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THE BUSHVELD COMPLEX:
A TIME TO FILL AND A TIME TO COOL

R.G. CAWTHORN AND F. WALRAVEN

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A TIME TO FILL AND A TIME TO COOL

by

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ABSTRACT

Calculations of the rate of cooling and consequent mineral differentiation trends in the Bushveld Complex place constraints upon the rate at which the chamber was filled. These calculations show that the 8 km-thick Bushveld Complex crystallized in a little over 150,000 years, but that magma addition ceased after 60,000 years. The volume of magma exceeded 384,000 km^3, giving an emplacement rate of over 6 km^3 per year. Such eruption rates far exceed any known rates related to mantle plumes.

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INTRODUCTION

Almost all studies on layered intrusions have focused on the processes and consequences of cooling of magma once emplaced into the crust. Little attention has been given to the rate of emplacement and filling of the magma chamber. With reference to the Bushveld Complex, Wager and Brown (1968) recognised an "Integration Stage" during which the chamber was being replenished with magma more quickly than it was cooling. This interpretation was based on the near-constancy of mineral compositions through large vertical intervals of the lower half of the intrusion (Figures 1 and 2). This begs the question as to how quickly the chamber was inflated. The purpose of this paper is to use a thermal model to address the rate of magma injection and compare it to flood-basalt eruption rates.

THERMAL MODELS

In a series of papers culminating in 1968, Jaeger presented a mathematical treatment for the cooling of magma bodies. While these models allowed estimates of the time for a magma to crystallize and the temperature regime outside intrusions to be determined, they had certain limitations when applied to large layered intrusions. One of these limitations is that, when crystals accumulate on the floor, the upper part of the magma chamber may have a near-uniform temperature due to convection, and as a consequence heat loss through the roof is far greater than through the floor. Irvine (1970) developed a more advanced set of equations to handle such processes. As a result of these refinements the rate of accumulation could be predicted (Irvine, 1970; his Fig. 5). He showed that the rate of accumulation varied linearly with the elapsed time. Using his diagrams the 8 km-thick Bushveld Complex would have solidified in about 100,000 years.

Magma addition (and extraction) are important processes in the evolution of layered intrusions. Such events cannot be included in the formulations of Jaeger or Irvine. Consequently, a different approach to the modelling and calculations is adopted here, which permits the incorporation of such processes. The thermal modelling procedure is as follows. The entire thickness of crust considered to be involved in the cooling process is divided into a number of horizontal cells of equal thickness. A temperature is assigned to each one (equivalent to a geothermal gradient). A magma, of known temperature (and known liquidus and solidus temperatures), is added at the required depth and subdivided into horizontal cells comparable in thickness to those in the floor and roof. Thus, at the initiation of the calculation each cell (depth interval) has a predetermined, uniform temperature, *i.e.* temperature profiles are stepped rather than smooth. The temperature differences between each cell and the overlying and underlying cells are used to calculate the heat flow, and changes in heat content between cells for a small time increment:

$$\Delta H = K.t.(T_2 - T_1)/D_s$$

and $\Delta E = \Delta H_2 - \Delta H_1$

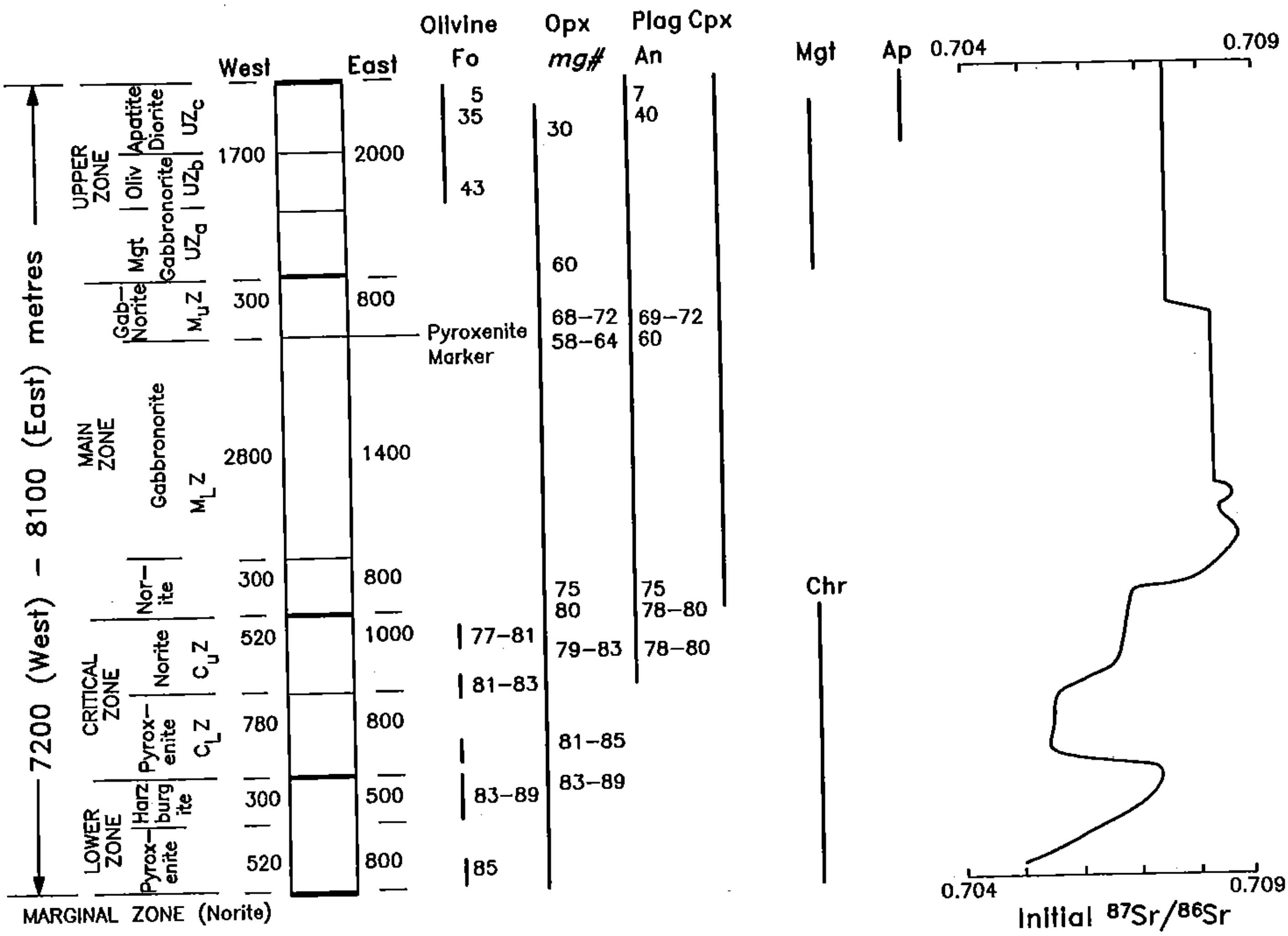

Figure 1: Simplified stratigraphy of the eastern and western limbs of the Bushveld Complex, showing the appearance of cumulus minerals and the initial $^{87}Sr/^{86}Sr$ (R_o) ratios. Compiled from references in Eales and Cawthorn (1996) and Kruger (1994).

where ΔH is the heat transferred from one cell to its neighbour, K is the thermal conductivity, t is the time interval, T_1 and T_2 are the temperatures of adjacent cells, D_s is the thickness of the cell, and ΔE is the change in heat content from one cell to the next.

New temperatures for each cell are calculated from their heat contents using different equations for the cases where the cells are totally solid, totally liquid or partially molten, *e.g.*

$$T = E/C.D_s.\rho \text{(totally solid)}$$

and $$T = (E - L.D_s.\rho)/C.Ds.\ ) \text{ (totally liquid)},$$

where E is the heat content of a cell, C is the specific heat, ρ is the density, and L is the latent heat of solidification. Sequential heat flow and temperature changes are calculated for successive time increments until the total time elapsed is that required for the specific stage of the model.

This calculation is an approximation, which increases in accuracy as the number of cells is increased and as the time interval for each step is decreased. The significance of this method is that, at any stage of the cooling process, magma can be added or subtracted to mimic inferred magmatic processes.

Two options in this calculation procedure can be adopted. If the cells in the magma chamber are considered not to convect, but merely solidify in situ, the process approximates to that of Jaeger (1968), where heat transfer is purely by conduction, and would be appropriate for relatively thin sills. Alternatively, it is possible to combine all the cells in the magma chamber into a single cell at a uniform temperature, producing conditions approximating to rapid convection and comparable to the model of Irvine (1970). The latter is used in the following calculations.

APPLICATION TO THE BUSHVELD COMPLEX

The stratigraphy and range of mineral compositions in the Bushveld Complex is shown in Figure 1. Magma replenishment within this intrusion can be identified in two ways. Initial $^{87}Sr/^{86}Sr$ (R_o) ratio data (Kruger, 1994) suggest that the entire intrusion can be subdivided into three magmatic lineages, the ultrabasic Lower and Critical Zones (R_o = 0.705-0.707), and tholeiitic events producing the Lower Main Zone (R_o = 0.709) and Upper Main Zone and Upper Zone (R_o = 0.707). These breaks in ratio are relatively abrupt, persist vertically and are traceable laterally, and leave little doubt that there has been addition of significant volumes of fundamentally different magma composition. With regard to the Lower and Critical Zones, the oscillating, but gradually decreasing, bulk-rock and mineral *mg#* (100*Mg/(Mg+Fe)) values shown in Figure 2 suggest periods where addition of less differentiated, but isotopically similar magma occurred (Eales & Cawthorn, 1996).

The variation in *mg#* for whole-rock and mineral composition (alternating olivine and orthopyroxene) can be converted to a trend of decrease and increase in temperature related to fractionation and magma addition, by using the known phase relations of the magma

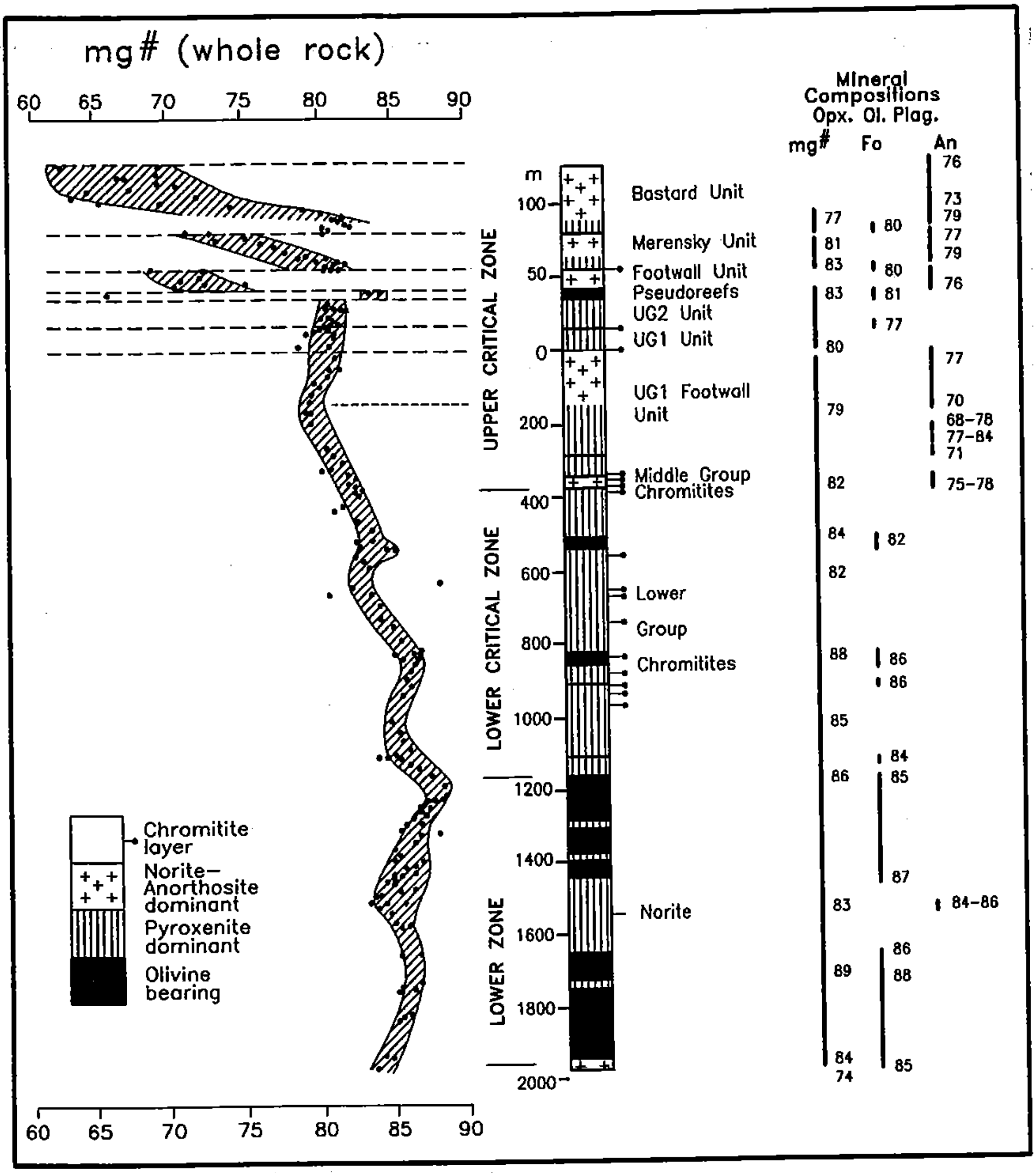

Figure 2: Section through the Lower and Critical Zones illustrating cyclic variations. Whole rock mg# calculated assuming a Fe_2O_3:FeO ratio of 0.1. Note scale change at base of UG1 (Upper Group 1 chromitite layer).

parental to this part of the intrusion. Experimental studies by Cawthorn and Biggar (1993) indicated a liquidus temperature of 1300°C. Crystallization of minor olivine was superseded by orthopyroxene until 1180°C when plagioclase appeared. Quantitative modelling using glass compositions from the experiments indicated that approximately 20% crystallization occurred before plagioclase appeared. Use of the changing orthopyroxene compositions is consistent with this degree of fractionation, changing from a maximum of En_{89} near the base of the Lower Zone to En_{82} where plagioclase appears in the middle of the Critical Zone (Figures 1 and 2). Modelling of fractionation of plagioclase plus orthopyroxene indicated a further 20% fractionation to the level of the Merensky Reef, very close to the top of the Critical Zone (Figure 2) where the orthopyroxene composition is En_{79}, and the temperature 1140°C. The combined Lower and Critical Zones have a variable thickness, but on average are approximately 2.5 km (Figure 1), and as they represent 40% fractionation demand a total thickness of 6 km of magma.

The emplacement and cooling of magma producing the Lower and Critical Zones are modelled by assuming magma is added in six batches each 1 km thick, with each addition taking place after the temperature has fallen to 1200°C. After the first increment of 1 km of magma the mineral composition decreases from En_{89} to En_{84} (Figure 3) within 400 years and 200 m of cumulates had formed, at which time another 1 km-thick batch of magma was added. After five repetitions, over 1 km of orthopyroxene-dominated cumulates formed, which approximates to the Lower and Lower Critical Zones. The intrusion was then allowed to fractionate to produce the plagioclase-orthopyroxene cumulates of the Upper Critical Zone as the orthopyroxene evolved to En_{79}. Using the multiple addition model, it can be seen from Figure 3 that the entire 2.5 km-thick Lower and Critical Zones formed within 24,000 years.

The cyclicity of dunite-harzburgite-pyroxenite within the Lower Zone (Cameron, 1978) and the pyroxenite-norite-anorthosite in the Upper Critical Zone (Eales & Cawthorn, 1996) could be interpreted as indicating far more than six additions of magma. Increasing the number of additions and decreasing their individual thicknesses in these calculations leads to a more rapid accumulation in the early stages of the layered sequence than calculated in Figure 3, but the final difference in total time to accumulate the Lower and Critical Zones is not reduced substantially.

At the level of the base of the Main Zone there is a dramatic change in R_o, indicating substantial addition of magma, and at this stage in the computed model a further 2 km of magma was added. This was allowed to cool and fractionate to produce a further 2.5 km of cumulate of the Lower Main Zone. As there is now a total thickness of 8 km of combined magma and cumulate, and as the country rocks have been significantly heated during formation of the Lower and Critical Zones, the cooling rate for the Lower Main Zone is much slower than for the underlying zones, and a period approaching 40,000 years is required (Figure 4).

The Pyroxenite Marker defines the boundary between Lower and Upper Main Zones and has been identified as an interval where magma addition occurs (Sharpe, 1985). There is 8 km of magma and cumulate rock already in the chamber at this stage, and the total thickness of the intrusion is also 8 km (Figure 1). Thus, at this level it is envisaged that

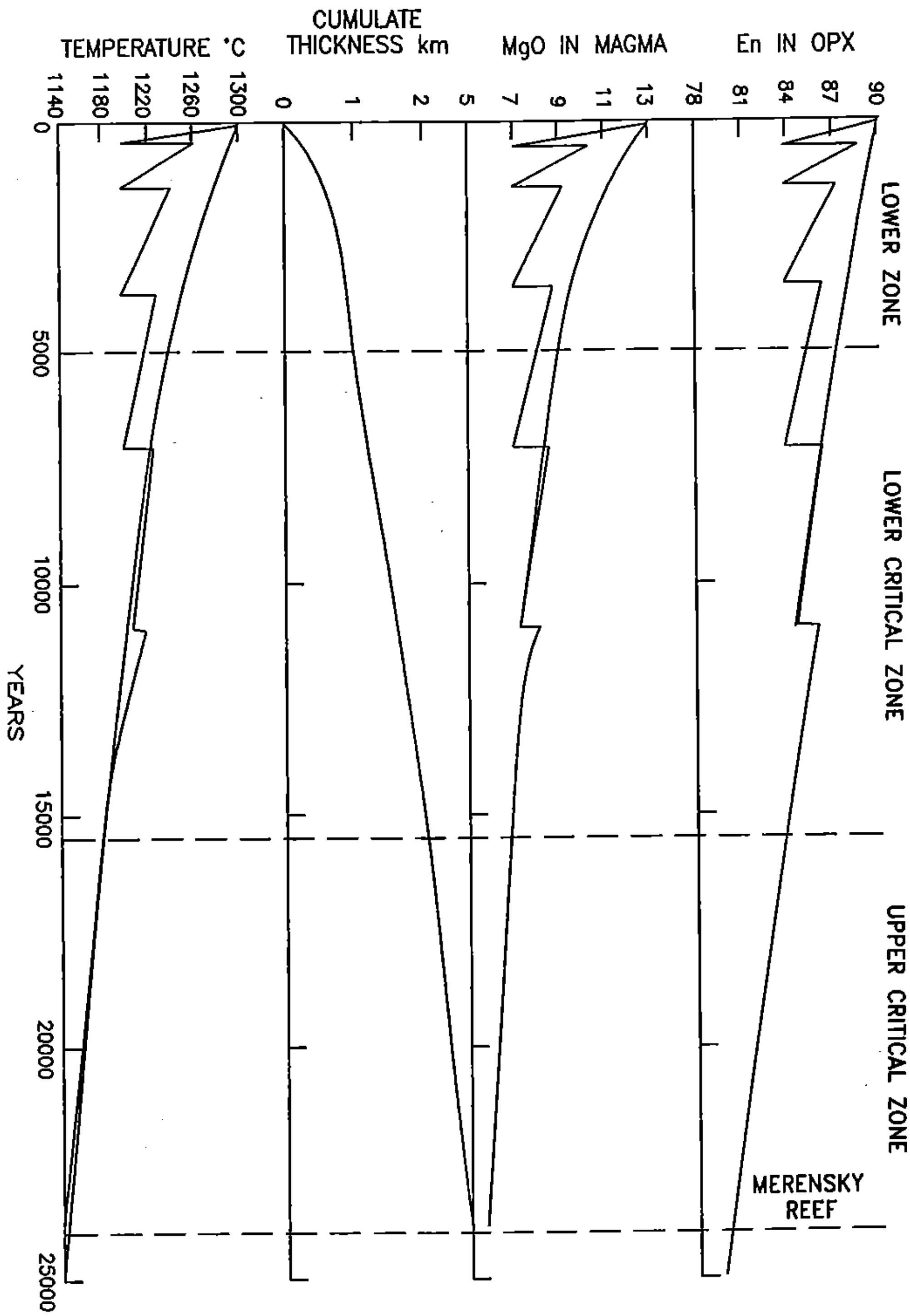

Figure 3: Model for emplacement and crystallization of the Lower and Critical Zones of the Bushveld Complex. Four different parameters relating to the magma and the accumulated rock are plotted against time. In each diagram the trends for two models are plotted, one involving addition of a single pulse of magma, and the other involving addition of six discrete 1 km-thick events. Liquidus temperature is 1300°C and cools to 1200°C crystallizing mainly orthopyroxene which changes from $En_{89\text{-}84}$ and the magma changes from 12.5% to 7% MgO before recharge. With each addition of magma, complete mixing is assumed and hence the maximum temperature, En, and MgO contents after each addition decrease.

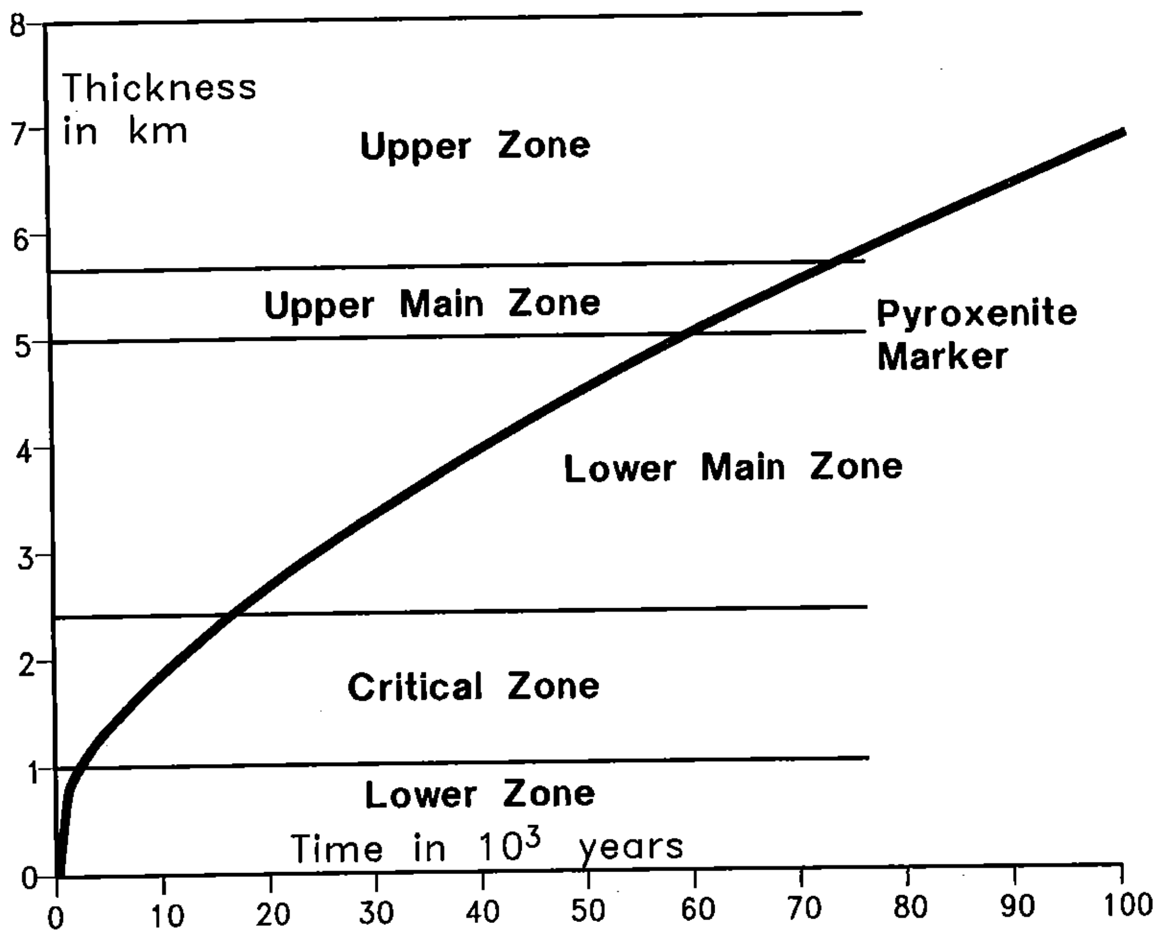

Figure 4: Model for crystallization of the entire Bushveld Complex. Trends for Lower and Critical Zone are taken from Figure 1. The Pyroxenite Marker represents the last addition of magma into the intrusion (Kruger et al., 1987).

lateral expulsion of magma occurred concurrently with magma addition so that there is no net increase in vertical extent of the chamber. For the purposes of this model no net addition of magma is introduced into the calculation. It could be argued that the new magma was slightly hotter than the resident magma, but in view of all the other uncertainties in this model, such an effect is considered trivial. The cooling rate continues to decrease as magma temperature falls and country rock temperatures remain high, and the time taken for total solidification (at 900°C) of the Upper Zone is in excess of 80,000 years (Figure 4).

These calculations show that the entire intrusion crystallized in over 150,000 years, slightly longer than predicted by the single injection model of Irvine (1970). However, of greater relevance here is the fact that the isotopic uniformity of the Upper Zone has been interpreted by Kruger *et al.* (1987) to indicate that no further addition of magma occurred above the level of the Pyroxenite Marker. Hence, addition of magma was complete after some 60,000 years.

VOLUME OF MAGMA

The Bushveld Complex is by far the largest preserved layered intrusion. The main limbs are exposed over an area of 64,000 km^2 (Wager & Brown, 1968). However, there are several small outliers and borehole evidence for hidden occurrences of layered mafic rocks, presumably consanguineous with the main exposed lobes (Eales & Cawthorn, 1996) and so this can be confidently regarded as a minimum area. The maximum thickness of layered rock is 8 km (Figure 1), but not all units all equally developed. The Lower and Critical Zones are restricted in lateral extent, while the Main and Upper Zones can be traced over the entire exposed area (Eales & Cawthorn, 1996). Assuming an average thickness of 6 km, the volume estimate becomes 384,000 km^3. This estimate excludes down-dip extensions, and can be considered a conservative estimate of the volume of layered rocks. For this volume of magma to be emplaced within the calculated time interval of 60,000 years the average emplacement rate must have exceeded 6 km^3 per year.

COMPARISON WITH PLUME MAGMATISM

Periods of relatively short but intense eruptive activity and emplacement of basic magma are normally attributed to mantle plumes. Estimates of the volumes of magma related to such plumes are difficult to obtain, and estimates of the time period for emplacement are even more problematic. Hence, comparison between the Bushveld Complex and other plume products in terms of volumes and eruption rates is difficult. Nevertheless, some comparisons can be attempted. The Hawaiian chain volcanicity is the best example of plume activity on the ocean floor, and its relative geological and geometrical simplicity make calculations plausible. Clague and Dalrymple (1987) and Dzurisin *et al.* (1984) calculated magma production rates of 0.01 km^3 per year.

The Columbia River Basalts provide the best example of recent plume activity on continents. The total volume erupted was 175,000 km^3 (Hooper, 1988), most of which was erupted during a 2 Ma period, yielding an average eruption rate during this peak of 0.09 km^3 per year, although there were probably shorter periods of more intense activity. However, igneous activity continued for a further 10 Ma.

Richards *et al.* (1989) suggested that flood basalt represented melt from the head of a plume, which was produced at rates of 1 km^3 per year, whereas linear tracks of magmatic activity were derived from the tail of a plume at rates of only 0.02 km^3 per year. Campbell *et al.* (1992) also suggested a rate of 1 km^3 per year for the Deccan and Siberian Traps. These estimates of 1 km^3 per year are unconstrained as the period of eruption cannot be determined sufficiently precisely.

In terms of intrusive activity, the Duluth Complex provides the best example of plume-related magmatism. Several discrete intrusions can be identified within this suite, together with considerable volcanic outpouring (Miller & Ripley, 1996). The intrusions spanned an age range of 11 Ma from 1107-1096 Ma, although most bodies may have been emplaced within 2 Ma (J.D. Miller, personal communication). As a result, each shows evidence of intruding, truncating and displacing immediate predecessors. The time interval between each event was such that the previous body was solid enough to be disrupted by the next. Furthermore, the locus of emplacement evidently moved in order to produce the discordant relations. In the case of the Bushveld Complex, neither of these processes can be identified. Emplacement of subsequent pulses occurred after extremely small proportions of previous magma had crystallized as evidenced by the small oscillations in mineral composition. With the exception of two discordant relations in the Western Bushveld, whose origin need not necessarily be related to addition of magma (Wilson *et al.*, 1994), magma piled up vertically, rather than disrupting the previously accumulated sequence. The exact loci of feeders to the Bushveld Complex remain to be identified, but there are no observations to suggest that they migrated systematically with time. Furthermore, nowhere is the Upper Zone cross-cut by younger bodies suggesting that there was no further igneous activity after the Upper Zone had crystallized, indicating an extremely abrupt termination to mafic magmatism after less than 60,000 to 150,000 years from initiation of the Bushveld Complex.

CONCLUSIONS

By assessing the time intervals between interrupted crystallization sequences in the Bushveld Complex it is possible to place quite precise constraints on the timing of filling of the magma chamber. This thermal modelling requires that the entire intrusion crystallized in a little over 150,000 years, but that magma addition terminated abruptly after about 60,000 years. The volume of magma emplaced during this period was at least 384,000 km^3, indicating injection rates at least two orders of magnitude faster than observed at Hawaii or in the Columbia River Basalts, and close to one order of magnitude greater than even the fastest rates proposed for flood-basalt activity.

Estimated rates for flood-basalt eruption are very poorly constrained as age determinations cannot be made as precisely as would be required for this kind of calculation. However, if the Bushveld Complex was generated by the same kind of plume activity as is proposed for more recent flood basalts the precise time span determined here for the emplacement of magma into the Bushveld Complex demands a reassessment of inferred rates of eruption for flood basalts.

Discordant emplacement features are extremely rare in the Bushveld Complex, in contrast to the plume-related Duluth Complex, and suggest that the igneous activity in the Bushveld Complex abruptly terminated in less than 150,000 years after initiation. In this regard, the Bushveld magmatic episode appears different from more recent plume-related activity.

*(Footnote - a copy of the thermal modelling computer program is available from Dr F Walraven.)

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