volatile phases were concentrated as excluded components in the remaining melt in the central portion of the uncrystallized dike. Had crystallization proceeded along these lines uninterrupted, a typical coarsening inward pegmatite would have resulted, with an overall granular texture up to the point of core formation, where, if the volatile content was sufficient and overall pressure was low enough, miarolitic cavities could have formed. However, in the case of the GAB, after ~75% crystallization, when the magma was close to vapor saturation, the normal progression of crystallization was interrupted and rapid nucleation and crystallization ensued, producing the line rock portion of the dike. We believe that the trigger for the destabilization of the system was a sudden pressure loss within the dike caused either by dike rupture or the dilatancy of the dike because of fracture propagation of the stressed country rock. This sudden decrease in confining pressure would increase the degree of undercooling and change the crystallization kinetics, initiating rapid heterogeneous nucleation and disequilibrium crystallization, as indicated by the CSD results. The ensuing rapid nucleation and growth of albitie, quartz and muscovite would result in the development of a boundary layer of excluded components (initially Mn and Fe) ahead of the crystallization front which would accumulate to the point of local saturation and thus set up the right conditions for oscillatory crystallization to begin. In this peraluminous melt, the saturation of Mn and Fe at the crystallization front led to the nucleation and growth of garnet which continued until the concentration of Mn and Fe decreased. The nucleation and growth of garnet would then slow dramatically until the concentration of the excluded elements built up again to saturation levels. During line rock formation, boron concentrations were not sufficiently high to permit the crystallization of significant amounts of tourmaline. Albite and quartz nucleation and growth continued throughout. With the melt already close to vapor saturation it appears likely that the drop in pressure and subsequent rapid nucleation and crystallization of the line rock would trigger the exsolution of a vapor phase. The volatiles would continue to be concentrated in the residual melt along the upper crystallization boundary layer where they would inhibit nucleation and by their flushing effect possibly cause resorption of crystals (McBirney, 1987). Line rock formation ceased when sufficient volatile phases had accumulated and the degree of undercooling decreased. Normal equilibrium granular crystallization then resumed. Thus, the upward concentration of volatiles may be the factor that restricts line rock to the footwall portion of many sheet-like pegmatite-aplite dikes.

In conclusion, it appears that the GAB cooled rapidly by heat conduction. The rapid crystallization of the dike,
slow calculated settling velocities for garnet and albite, high magma viscosity, and thin dike size preclude the formation of layering in the GAB by crystal settling. Additionally, because of the thin dike size, crystal settling velocities are not high enough for crystals to escape the downward solidification front. Finally, we suggest that any event that promotes undercooling has the potential to initiate line rock formation. Loss of volatiles to the country rock, crystallization of tormulane which effectively removes boron from the magma (Rockhold et al., 1987), dike rupture or dike dilatancy can increase the degree of undercooling of the melt and act as a trigger to destabilize the crystallization dynamics of the pegmatite system. Such events can potentially initiate rapid heterogeneous nucleation and oscillatory crystal growth, the development of a layer of excluded components in front of the crystallization front, and the formation of line rock.

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