Crystallization history of a zoned plagioclase

(Microprobe analysis of zoned plagioclase from the Grand Manan tholeiite sheet)

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SUMMARY. Electron microprobe and textural data are used to deduce a crystallization history for plagioclase in the Grand Manan Island (Canada) diabase sheet. The plagioclase crystals consist of a prominent core (An_{79-67}) surrounded by a distinctly more calcic zone (An_{82-76}), which grades into a normally zoned mantle (An_{73-40}). This zoning pattern is interpreted as reflecting the elevation of a dry diabase magma from at least 10 to 20 km to near surface conditions.

THE crystallization history of an igneous rock unit may be inferred from the compositional zoning of the plagioclase feldspars. Vance (1962, 1965) reviewed the theories affecting compositional zoning and Bottinga et al. (1966) further evaluated the role of diffusion in crystallization from a multicomponent melt. Correlating compositional zones from one crystal to another is difficult because thin sections include crystal sections that are random both in orientation and in relative position within the crystals. Large scale non-cyclical features such as normally-zoned mantles are more easily correlated and reflect conditions that have affected the body as a whole. Small-scale oscillatory zone patterns cannot be correlated with any degree of confidence possibly because these features reflect local conditions. This paper provides a description and interpretation of zoning patterns in plagioclase from a tholeiite sheet.

General petrography. Gunter (1967) described the general petrography and chemistry of a 500 ft thick basalt sheet of Triassic age exposed on the west coast of Grand Manan Island, Bay of Fundy, New Brunswick, Canada. The sample chosen for the present study was taken from the 357·9 ft level (134 ft above the base) and consists of seriate plagioclase and pyroxene up to 2 mm in length with mesostasis consisting of smaller feldspar grains and micropegmatite. The modal composition is 40 % plagioclase, 40 % clinopyroxene, 5 % orthopyroxene, and 15 % mesostasis and the normative feldspar composition is $Or_{8.8}Ab_{34.6}An_{56.6}$.

The larger plagioclase crystals (hereafter referred to as phenocrysts) in the central 450 ft of the sheet contain an optically distinct core surrounded by a distinctly more calcic zone, which is the inner portion of a normally zoned mantle. Although the proportion of core to mantle increases with depth, the general compositional pattern is retained.

Three lines of evidence indicate that the plagioclase and augite co-crystallized throughout the event that resulted in the major oscillation of the plagioclase crystals: glomerocrysts of plagioclase, augite, and orthopyroxene occur near the

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contacts of the sill, a few pyroxene and opaque grains are included along the core-mantle boundary in plagioclase phenocrysts, and an optically distinct zone is observed in many of the larger augite grains. More detailed petrography and textural features of the sheet will be published elsewhere.

Technique. In each phenocryst optically distinct features such as twin planes, composition planes, and zoning were located and marked on photomicrographs. Areas large enough for spot electron microprobe analysis were selected and compositions determined. Details of analytical procedures and a comparison of microprobe
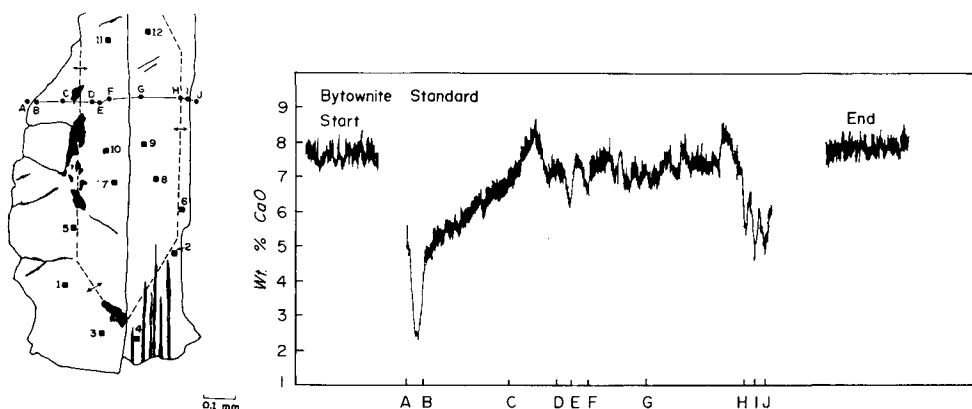

Fig. 1*a*, left: Sketch of a plagioclase phenocryst with the microprobe traverse line and sample points where compositions were determined. The dashed line marks the anorthitic zone between the oscillatory zoned core and the normally zoned mantle. The mantle is poorly developed on the side (I–J) due to an adjacent phenocryst. *b*, right: Continuous scan microprobe traverse line of wt. % CaO. Full capitals define reference points on the traverse in fig. 1*a*.

and optical determinations are published elsewhere (Pringle *et al.*, 1973). Total analyses were completed for 214 spot locations (5 μm in diameter) in thirteen pheno-crysts and some mesostasis areas and additional scanning traverses made on four phenocrysts. An example of the control for the locations is present in fig. 1. Micro-probe spot locations were divided into four classes: core, inner mantle,[1] outer mantle, and mesostasis, on the basis of observed position within the crystal and not on the basis of determined compositions.

Plagioclase compositions. Values of weight per cent CaO, Na_2O, and K_2O were recalculated as mole per cent Or, Ab, An molecules assuming aluminium and silicon stoichiometry. All compositions have been plotted on the feldspar ternary diagram (fig. 2) which illustrates the extent of solid solution observed in a single hand specimen. Plagioclase compositions range from $An_{82.0}$ $Ab_{17.5}$ $Or_{0.5}$ to $An_{40.0}$ $Ab_{56.0}$ $Or_{4.0}$ with Or content increasing with decreasing An content. Alkali feldspar analyses range from $Or_{50}(Ab+An_{10})_{50}$ to $Or_{60}(Ab+An_{10})_{40}$. Points on the diagram indicating plagioclase

[1] The term inner mantle refers to the more calcic zone, which is an optically distinct unit within the mantle.

analyses more sodic than An_{40} represent mesostasis analysis occurring as intimate intergrowths with the alkali feldspar. The one point plotted in the middle of the ternary composition diagram probably represents either a determination on a boundary that was not visible optically or a cryptoperthitic intergrowth.
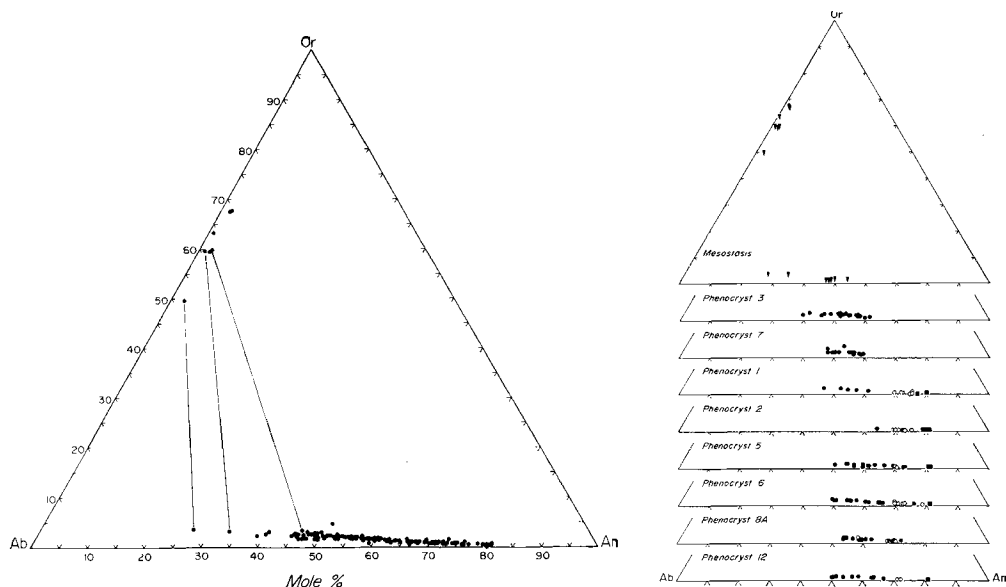

FIGS. 2 and 3: FIG. 2 (left). Composite feldspar ternary composition diagram for 214 spot electron-microprobe determinations on thirteen phenocrysts and some mesostasis areas in one rock at the 357·9 ft level of the sheet. The lines joining mesostasis compositions indicate compositions that occurred in close proximity and are meant to imply coexisting pairs but do not necessarily imply equilibrium coexistence. FIG. 3 (right). Microprobe determinations of compositions for separate feldspar phenocrysts and mesostasis. Positions within the crystal were used to divide the locations into four classes: ○, core; ■, inner mantle zone; ●, normally zoned mantle, and ▼, mesostasis.

Zoning of the feldspar phenocrysts precludes the use of average values and even presentation of separate analyses may be misleading if the position within the crystal is not identified. For example, examination of fig. 2 does not reveal any compositional gaps between phenocryst cores and mantles nor is it evident that optically the core consists of numerous small oscillation zones. If the classes established by positional criteria are kept separate and ternary plots made for each phenocryst, we obtain fig. 3, and a consistent pattern emerges. It should be emphasized that the data points are selected, i.e. any points that could not be assigned to one of the four classes on the basis of optical studies, and appeared likely to be composites of two classes due to overlap of the electron beam, have been eliminated. If we examine the plots for individual phenocrysts, nos. 1, 2, 5, 6, 8, and 12, we observe a restricted compositional range for the core of An_{79-67}. The inner mantle, which immediately surrounds the core, is An_{82-76} and the normally zoned mantle ranges from An_{73-40}. The scarcity of analyses

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in the mantle compositions from An_{80-65} is probably more apparent than real and is attributed to an insufficiency of data points on the steep chemical gradient defining the boundary between the inner mantle zone and the outer mantle. Optical observations and continuous electron–microprobe–traverse scans (fig. 1b) indicate a complete range of An compositions from An_{80} to An_{40} in the mantle. This is in contrast to the Picture Gorge basalt (Lindsley and Smith, 1969–70) where there is a compositional gap of (An_{74-70}) in the normally zoned mantle.

The five phenocrysts that do not have a prominent core are normally zoned from a maximum of An_{75} to the lowest observed composition on a phenocryst rim of approximately An_{40} (examples—phenocryst no. 3 and no. 7). These normally zoned
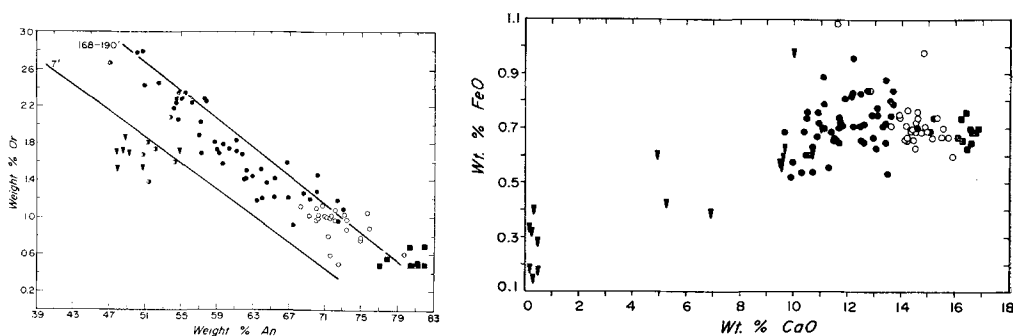
FIGS. 4 and 5: FIG. 4 (left). Relationship between % Or and % An of plagioclase. Solid lines are approximate fit for Makaopuhi Lava Lake feldspars for depth 7 ft and 168 to 190 ft (after Evans and Moore, 1968). Symbols are the same as fig. 3 with the additional class of half-filled circles denoting mantle compositions near the phenocryst boundary. FIG. 5 (right). Relationship between wt. % FeO and wt. % CaO for all data points. Symbols same as in fig. 3.

phenocrysts may either be sections through the mantle of phenocrysts having cores similar to that shown in fig. 1, or they may represent phenocrysts that nucleated after the cores and oscillation zones formed. In any case, the total range of compositions for the normally zoned phenocrysts such as nos. 3 and 7 (fig. 3) are within the range of mantle compositions established by continuous scans and optical observations on those phenocrysts with cores that appear to be approximately the same size; and therefore both are probably from the same population. The points representing mesostasis compositions show that both alkali feldspar and plagioclase compositions occur in the mesostasis. The lines joining mesostasis compositions in fig. 2 simply indicate compositions that occurred in close spatial proximity and are meant to imply co-existing pairs but not necessarily to imply equilibrium coexistence.

To facilitate comparison with plagioclase compositions from the Makaopuhi Lava Lake, Hawaii (Evans and Moore, 1968) the same compositions as used in fig. 3 have been plotted on an expanded Or–An diagram in fig. 4. Approximate lines representing Evans and Moore's data for plagioclase from the top (7 ft) and bottom (168–190 ft) of the lava lake are included for comparison. Compositions near the crystal periphery

(designated by half-filled circles) plot at an intermediate position between the meso-stasis and the majority of the 'mantle' compositions. The general trend of the pheno-cryst mantle and inner mantle zone compositions differ from the representative slopes for the 7 ft and 168 to 190 ft curves from the Makaopuhi Lava Lake. Plots represent-ing mesostasis feldspars and the phenocryst periphery are lower in Ab than any compositions given in the lava lake study. It is apparent that there is a wider range in feldspar compositions from a single sample than there was for all levels of the Maka-opuhi Lava Lake section. For example, at a fixed Or content of $\approx 1\cdot 8\%$ we have plagioclases of composition An_{62-47}.

In fig. 5, weight per cent FeO is plotted versus weight per cent CaO. Iron is at a minimum concentration in the Or-rich alkali feldspars and highest in plagioclases more calcic than 10 wt. % CaO. The core, the inner mantle zone, and the more calcic portion of the mantle have a nearly constant iron content. Apparently at the last stages of crystallization, as represented by compositions of the mesostasis and the periphery of the phenocrysts, a lower concentration of iron is incorporated into the plagioclase structure. Additionally it appears that in the mesostasis the iron is preferen-tially incorporated into the plagioclase rather than the alkali feldspar.

Considerations on the crystallization of plagioclase

The following discussion serves to consolidate some principles governing the crystallization of plagioclase. Many of these principles have been discussed in part in published works (Harloff, 1927; Bowen, 1913, 1928; Carr, 1954; Vance, 1962, 1965; Bottinga et al., 1966; Wiebe, 1968; Yoder, 1969). However, the previous discussions do not individually provide a clear concept of the factors involved in growth zoning of plagioclase.

Pressure changes at constant composition. Although the exact shapes of the liquidus and solidus curves are not known for magmatic systems, some constraints on the events giving rise to the zoning can be obtained from a consideration of the binary system Ab–An at different pressures. In this section we are concerned only with the equilibria between melt and crystals through and immediately after the event of a pressure change under conditions approximating constant composition. Under these circum-stances, the changes in the liquidus and solidus curves can be considered in binary projection.

In a general case of pressure reduction by vertical movement, the magma will undergo cooling through heat exchange with the environment, volume expansion, and possible resorption of crystals. Although the effect of volume expansion is small, and that of resorption of crystals variable from case to case, all these factors operate to lower the temperature of the rising magma.

Anhydrous systems. The Ab–An phase relationships for 1 bar and 10 kb pressure under anhydrous conditions (after Lindsley, 1968) are given in fig. 6. If liquid A is in equilibrium with crystals of composition B, a drop in pressure can give rise to two limiting cases:

If some of the crystals were resorbed as a result of the pressure drop, the remaining crystals could react with the liquid to re-establish equilibrium at approximately A′

and B'. The heat energy required to resorb some plagioclase will result in a temperature drop. For each per cent by weight of plagioclase An_{70} that melts in the system, the magma would cool adiabatically about 2 °C assuming the heat of crystallization of An_{70} to be $\approx$ 60 cal/gm (Kracek and Neuvonen, 1952) and the heat capacity of the magma to be $\approx$ 0·30 cal/(gm. °C) (Leonidov, 1967). A system that is largely crystalline could change its liquid composition appreciably with the least amount of melting
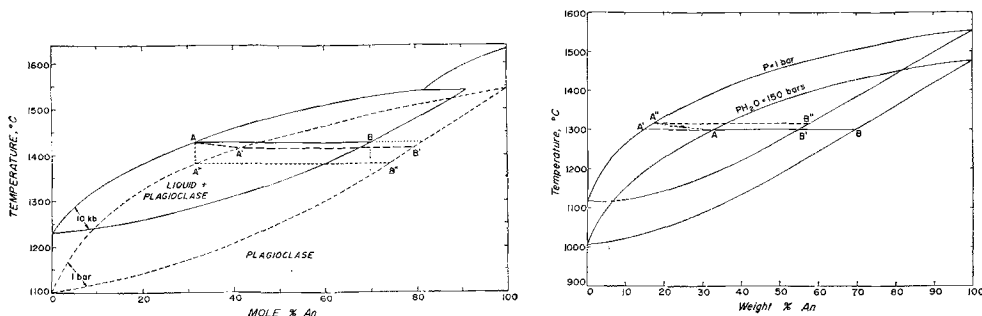
FIGS. 6 and 7: FIG. 6 (left). Initial equilibrium at 10 kb between liquid A and plagioclase B. A subsequent pressure reduction to 1 bar with consequent melting of some solid could lead to crystallization of plagioclase B' when the liquid composition reached the liquidus surface at A'. The exact position of the resulting tie-line A'B' depends on the thermal history of the system during and immediately after the pressure drop. Plagioclase B' could crystallize on cores of unmodified B or cores that have been completely or partially modified by liquid–crystal equilibration. If no melting occurs, the liquid A and plagioclase B remain unchanged until the system cools to the lower-pressure liquidus at A''. Plagioclase B'' will then crystallize on core of plagioclase B because no nucleation can occur at A''. (Ab–An system after Lindsley, 1969.) FIG. 7 (right). Initial equilibrium at 150 bars P_{H_2O} between liquid A and plagioclase B. A subsequent pressure reduction to 1 bar (anhydrous) would result in liquid A being supercooled by 100 °C. If the temperature of the system was not lowered below 1300 °C by heat loss, and the degree of supercooling did not reduce the crystallization rate to zero, the super-cooled system would crystallize plagioclase B' either on cores of plagioclase B or form nuclei of B' or both. The heat of crystallization could increase the temperature giving rise to a liquid path A to A' and plagioclase compositions ranging from B' to B''. (Ab–An system after Yoder, 1969.)

and the process could be essentially isothermal. This case will give rise to the largest difference theoretically possible between the composition of plagioclase before the pressure drop and after. If sufficient of the plagioclase melted to enable the liquid composition to reach the liquidus surface, but no reaction took place with the remaining plagioclase, heat loss from the system could cause crystallization of more anorthitic plagioclase on corroded cores of pre-existing but otherwise unaltered plagioclase. Reported resorbed grain boundaries that cut across oscillation zones in plagioclase (Bowen, 1928; Fries, 1939; Vance, 1965) attest to the fact that the rate of resorption can exceed the rate of diffusion in plagioclase in magmatic systems.

On the other hand, the heat loss from the system could be sufficiently rapid to prevent melting of the crystals and crystallization of plagioclase B'' would occur on unmodified pre-existing plagioclase crystals when the liquid reached A'' (fig. 6). The

exact composition of plagioclase B″ would be dependent on the relative shapes of the liquidus and solidus envelopes (cf. Carr, 1954; Vance, 1965). Because it is probable that the 'thinning' of the projected liquidus and solidus envelope occurs with higher pressure in a more complex system, the illustration is believed to be correct to the extent that the plagioclase at lower pressures would be more anorthitic in both limiting cases. The plagioclase crystallizing in this second case would have a considerably lower anorthite content than the plagioclase resulting from the previous case. Any supercooling of the magma before the resumption of crystallization would further decrease the anorthite content of the plagioclase to the extent that the plagioclase would be either the same composition or more albitic than that which crystallized at the higher pressure.

The maximum change in composition that can be expected for an anhydrous pressure differential of 10 kb from the conditions of fig. 6 would be about 10 wt. % An, an average ΔAn of 1 % per kb.

Hydrous systems. In a system containing H_2O but undersaturated with respect to water throughout the drop in confining pressure, the change in the liquidus and solidus surfaces would be in the same direction as that in the anhydrous system and would be subject to similar considerations as the anhydrous case.

As noted by Yoder (1969), relatively small pressure changes in H_2O-saturated systems can lead to significant changes in the composition of crystallizing plagioclase. A sudden reduction of pressure in a saturated system is equivalent to supercooling and results in the precipitation of more albitic plagioclase after the pressure drop. In fig. 7 the path A–B to A′B′ illustrates a H_2O-saturated system undergoing a confining pressure drop from 150 bar to 1 bar. Plagioclase subjected to a pressure reduction during growth in a H_2O-saturated system would have a more albitic outer zone grown directly on the unmodified growth surface of the core. Crystallization could be rapid for moderate supercooling because the rate of crystallization is not controlled by the rate of heat loss from the system until the liquid reaches the lower pressure liquidus and because the initial liquid composition (A) is closer to the solidus composition (B′) after the pressure drop than at equilibrium (A–B or A′–B′). In fact, the temperature of the system may actually rise (A″–B″) if the crystallization were sufficiently rapid relative to heat loss.

Anhydrous or H_2O-undersaturated systems cannot produce nuclei solely through a reduction in pressure whereas saturated systems can be expected to do so.

Crystallization under constant pressure in anhydrous systems. Bottinga et al. (1966) have presented the argument, following Harloff (1927), that the composition of plagioclase crystallizing during a single cycle in oscillatory growth is controlled by the changing composition of the melt at the crystal–melt interface. This change in the melt composition is brought about by the progressive development of concentration gradients whenever the growth rate is greater than the rate of supply of growth components from the melt by diffusion. If the concentration gradient is destroyed by movement, or if crystallization ceases and diffusion of the components obliterates the gradient, resumption of crystallization would repeat the oscillation. Bottinga et al. (1966) have pointed out that crystallization under these conditions is effectively

diffusion-controlled. Since diffusion-controlled crystallization will occur only under restricted conditions, i.e. when the rate of heat removal is approximately equal to the rate of heat production from crystallization, it is therefore necessary to examine crystallization under various rates of heat removal.

In the case that crystal growth is controlled by heat removal from the system, the melt composition will effectively remain on the liquidus surface and will represent the simplest case of non-equilibrium crystallization. Under these conditions the rate of heat removal from the system would be slower than the potential rate of heat production from crystallization. The variation in the rate of heat removal for simple nonequilibrium crystallization can vary between a rate allowing solid–melt reaction (equilibrium crystallization) and a rate at which diffusion in the melt controls crystal growth. The difference in the diffusion coefficients of plagioclase components in melts and in solid plagioclase indicates a broad range of cooling rates under which nonequilibrium crystallization controlled by the rate of heat loss will occur. Under these conditions, no irregularities or departures from progressive normal zoning can occur.

Where cyclic oscillatory zoning has occurred, the rate of heat removal must have been equal to the rate of heat production during crystallization to maintain melt composition control on the crystallizing plagioclase; i.e. essentially isothermal crystallization. Furthermore, the intercrystal melt must have been maintained in a supercooled condition to allow sufficient compositional difference between the liquidus composition and the composition of the intercrystal melt to permit significant zoning to result.

If, however, heat removal was greater than heat production from crystallization, the system would become progressively more supercooled and the composition of the crystallizing plagioclase would effectively be controlled by variation of both temperature and composition of the melt adjacent to the growing crystal, giving rise to oscillatory perturbations superimposed on normally zoned plagioclase.

It is possible to show that diffusion-controlled crystallization would not be effective for a near-surface intrusion under the more extreme conditions of progressive supercooling. The plagioclase concentration of a melt containing about one-half by weight of calcic plagioclase is about 5×10^{-3} mole/cm^3. If the diffusion rate of the slowest component is $\approx 10^{-11}$ cm^2/sec (Towers and Chipman, 1957) and the thermal conductivity is 4×10^{-3} (lime-glass, Kingery, 1960), diffusion to provide a linear growth (perpendicular to a crystal face) of 1 μm for a concentration gradient of 10^{-3} mole cm^{-3} mm^{-1} under steady state conditions would occur in ≈ 1 year. The linear growth rate of 1 μm/yr approximates the average growth rate of plagioclase in the central part of the 225 ft thick Makaopuhi Lava Lake (Evans and Moore, 1968). The crystallization of 1 μm of calcic plagioclase liberates about 10^{-2} cal/cm^2 of heat energy. The removal of this heat energy from the interface in one year (heat energy removal equal to heat energy produced from diffusion-controlled crystallization) would occur if the thermal gradient in the vicinity of the crystal was 10^{-7} °C/cm. This calculated value for a thermal gradient would only be achieved in deep-seated magmas or for relatively short periods of time in the centre of shallow intrusions and thick extrusive bodies. For example, thermal gradients obtained from calculated time–temperature–depth

profiles for the Makaopuhi Lava Lake range from ≈ 1 °C/cm to $\approx 10^{-2}$ °C/cm five years after intrusion (Evans and Moore, 1968). It is probable, therefore, that extrusive and shallow intrusive bodies crystallize under conditions of progressive supercooling. Yet many of these igneous bodies contain normally zoned plagioclase or plagioclase with normally zoned rims (Vance, 1962; Evans and Moore, 1968; Wiebe, 1968, 1973), similar to the normal zoning that could be theoretically produced under simple non-equilibrium conditions where crystallization at the liquidus was heat-conduction controlled. Normal zoning is also observed in plagioclase from lavas where the conditions for progressive supercooling are even more certain. Under conditions of extreme progressive supercooling, compositional perturbations in the melt would cease to cause any observable alterations to normal zoning in plagioclase crystals; i.e. temperature control of the crystallizing plagioclase composition. This effect may be ultimately related to the 'early' nucleation of all solid phases under extreme super-cooling resulting in an apparent univariant condition in the system.

In summary the zoning patterns of plagioclase may be separated into four categories according to the relative effectiveness of the rates of diffusion in the magma and heat removal from the system. These are listed below in order of increasing rate of heat removal:

a. Crystallization is controlled by rate of heat conduction under conditions where the rate of heat removal is slower than the diffusion rate of components in the melt. No concentration gradients exist in the intercrystal magma and the zoning is normal without perturbations.

b. Crystallization is controlled by the diffusion rate of plagioclase components in the magma when the rate of heat removal is equal to the rate of heat production from crystallization. Concentration gradients can exist in the intercrystal magma and the composition of crystallizing plagioclase is controlled by the magma composition at the crystal–melt interface under conditions approximating constant temperature. The zoning could be oscillatory with a cyclic compositional range over a limited interval of crystallization.

c. Crystallization is controlled by the diffusion rates in the melt and the rate of heat removal. The relative rates must be such that crystallization proceeds at an approximately constant degree of supercooling. The resulting composite zoning could be oscillatory superimposed on a normally zoned trend.

d. Crystallization is controlled by rapid heat removal leading to progressively higher degrees of supercooling. The composition of crystallizing plagioclase would be largely controlled by temperature and apparently results in normally zoned plagioclase without apparent perturbations, as is evidenced by plagioclase in rapidly cooled igneous bodies.

Interpretation

Any model proposed for the plagioclase crystallization history of this rock unit must be consistent with the presence of a core, a more calcic inner mantle zone, a normally zoned mantle in the phenocryst plagioclases, and the crystallization of two feldspar phases in the final stage. In general, the feldspars must have crystallized in

a H_2O-undersaturated environment (no observed primary hydrous phases) after emplacement as a sheet at quite shallow depths (intercalated with surface flows) and hence low load pressures (maximum 100 bar). Extensive blocks of volcanic rocks that have sunk into the sheet indicate that the roof rocks were extensively fractured and that significant volatile pressures were unlikely.

The sharp boundary between the core and the calcic inner mantle represents a feature of plagioclase phenocrysts throughout the sill and, therefore, records an event that affected the whole system that now constitutes the sheet. This event caused a change from the oscillatory zoned core (similar composition cycles) of An_{79-67} to the inner mantle composition of An_{82-76}; an increase of about 10 wt. % An, which would correspond to an anhydrous pressure differential of -10 kb assuming that anhydrous conditions leading to a maximum change in composition existed.

Outer boundaries of the cores are generally euhedral but occasional subhedral core outlines suggest limited resorption prior to crystallization. The anorthitic zone ($\approx An_{82-76}$) of the inner mantle is generally euhedral. If the heat loss during ascent was relatively rapid, only partial resorption would result and primary oscillatory zoning would be retained in the cores. A more calcic plagioclase zone would then be deposited around the core when crystallization resumed. As the inner mantle crystallized more or less directly onto the core, the indicated pressure change may represent a minimal estimate.

The crystallization of the cores, therefore, would have taken place at a minimum depth of at least a few tens of kilometres. At this depth, the core formed under conditions at which the crystallization was controlled by the diffusion rate of components (category b) in the melt and minor oscillations in the phenocryst cores are probably due to component gradients established around crystals during growth. The compositional range of the oscillation zones reflects the range of compositional gradients developed at the crystal–liquid interface during crystallization and resulted from the alternating establishment and destruction of the gradients. Although it is possible that changes in the controlling gradients can occur without physical disruption (see Vance, 1965; Bottinga et al., 1966) it is likely that in an actual magma there is sufficient differential movement at times between crystal and liquid to affect gradients established by diffusion.

Following intrusion of the magma into a near surface environment, the inner mantle zone (An_{82-80}) crystallized. In this environment thermal gradients would be such that crystallization would be taking place under progressively higher degrees of super-cooling (category d). The composition of the crystallizing plagioclase would be controlled by the temperature, resulting in a normally zoned plagioclase without apparent perturbations. There is little doubt that crystallization of the mantle was continuous after emplacement without further sudden changes in intensive parameters. The crystallization would appear to have occurred under conditions of a high rate of heat removal resulting effectively in temperature control of the crystallizing plagioclase. These conditions are envisaged to be broadly similar to those resulting in the normally zoned plagioclase in the Makaopuhi Lava Lake (Evans and Moore, 1968) even though the cooling rate in the lava lake would be somewhat faster than in the diabase sheet.

The continuous normally zoned mantle terminates against the mesostasis, which represents the final liquid residuum. At a plagioclase composition of $\approx An_{40}$ the liquid reached the boundary surface resulting in co-crystallization of alkali and plagioclase feldspars.

In summary the crystallization history may be interpreted from a synthesis of compositional and textural features observed in the plagioclase phenocrysts. The phenocryst cores crystallized in a P_{H_2O}-undersaturated environment under a few tens of kilobars pressure. Minor zonation (An_{79-67}) of the phenocryst cores is a function of compositional gradients at the crystal–liquid interface. Following a marked decrease in total pressure (intrusion at shallow depths), a more calcic shell of $\approx An_{82-80}$ crystallized around the core. Finally the normally zoned outer mantle crystallized under conditions of progressive supercooling at relatively low total pressures in a relatively quiescent environment.

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