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Implications of mantle plume structure for the evolution of flood basalts

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ABSTRACT

Morgan [1,2] suggested that continental flood basalts appear as the first volcanic expression of new mantle plumes. Experimental studies in viscous liquids have shown that new, or "starting", plumes should consist of a large bulbous head followed by a narrow feeder conduit. Analysis of the plume flow [3] indicates that, if the plume ascent is driven by thermal buoyancy, the head will entrain the surrounding mantle as it rises. The head cools and enlarges, and develops a compositional zonation of source and entrained mantle. Uncontaminated, relatively hot material from the plume source continues to flow up the trailing conduit making the temperature of the plume axis greater than that of the remainder of the head. We explore the implications of this plume structure by comparing the physical and chemical characteristics of two flood basalt provinces (the Deccan and Karoo) with predictions of the dynamical model.

The chronology, tectonics and geochemistry of the two provinces all fit well with the starting plume hypothesis. We attribute the sudden onset and short duration of continental flood volcanism, over an equant area 2000–2500 km across, to melting the plume head and its subsequent decline to a narrow chain of volcanic activity, ~ 200 km wide, to melting in the plume tail. A surface uplift of 500–1000 m is predicted but this gives way to subsidence due to lateral spreading of the plume head before the onset of the main period of volcanism. A period of enhanced subsidence is then predicted to occur as magma escapes from the mantle and loads the earth's surface, followed by slow subsidence over 10^9 years as the plume's thermal anomaly gradually decays. The timing and duration of volcanism has not been predicted with certainty, but activity is expected to begin as a burst and to die away rapidly over a total time of order 20 Ma, in agreement with dating which indicates that the bulk of the magmas in each province were ejected within 2–3 Ma and followed by smaller volumes over a further 5–10 Ma. The model predicts that the high-temperature picritic melts associated with continental flood basalts are derived from hot, relatively uncontaminated plume-source mantle at the plume axis and that the more voluminous tholeiitic basalts are produced by melting of cooler hybrid mantle in the plume head. This explains for why the picrites of the Karoo and Deccan are strongly enriched in highly incompatible elements, consistent with melting of an OIB-type source, whereas the associated basalts are weakly enriched in incompatible elements, consistent with derivation from a mixed OIB–lower-mantle source.

1. Introduction

Plate tectonic reconstructions place a number of major continental flood basaltic (CFB) provinces over known hotspots at the time of volcanism. Examples include the Deccan Traps, the British Arctic province, the Parana and the Karoo. These provinces overlay the Reunion, Iceland, Tristan de Cunha and Crozet hotspots respectively [1]. Furthermore, there is no evidence of volcanism above these hotspots prior to their association with the flood basalts [1,2]. It is also apparent, on the basis of the maximum MgO content of the basalts, that hotspots are zones of anomalously hot material rising from deep within the mantle [4]. Thus it has been suggested that CFBs are the initial volcanic product of new mantle plumes [1,2,5].

Mantle plumes must originate from a hot boundary layer deep within the mantle, possibly the thermal boundary layer above the core–mantle boundary, where the required heat flux can be drawn from the core. Plumes cannot be formed from localised anomalous concentrations of radioactive heating being carried about in the mantle since these will not generate temperatures high enough to provide the buoyancy needed to bring a plume to the top of the mantle [6].

A newly formed plume originating from an unstable boundary layer consists of two parts: a

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large bulbous head and a relatively narrow tail [7]. The head is large because the buoyant material escaping from the source region cannot displace the overlying mantle and ascend rapidly until a large amount of buoyancy has collected. The tail, or feeder conduit, is narrow because the hot, low-viscosity material flowing up an existing pathway can rise much faster than the head despite the small diameter of the conduit. We call the overall structure a "starting plume". Previous attempts to establish a link between CFBs and starting plumes [5,8] assume plume dynamics based on laboratory experiments in which the ascent of the plume is driven by compositional (hence effectively non-diffusing) buoyancy. Discussions of flood basalts have not considered the structure of plumes driven by thermal buoyancy. We will argue that this structure has profound effects on the dynamics of the plume and on the petrology and geochemistry of the resulting magmas.

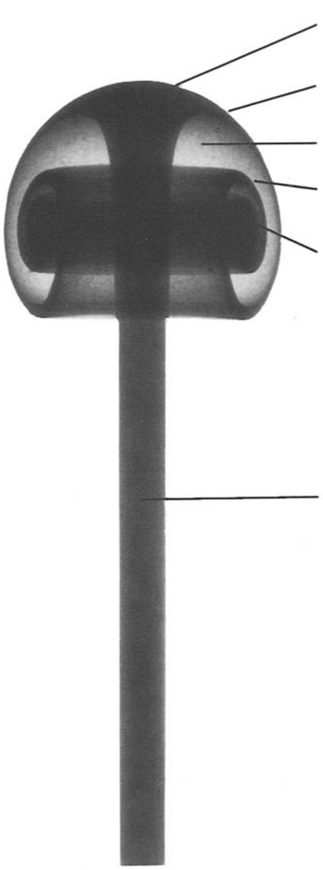

Fig. 1. A photograph of a laboratory starting plume [45] showing the structure caused by conduction of heat and consequent entrainment of surrounding fluid. Key features of the plume are labelled. Dark areas represent dyed material derived from the source, which becomes wrapped between layers of entrained mantle in a bulbous head. Hot source material is continuously supplied through the tail to the top of the plume head. The mean temperature in the head governs the dynamics of the plume ascent, and this temperature decreases as the volume of the head increases with time. The feeder conduit in these laboratory experiments is much broader, relative to the head, than is expected for mantle plumes because the viscosity of the hot conduit material in the experiments is only two orders of magnitude less than that of the surroundings, whereas the viscosity contrast in the mantle may be 10^{-2}–10^{-5}. Much narrower conduits are observed in experiments with compositional density and viscosity contrasts [7].

In laboratory experiments with plumes driven by buoyancy resulting from thermal expansion the conduction of heat leads to dramatic effects [3,6,9,10]. If there is a continuous supply of hot fluid from a source region a starting plume again consists of a large spheroidal head and a narrow feeder conduit. However, as the head rises, a thin boundary layer of adjacent mantle is heated by conduction until its temperature and density become comparable to that of the plume. This material is then a part of the plume and is stirred in with the source material by a recirculating flow within the head. The head is therefore a mixture of material both from the hot source and from the cooler mantle through which the head passes. Flow in the trailing conduit, on the other hand, at least for a time before it is tilted by larger scale shears in the upper mantle, does not entrain a significant amount of the adjacent mantle or cool as it ascends [11]. Its temperature remains appreciably higher than the average temperature in the plume head and its composition the same as that of the boundary layer from which the plume originates. The source mantle flows up the axial conduit until it reaches the top of the head, where it spreads radially outward over the cap and is wrapped into a spherical vortex (Fig. 1) [3]. As a consequence of the radial flow, the source material becomes a thin sheet with a large surface area which loses heat to the entrained material as the two are wrapped together. The head of the plume is then compositionally and thermally zoned.

The hottest part of the head, apart from the plume axis, is the thin layer of source material at the top of the head. Also warmer than the average temperature of the head is an expanding doughnut shaped mass containing the source mantle that formed the head of the starting plume early in its development (see Fig. 1). However, large temperature gradients are confined to the axial conduit and to the perimeter of the head, with gradients in the interior of the head smeared by conduction. In contrast, if the plume source and the overlying mantle have different compositions, compositional zoning will persist, on scales of order 10 km, throughout the head [3]. Modern hotspot basalts show enrichment in the light rare earth elements (LREE) and a distinct isotopic signature. Thus, if the source for starting plumes is assumed to be OIB-type mantle (that which forms the enriched source for ocean island basalts) the plume head should be zoned with layers of enriched OIB-type mantle separating regions of normal (entrained) mantle (Fig. 1).

Using a dynamic model for starting plumes Griffiths and Campbell [3] extend the results of laboratory experiments and predict that the heads of thermal plumes originating at the core–mantle boundary are likely to have diameters in the range from 800 to 1200 km by the time they reach the lithosphere. This size is insensitive to the buoyancy flux, but does depend on the lower-mantle viscosity. Because the final diameter of the plume head is comparable to the depth of the upper mantle, almost all of the entrainment takes place in the lower mantle. Source mantle will make up only 20–40% of the volume of the head, depending primarily on the volume flux from the source, the balance being entrained material from the lower mantle. Plumes originating from the transition zone at 650 km depth are predicted to develop heads with diameters less than 300 km.

Our model differs from that of White and McKenzie [8] since the latter, although it recognizes the essential link between CFBs and mantle plumes, does not make allowance for the highly time-dependent nature of starting plumes or the role of entrainment. We discuss the merits of the two models towards the end of the paper.

In this paper we compare the predictions of the starting plume model of Griffiths and Campbell [3] with the physical and chemical characteristics of Mesozoic/Cenozoic CFBs. The model provides a simple explanation for the sudden appearance of basaltic magmas over an equant area 2000–2500 km across followed by the rapid contraction of eruptions to a chain of volcanoes 100–300 km wide. We argue that flood basalts come from melting of the relatively cool regions of the plume head which are a mixture of source mantle and entrained lower-mantle material, whereas the associated picrites originate from the hot and relatively uncontaminated source mantle of the plume axis or tail. Calculations from the fluid dynamical model are consistent with estimates of the volumes and areal extents of volcanism and with the chronology of a sequence of uplift, followed by subsidence and volcanism.

2. Characteristics of continental flood basalts

We begin by reviewing the principal petrological and geochemical characteristics of the Deccan Traps of India and the Karoo of southern Africa. We have chosen these as examples of CFBs because they are well-documented. Both include picritic basalts with MgO contents in excess of 16% by weight, implying that they originate from an anomalously hot mantle source [4]. In addition, plate tectonic reconstructions place the volcanic province over a known hotspot at the time of volcanism [2], and the volcanics represent the oldest activity associated with the plume, suggesting that they are the product of a starting plume. White and McKenzie [8] give a detailed review of a number of other examples of continental flood volcanism.

2.1. Deccan Traps

Geomorphological studies have shown that the surface onto which the Deccan Trap volcanics were erupted underwent extrusive peneplaination, lateritization and uplift prior to volcanism. Uplift took the form of a gentle arch centred on the Pachmartic Hills south of the Narmada Valley, where the contact between the Traps and the underlying Gondwana Beds is at present approximately 1000 m higher than it is at the margins of the traps [12]. $^{40}Ar/^{39}Ar$ dating places the volcanism between 70 and 60 Ma. Duncan and Pyle [13] showed that the ages from the bottom and top of a 2 km thick composite section near Bombay were indistinguishable at 67.4 ± 0.7 Ma, suggesting rapid volcanism. This conclusion is supported by magnetostratigraphic studies that place most of the volcanism within a three chron (NRN) polarity sequence which can not have taken more than 3 Ma and probably took less than 1 Ma

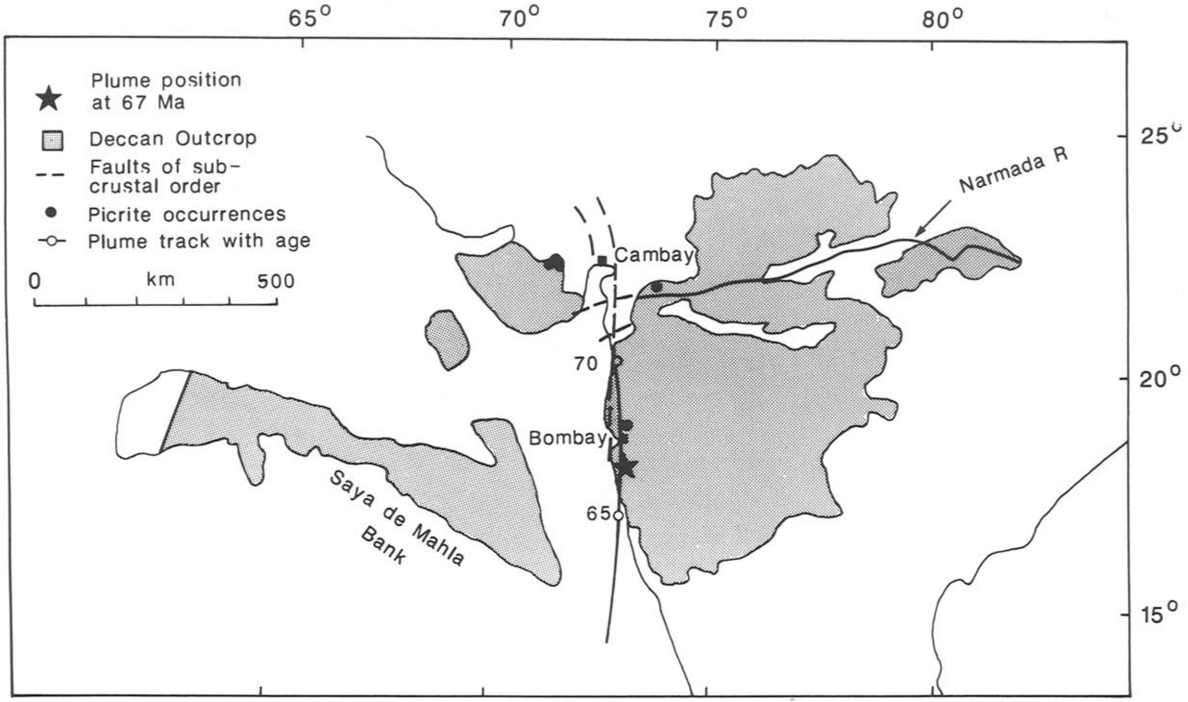

Fig. 2. A map of the Deccan Traps prior to the separation of the Saya da Malha Bank from India after [8] showing the extent of basaltic outcrop, known occurrences of picrite (from [21]), the track of the Reunion hotspot (from [47]), and the hotspot location at the time the Deccan Traps were forming.

[14]. Concomitant with volcanism two major grabens formed: the NNW trending Cambay Basin and the ENE trending Narmada Graben [12,15]. These grabens intersect a little to the south of the Bay of Cambay (Fig. 2). They contain up to 2.4 km of basalt and 6 km of terrigenous sediments [15]. The area covered by the Deccan, including the Saya de Malha Bank, is approximately 2000 × 1000 km and the volume of the extruded magmas is greater than 1.5×10^6 km^3 [5].

The most detailed studies of the geochemistry of the Deccan basalts have been made in the Western Ghats area near Bombay. Here eleven formations have been recognised [16–19]. The overwhelming majority of the flows are tholeiitic in character, but picritic basalts are found at a number of levels, especially near the base of the sequence [16,18]. The sequence generally dips southward at less than 1^0, with younger formations overstepping older ones. Watts and Cox [20] have interpreted this relationship as being due to the flexural response of the continental lithosphere as the volcanic load migrates southwards.

High-Mg picrites have been found low in the stratigraphic succession at a limited number of localities, mostly near the Bay of Cambay, close to the line traced out by the plume axis (Fig. 2). These rocks have crystallized from magmas containing at least 16% MgO [21] but are a volumetrically minor component of the Traps as currently exposed. Detailed geochemical data are available for some of the high-Mg picrites [21] and for the upper five formations [17,18] and preliminary data for the lower six [16]. These data, with the noteable exception of the Bushe Formation, show a general upward decrease in $^{87}Sr/^{86}Sr$ [16–18] and in the ratio of highly incompatible to weakly incompatible elements (e.g. Ba/Y).

Trace element abundance patterns for the upper five formations in the Bombay area and an example of a picrite are plotted on Fig. 3. The lowermost of these formations (the Bushe and Poladpur) have pronounced and weak negative Nb anomalies, respectively. This, together with the isotopic evidence, is taken by Cox and Hawkesworth [18] as evidence of crustal contamination.

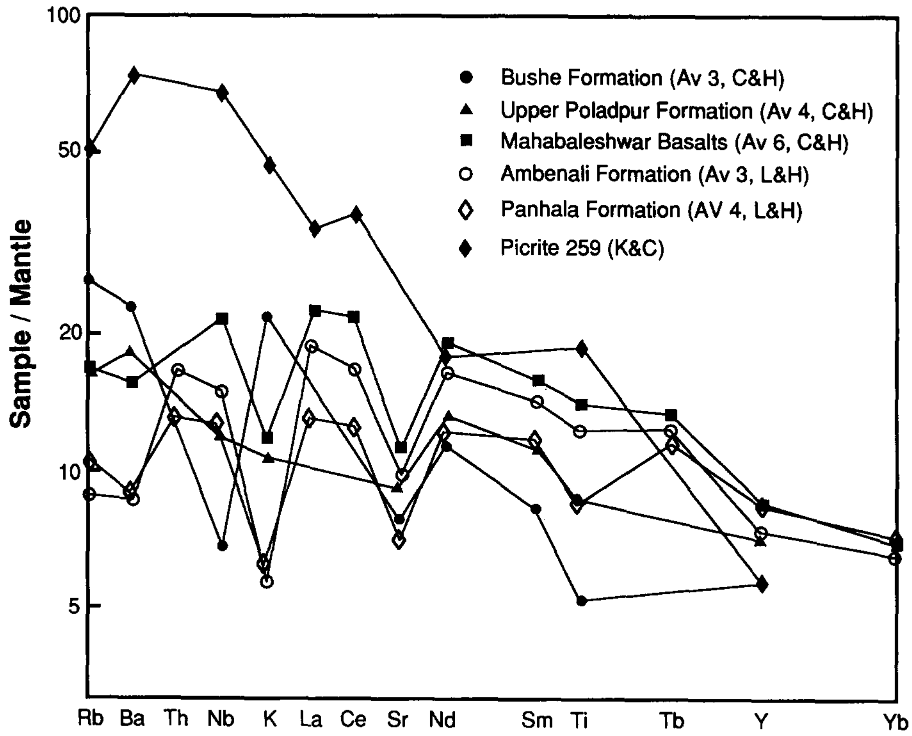

Fig. 3. Mantle normalized trace element abundance patterns for the Bushe, Upper Poladpur, Mahabaleshwar, Ambenali and Panhala Formations and a picrite from the Deccan Traps Mantle normalizing values from Sun and McDonough [48]; data from [17,18,21].

However, the upper formations (Ambenali, Mahabaleshwar and Panhala) and the picrite do not show Nb anomalies and are believed to have undergone little or no crustal contamination. Normalised abundance patterns for these formations show enrichment in highly incompatable trace elements. Enrichment in the picrite is particularly strong and approaches that seen in ocean island alkali basalts. The upper formations are isotopically and geochemically distinct from each other, suggesting that they have come from slightly different mantle sources [17]. On a plot of Sr/Nd isotopes samples from the uncontaminated formation plot subparallel to the mantle array in a similar position to the Kerguelen ocean island basalts [18].

2.2. The Karoo

The Karoo volcanics occur as scattered outliers representing what was originally a much more extensive volcanic province (Fig. 4). Outcrops extend from Lesotho in the south to Zimbabwe in the north and from Mariental in the west to Mozambique in the east. The preserved outcrops cover only 140,000 km^2 [22] but these outcrops are scattered and lie in a region approximately 2200 km long $\times$ 1800 km wide, suggesting that the original area covered by the volcanics may have been as great as 3,000,000 km^2. The volume of the basaltic sequence is difficult to estimate because the top has been removed by erosion in most areas. It is also unclear how much of the province was originally covered by basalt. One possibility is that most of the 3,000,000 km^2 was covered and that much of the erupted material has subsequently been removed by erosion. Alternatively the Karoo plume may have been selective in its ability to penetrate the lithosphere, with eruptions largely confined to the observed areas of outcrop. Richards et al. [5] estimate the volume of the Karoo basalts as 2×10^6 km^3.

Most dates for the Karoo basalts lie between 190 and 193 Ma, whereas the Jozini Formation (the main rhyolite unit of the Lebombo area) gives an age between 176 and 179 Ma [23]. There is also a suite of later dolerite dykes with an age of 165 Ma.

The best exposures of Karoo volcanics are found in the Lebombo area where a 6–9 km thick volcanic sequence is draped over a monocline that formed at the same time as the basalts [22]. This area is important in that it provides a complete cross-section through a sequence of flood basalts close to the line of retreat of the plume axis. The volcanic stratigraphy consists of a thin basal sequence of nephelinites overlain by picritic basalts, followed by basalts including basic andesites and rare shoshinites, and finally a thick rhyolitic unit [22]. In Mozambique there is a return to basic volcanism with a succession of interbedded basaltic and rhyolitic flows.

Normalised trace element abundance patterns for a typical nephelinite, picrite and basalt from the northern Lebombo are shown in Fig. 5. The nephelinite shows strong enrichment in the highly incompatible elements and is indistinguishable from ocean island nephelinites. The picritic pattern is similar to the nephelinite patterns but, unlike the nephelinites, has a pronounced Nb anomaly. The basalt also has a Nb anomaly and comparable highly incompatible element concentrations to the picrite but a distinctly lower La/Yb ratio. The basalts of the southern Lebombo area show less enrichment in incompatible elements than their northern counterparts. They have similar heavy REE concentrations but only half the average concentration of light REE. Lavas with low K_2O and other incompatible element contents are also found in the northern Lebombo but are invariably high in the succession [24]. Here the basalts above the rhyolites include examples that are amongst the most primitive found in the Karoo. The most depleted magmas of the Lebombo area are the Rooi Rand dykes which intrude the hinge zone of the Lebombo monocline. These dykes have MORB-like geochemistry with depleted LREE and low initial $^{87}Sr/^{86}Sr$ [22].

Basalts from other parts of the Karoo are di-

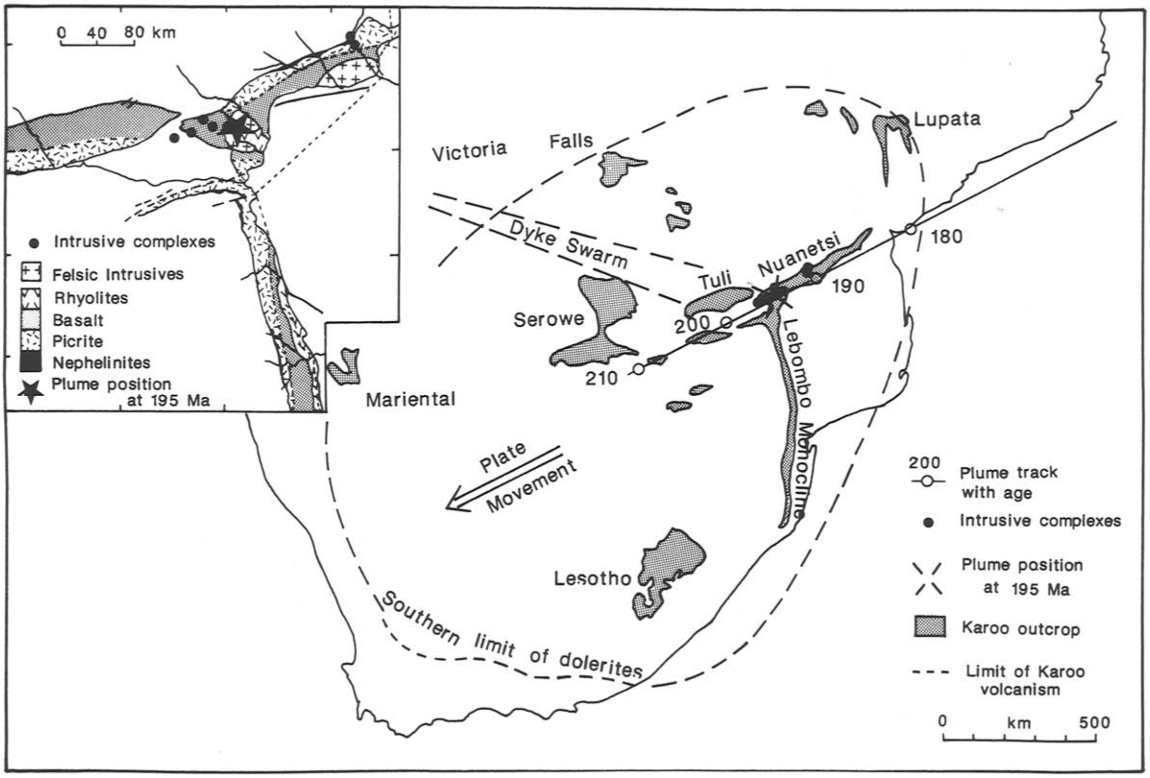

Fig. 4. A map of the Karoo basaltic province showing outcrop distribution, the path of the Crozet hotspot at the time the Karoo basalts were forming, and our estimate of the original extent of Karoo volcanism. The insert shows the distribution of picrites and nephelinites in the northern Lebombo region. They occur close to the line of retreat of the hotspot. Data from Eales et al. [49] and Bristow [50]. Plume position from Morgan [2].

vided into a low-Ti type which shows weak enrichment in highly incompatible elements and a high-Ti type which is strongly enriched in these elements. The low-Ti basalts are dominant in the south although high-Ti basalts have been recorded near the base of the sequence in the Stromberg area. In the north high-Ti basalts predominate but there is a return of the low-Ti basalts in the extreme north.

Picrites are found only in the northern Lebombo area close to the hotspot track (Fig. 4). Those with MgO contents between 12 and 18% MgO consist mainly of glass and quenched olivine, whereas picrites with greater than 18% MgO contain olivine phenocrysts [24]. The most Mg-rich olivines have forsterite contents of Fo_{92}. These observations suggest that the highest temperature pricritic melts in the Lebombo area had MgO contents of at least 18%. All of the picrites within the Lebombo area are enriched in highly incompatible elements but the degree of enrichment is highly variable. $(Ce/Yb)_n$ varying between 2.5 and 25 Ba/Nb between 19 and 51 [25].

3. A comparison with model predictions

The Deccan and Karoo show a number of common features, many of which are also found in other examples of CFBs. In each case volcanism starts suddenly and extends over a large and roughly equidimensional area 2000–2500 km in width. It then contracts rapidly to a narrow linear volcanic chain, 100–300 km across, marking the later track of the hotspots. The eruption rates for CFBs are anomalously large: minimum estimates for the mean rate of eruption of basalt during the most active periods are 10 m^3 s^{-1} for the Deccan (taking a 4 Ma period) and 70 m^3 s^{-1} for the Karoo (over a 3 Ma period). These rates are 10–100 times greater than the continuing eruption

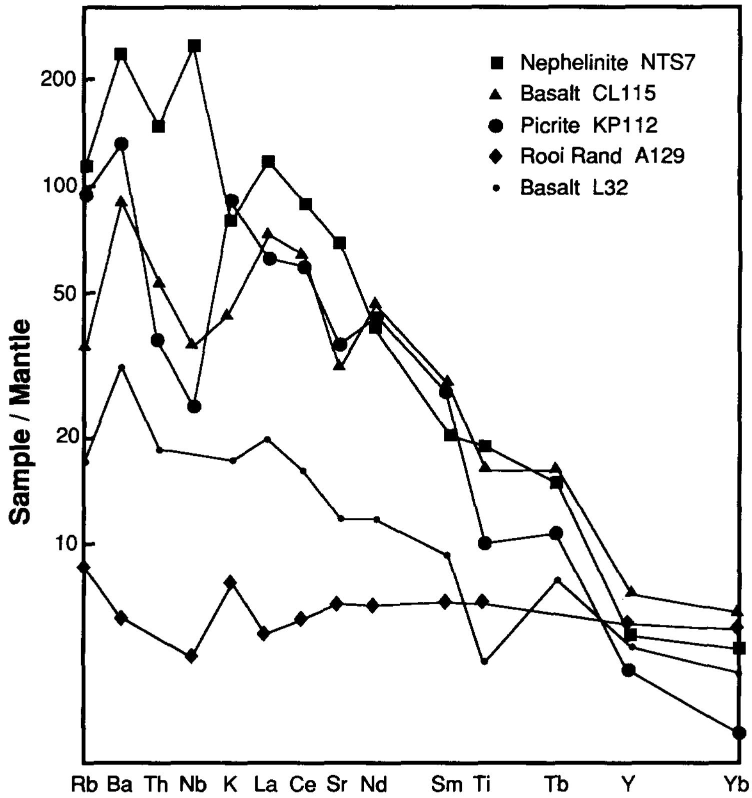

Fig. 5. Mantle normalized trace element abundance patterns for five Karoo samples: a nephelinite (NTS 7), a picrite (KP112) and a basalt (CL115) from the northern Lebombo; a basalt (L32) and a Rooi Rand dyke (A129) from the southern Lebombo. Data from Duncan et al. [51].

rates of 0.7–1.3 m^3 s^{-1} for the Reunion and Crozet hotspots, respectively [5]. By way of an additional comparison, even the Hawaiian hotspot track, the strongest of the modern plumes, represents a maximum eruption rate of only 3 m^3 s^{-1} [26]. For both the CFB provinces considered volcanism extended over a period of at least 7–15 Ma, although the bulk of the basalts were probably erupted during an initial short period of 1–4 Ma.

We note that in Figs. 2 and 4 the apparent centre of the volcanic province is displayed relative to the position of the hotspot at the time of volcanism and that the displacement is in the direction of continental drift. This is consistent with the plume head becoming effectively attached to the plate 10–15 Ma (see Figs. 2 and 4) before the onset of volcanism. Our laboratory observations indicate that at this time the top of the plume may have reached a depth less than 1/4 the diameter of the plume head (i.e. < 250 km).

Another feature common to the Deccan and Karoo is the occurrence of picrites. These are found only within an area close to the line of retreat of the hotspot (see Figs. 2 and 4). The picrites are LREE enriched relative to the basalts which are, in turn, enriched relative to MORBs. These and other characteristics are discussed briefly below, where they are compared with predictions based on the starting plume hypothesis.

3.1. The diameter of the plume head

The diameter of the mantle thermal anomaly required to form a CFB province can be estimated from the lateral extent of the volcanism and, more accurately, from the extent of the associated uplift. The width of major provinces, including the Deccan and Karoo, lies between 2000 and 2500 km [8]. This figure may be slightly greater than the width of the mantle thermal anomaly because it ignores the possibility that the lavas may flow beyond their point of eruption. Unfortunately, the lateral extent of individual flows is rarely recorded. In south eastern Australia, where flow lengths have been measured and where the basalts have similar compositions to the Deccan and Karoo, the largest flow is 60 km long (R. Price, pers. commun., 1989 [27]). The longest recorded flow of the Deccan is 150 km in length [19]. Further evidence for the mantle thermal anomaly extending close to the margins of the volcanic provinces is provided by the occurrence of dykes at the southern margin of the Karoo (see Fig. 4) and at the north eastern margin of the Deccan [19]. The basalts of the northeastern Deccan also contain abundant phenocrysts leading Mahoney [19] to conclude that they have not flowed far from their point of eruption. Finally, and perhaps most importantly, the area of volcanism corresponds closely with the area of uplift, providing direct evidence that the basalts were underlain by a mantle thermal anomaly of similar dimensions. We therefore concur with the conclusion of White and McKenzie [8] that the mantle thermal anomaly giving rise to CFBs had a diameter of between 2000 and 2500 km.

From the size of the mantle thermal anomaly we can estimate the size of the plume head required to produce it. A new plume must displace the overlying mantle in order to rise. As the top of the head approaches the surface, proximity to the surface and, to a lesser extent, the stiffness of the lithosphere slow the rate at which the head can displace the remaining overlying mantle. The result is horizontal spreading and flattening of the head. Experiments with buoyant spheres of liquid [28] show that flattening will become obvious once the top of the plume is within 0.5 head diameters from the surface, and that the diameter of the head will continue to increase by a factor of approximately two before the radial velocity decays significantly. Hence the plume diameter, before flattening, will be roughly one half the width of the flood basalt provinces, or ~ 1000 km for the Deccan and Karoo plumes. This result is consistent with the diameters predicted from a theoretical analysis of thermal plumes [3].

Calculations presented by Griffiths and Campbell [3] assume a lower mantle viscosity of 10^{22} Pa s (2.5×10^{18} m^2 s^{-1}) and a plume-to-mantle temperature anomaly near the core–mantle boundary of 200–800 °C. From the present day hotspot buoyancy fluxes, which are inferred from the seafloor topography [29,30], and a source temperature anomaly of 200–400 °C, we obtain source volume fluxes of 15–30 m^3 s^{-1} for the Crozet hotspot and 60–120 m^3 s^{-1} for Reunion. The head diameter at the top of the mantle is, however, predicted to be similar for these two flowrates and is 800–1200 km for plumes originating at the core–mantle boundary. Plumes originating at the upper mantle–lower mantle boundary have a calculated diameter of only 250–300 km [3,8] and can not produce a thermal anomaly with the dimension required to explain CFB provinces. The excellent agreement between the theoretical predictions and the observed size of CFB provinces supports the hypothesis that the strong mantle plumes responsible for CFBs originate at the core–mantle boundary [3]. The enormous size of plume heads also implies that the upper mantle will be displaced by, but not wrapped into, the head. *Only lower mantle and source material will be stirred into the plume.*

The volume of a plume head, with a diameter of 1000 km, is $\sim 5 \times 10^8$ km^3. When it flattens beneath the lithosphere it forms a disk $\sim$ 2000 km in diameter and $\sim$ 170 km thick. Only the mantle near the top of this disk can achieve the low pressures required for melting Tholeiitic basalts, of the type found in flood basalts, typically form within a pressure range of 1–1.5 Gpa [31], equivalent to depths of 40–55 km, suggesting that melting is confined to the top 10% or less of the volume of a plume that ascends beneath continental crust (although the figure may be greater for plumes ascending under new oceanic crust). This conclusion is consistent with the decrease in element ratios such as La/Yb and Ba/Nb in successive volcanic units, which we interpret as indicating that the earliest basalts in the Karoo and Deccan come from a mixed OIB-lower mantle source. The lower mantle component becomes increasingly important in the later flows. There is no evidence for a return to more enriched basalts in the youngest flows suggesting that the third (OIB-rich) layer in the plume head does not enter the melt zone. Our theoretical model for the stirring in thermal starting plumes [3] predicts that the total thickness of the top two layers, after the flattening of the head, will be of the order of 10 km, accounting for 5–10% of the volume of the head.

The degree of partial melting required to form flood basalts is probably in the range 10–15% [4]. If we assume that the percentage of melting is 15% and that melting is confined to the upper 10% of the plume, the volume of melt produced by a starting plume will be $\sim 10^7$ km^3. This figure is consistent with White and McKenzie's [8] estimate of the volume of magma produced by the Iceland starting plume but is greater than estimates for the volume of basalts in other CFB provinces. Some of this difference could be due to magma ponding in the crust during ascent and some (particularly in the Karoo) to a smaller degree of thinning of the lithosphere.

3.2. *Uplift and subsidence*

As a plume head rises beneath continental lithosphere the buoyancy it introduces into the upper mantle will lift the overlying mantle and crust in an approximate isostatic balance. Significant uplift of the earth's surface begins 10–20 Ma before the onset of volcanism when the top of the plume is one to two diameters below the surface and it reaches a maximum when the plume is 0.1–0.2 diameters from the surface [28]. For the estimated plume head diameter of order 1000 km and a temperature anomaly of 100 °C, the maximum rate of uplift at the plume axis is predicted to be 20–40 m/Ma. A maximum uplift of 500–1000 m should be attained when the plume is at a depth of 100–200 km. Thus, unless extensive lithosphere stretching has occurred as a result of other forces before the ascent of the plume [8], melting at depths of 40–55 km should not occur until uplift has ceased. This widespread uplift preceding plume-related volcanism places the lithosphere under tension and may cause stretching [32] similar to that required in the model of White and McKenzie [8].

Uplift is followed by subsidence. First, as the plume head spreads beneath the continental lithosphere the mantle buoyancy anomaly is dispersed over a larger area leading to subsidence above the plume axis, but to continued uplift near the margins of the plume, over a period of order 20–60 Ma [28]. Second, the removal of magma from the mantle deflates the head of the plume leading to enhanced subsidence of the overlying crust and mantle lithosphere. This process together with the weight of the volcanic material loaded onto the surface should cause the contact between the basalts and the surface they erupt upon to sink, although the top of the volcanic pile will rise slightly because the specific volume of basalt is about 10% greater than that of the mantle. Finally, decay of the mantle thermal anomaly by conductive heat loss to the surface causes extremely gradual, regional subsidence over a period of order 10^9 years. Subsidence can lead to the accumulation of sediments, particularly if the basin sinks below sea level.

Evidence for uplift preceding volcanism as predicted by the plume head model is sparse but this is not surprising because the early uplift is overprinted by later subsidence and it is the latest events that are most easily recognised. However, uplift prior to volcanism has been recorded in two flood basalt provinces: the Deccan Traps as already noted [12] and the traps of the Siberian Platform, where evidence for gentle arching prior to subsidence is reported [33].

The tension produced by a rising plume will lead to thinning of the crust and mantle lithosphere as both tend to slide off the raised area under the influence of gravity [32]. The best evidence for crustal thinning comes from the Siberian Platform where seismic studies show that the crust is about 30 km thick under the Tunguska basin compared with 35–40 km along the southern flank of the basin [34]. Crustal thinning is accompanied by deep crust faults that produce steps in the crustal thickness. Evidence for thinning of the mantle lithosphere, on the other hand, is indirect: experimental studies of the origin of basaltic magma suggest that tholeiitic magmas of the type found in continental flood basalt provinces form within a depth range of 40–55 km [31], requiring the top of the plume head to ascend to within 10–25 km of the base of the crust.

The evidence for subsidence is much clearer. Flood basalts occur in broad subsiding basins surrounded by uplands [35]. In the Karoo the basaltic sequence overlies aeolian sandstones and sediments deposited in lakes and river courses [22], implying gentle subsidence over a broad area prior to volcanism. In some areas subsidence leads to the development of grabens. Examples include the Cambay Basin and Narmada Graben of the Deccan Traps, and the Tuli and Nuanetsi synclines in the northern Lebombo area of the Karoo. Evidence from the Deccan suggests that enhanced subsidence coincides with the onset of volcanism [12,15] and we suggest that it results from withdrawal of magma from the plume head. This enhanced subsidence will be greatest above regions where the volume and degree of partial melting is greatest, which is likely to be near the axis of the plume. Thus the grabens of both the Karoo and Deccan are found, in close association with picrites, near the position of the plume axis at the time of volcanism (Figs. 2 and 4). The volcanics of the Tuli and Nuanetsi synclines are folded, suggesting that subsidence has continued after the main period of volcanism. This can be attributed to the preferential withdrawal of crustal melts from the high-pressure region below the thick volcanics in the synclines [36].

It is also interesting to note that the Cambay Basin (Graben) lies on the extension of the Reunion hotspot track when it is extrapolated beneath the Traps. A similar coincidence of graben, picrites and the hotspot track can be seen in the Karoo. The grabens, synclines and major faults in both provinces form a spoke-like pattern with an intersection which lies on the hotspot track (Figs. 2 and 4). There appear to be three radial lines in the Karoo and four lines forming a cross in the Deccan. We take this as evidence that the plume is active in inducing tension in the crust.

3.3. Melt production

Melting in the plume head occurs as a consequence of adiabatic decompression and takes place when the plume reaches the top of the upper mantle. In our model it will occur first in the high-temperature axis of the plume, which will begin melting at appreciably higher pressures than the cooler plume head. At these high pressures the density difference between the melt and mantle is small [37] and the melt will not acquire enough buoyancy to leave the mantle until the degree of partial melting becomes large. As a consequence, the first melts to leave the plume will normally be picrites derived from relatively uncontaminated hot source material at the plume axis (see Fig. 1).

Later the cooler "black" area at the top of the plume away from the axis in Fig. 1 and then the "white" area below may enter the melt zone and release lower temperature tholeiitic liquids. While the melt fraction is still small the tholeiitic melt will begin to ascend at a faster rate than the plume [38], eventually acquiring enough buoyancy to break away from the plume and ascend to the surface via a dyke. Details of this process need not concern us here. However, each batch of magma escaping from the mantle can be a collection of melt from a range of depths. A layered mantle may therefore act as a mixed source region for basalts. Thus, if the source mantle is OIB-type and the entrained material is depleted lower mantle, the first melts to be released from the cooler bulk of the head are likely to reflect a mixture of these two source regions with the contribution from the lower depleted mantle layer increasing with time. If melting extends to the second OIB-rich layer in the plume head, it is possible that the later magmas will again become more OIB in character.

Two factors may complicate this simple prediction. First, if sections of the plume source component of the mantle are rich in volatiles they will melt at high pressure to release a low-temperature melt rich in volatiles. Such a melt will have a low density and low viscosity and be able to escape from the mantle at very low degrees of partial melting. Small amounts of nephelinite that underlie the picrites in the northern Lebombo area may have formed in this way. Second, mantle may continue to flow up the tail after melting of the head has ceased. For this reason picritic magmatism need not be confined to the early stages of volcanism but may continue for some time along the track of the hotspot. On the other hand, tilting of the plume tail is expected to become more pronounced with time as the motion of the overlying plate drags the top of the plume farther from its source. The greater the inclination of the tail the greater will be the stirring and cooling due to entrainment of surrounding mantle into the tail [3,39]. Long lived plumes such as that under Hawaii are therefore less likely to yield high-Mg picrites (+16% MgO) than new plumes such as the one which formed the Karoo.

3.4. *The time scale for volcanism*

Basalt generation will continue as long as the thermal anomaly continues to rise, advecting more material into the melting zone. A maximum time scale for volcanic activity is therefore the time elapsed between when the top of the plume first reaches a high enough level in the upper mantle to allow significant melting and when the ascent is effectively arrested. More detailed analysis of the spreading of the plume head beneath the surface and the influence of the lithosphere rheology is needed in order to fully understand the production of melt. However, a first order estimate of the time can be obtained from the thermal plume model. Laboratory experiments with buoyant liquid spheres in a viscous fluid [28] imply that lateral spreading of the plume head beneath the surface in an upper mantle of viscosity 10^{21} Pas will continue for at least 20 Ma after the time of maximum uplift. The experiments also clearly show that vertical velocities decay very rapidly in time, implying that the volume of mantle entering the melt zone will be greatest during the early stages of volcanism and will subsequently decrease with time. The mechanism by which the plume penetrates the lithosphere remains uncertain but it is likely to involve small-scale convection. Hence the time that elapses between uplift and melting is uncertain. Thus the model predicts a burst of basaltic volcanism sometime after maximum uplift is reached, with most of the magmas erupted early but continuing at a decreasing rate, perhaps for a period of the order of 20 Ma.

These predictions are consistent with the timing of the volcanic activity for the Deccan, where most of the basalts were erupted in the first 1–3 Ma. Further less voluminous eruptions continued for at least 3–5 Ma. It is important to remember that the youngest volcanics in the Deccan and many other flood basalt provinces have been removed by erosion. The eruption rates for the Karoo are less clear but Bristow and Saggerson [22] also argue for an abrupt start to volcanism. As reported in Section 2, the bulk of these basalts were probably discharged within 3 Ma but activity trailed on for another 25 Ma. Other flood basalts show similar time scales. For example, in the British Arctic Province there was a burst of volcanism lasting less than 2 Ma, with smaller amounts of magma erupted during a further 5 Ma [40]. In the Columbia River, volcanism lasted for 11 Ma, but most of the material was erupted in the first 3 Ma [41].

3.5. *Geochemistry*

A full discussion of the geochemical consequences of our model is beyond the scope of this paper and we will therefore limit the discussion that follows to a brief summary of our main conclusions. A more detailed account will be presented elsewhere [42].

If flood basalts are to be interpreted in terms of a starting plume model, special emphasis should be placed on the geochemistry of the picrites since these should provide the least contaminated sample of the plume source. The picrites from the Karoo and Deccan Traps are enriched in highly incompatible elements. Hence we conclude that the Karoo and Deccan plumes originated from a reservoir of OIB-type mantle. The picrites of the central British Arctic Province [43] and the Siberian Traps are also enriched in incompatible elements and have apparently originated from a similar mantle source.

It has been suggested, with specific reference to the Karoo, that continental flood basalts are the product of fractional crystallization of a picritic parent magma [44]. However, the incompatible trace element contents of the picrites and basalts from the northern Lebombo region vary independently of the MgO content of the melt. This type of variation can not be due to differences in the degree of partial melting or to the amount of fractional crystallization required to produce these lavas, and is probably due to the mantle source being variably enriched in incompatible elements [24]. Furthermore, the available data [24] suggest that the average LREE content and La/Yb ratio for the picrites from the northern Lebombo is higher than for the associated high-Ti basalts and appreciably higher than for the low-Ti basalts from the southern half of the Karoo (Fig. 5). These differences are not readily explained by fractional crystallization of the basalts from the picrites unless the high-Ti basalts are derived from picritic magmas with a lower than average LREE content and La/Yb ratio, with the low-Ti basalts being derived from picrites with incompatible element concentrations towards the lower end of the observed range. Although we can not dismiss these possibilities we believe them unlikely.

A simpler explanation for the geochemistry of the basalts is that they form by melting the head of a starting plume. Because the plume head is cooler than the axis it will undergo a lower degree of partial melting, giving basalts with a lower MgO content than the picrites. Melting of a plume head also raises the possibility that the trace element geochemistry of the basalts may be controlled by melting two or more layers from a zoned source region. If one layer is OIB-type mantle, with trace element enrichment similar to that seen in the Karoo and Deccan picrites and the other is unenriched entrained lower mantle material, the mixing ratio required to produce the weak LREE enrichment of a typical low-Ti CFB is 1 part OIB-type mantle to 4 parts entrained mantle. This mixing ratio agrees well with the predicted ratio of the thicknesses of the upper source mantle (OIB-type) layer to a second layer of entrained mantle at the top of the plume head as calculated by Griffiths and Campbell [3]. As noted earlier, the CFBs of the Karoo and Deccan show a general upward decrease in the ratio of highly incompatible to weakly incompatible elements. We suggest that this is due to a systematic decrease in the contribution of the OIB-rich upper layer of the head to melts as the plume rises and flattens, bringing increasingly more of the second layer of entrained lower mantle material into the zone of melting. This type of trend is likely to be blurred if penetration of the lower lithosphere involves small-scale convection.

It has often been suggested that melting of the continental mantle lithosphere plays an important role in generation of CFBs. However the continental mantle lithosphere forms part of the cold boundary layer at the top of the mantle and it can only be melted by uplift and stretching if it gains sufficient heat by conduction from a hot plume [8]. Simple conductive calculations suggest that this is unlikely on the timescale of continental flood volcanism: if it is assumed that the top of a plume head is emplaced instantaneously to a depth of 50 km and then cooled by conduction, melting due to conduction is confined to a restricted zone above the central (hot) axis of the plume and no melting occurs in the mantle lithosphere over the remainder of the plume head. Since the instantaneous emplacement assumption used in this calculation leads to an overestimate of the amount of melting above the plume it is unlikely that the mantle lithosphere contributes significantly to CFBs. The lithosphere certainly can not achieve the high temperatures required to produce picrites.

4. Comparison with the White and McKenzie model

McKenzie [38] recognised that lithospheric stretching alone cannot explain CFBs and that anomalously hot mantle is also required. This led to the view that CFBs form at a continental margin when the opening of a new ocean basin coincides with a mushroom of hot mantle produced by an ascending plume [8]. In this model, the sudden onset of volcanism is attributed to the passive upwelling of hot mantle as the lithosphere stretches and thins in response to the opening of the ocean. White and McKenzie [8] also recognise that the head of a new plume must be involved in order to account for the coincidence between CFBs and the beginning of hotspot tracks. However, like Richards et al. [5], they suggest a head volume which is some 100 times smaller than that indicated by our calculations. The volume they suggest is only half their estimate of the volume of basalt erupted in the British Arctic Province and is less than 2% of the material required to form the large thermal "mushroom" spread beneath the lithosphere in their numerical model. The reason for this underestimation of the plume head volume is that they have neglected to include the effects of entrainment. White and McKenzie resorted to a steady-state model of a long-lived plume beneath a stationary lithosphere in order to explain the very large lateral scale of CFBs.

The diameter of the spreading hot mushroom required in the White and McKenzie model is 2000–2500 km. However, this large thermal anomaly can only be built up over a long period. The time required depends on the upwelling volume flux in the plume and the rate of drift of the overlying plate. If the plate is assumed to be stationary, and the volume flux is 200 m^3/s (the upper mantle flux inferred for strong modern plumes such as Iceland [30]), the time required to produce a mushroom 2000 km in diameter and 100 km deep is ~ 50 Ma. The steady-state plume model is therefore inadequate as an explanation for the voluminous but short-lived outpouring of basalt that characterises continental flood volcanism. The model requires an incubation time of at least 50 Ma for which there is no evidence. One might expect, for example, continental flood volcanism to be preceded by over 50 Ma of alkali volcanism. However, CFBs are invariably the first expression of volcanism from a new hotspot. Further, CFBs cover sub-equant areas ~ 2000 km across [8]. In the steady-state model this is only possible if the plate remains stationary above the hotspot for at least 50 Ma. If the plate moves at even a modest rate relative to the hotspot, the mantle thermal anomaly is appreciably less than 2000 km wide (e.g. 1200 km in the case of the Hawaiian plume) and the volcanic activity takes the form of a linear chain less than 200 km wide. At the time of the Karoo and Deccan volcanism, the African and Indian plates were moving at velocities ~ 30 mm/yr [2] and 150 to 300 mm/yr respectively [19].

Finally, we note that when a long-lived plume passes under a zone of extreme lithospheric extension, such as when the Reunion hotspot passed under the central Indian ridge, the amount of volcanic material produced is not unusually large. At the same time, many CFB provinces, including the Columbia River, the Siberian Traps and the main volcanic event of the Karoo (at 193 Ma), are not associated with the opening of a new ocean, although the British Arctic Province and Deccan clearly are. These observations suggest that, where the external forces that drive plate motions place the continental crust under strong tension, the uplift caused by arrival of a new plume may weaken the lithosphere, causing it to split and form a new ocean basin. This explains the apparent coincidence between new ocean basins and CFBs. However, where the tensional forces acting on the continent are weak, a plume head may induce the crust to rift without splitting. The amount of volcanism produced in this case will depend on the capacity of the plume to thin the lithosphere. Hence, although we agree that enhanced volcanic activity can be expected if the ascent of a large plume head coincides with lithosphere stretching during formation of a new ocean basin, by observation such a coincidence *is not* a prerequisite for flood volcanism. We suggest that the lithospheric thinning required for volcanism can result from the tension induced by the plume head itself.

5. Conclusions

The starting plume model for CFBs is not new but as previously stated was incomplete. Inclusion of effects of thermal entrainment in calculations is essential. The model then predicts the sudden onset of volcanism over an equidimensional area 2000–2500 km in width, its short duration, and its rapid contraction to a linear volcanic feature 100–300 km wide. Further, the plume structure induced by thermal entrainment provides a simple explanation for continental flood basalts and their associated picrites: the flood basalts are produced by melting the cool hybrid head of a starting plume, and the picrites result from melting the hot and relatively uncontaminated source mantle in the tail. As a consequence, the geochemistry of picrites can be used to establish the nature of the plume source. For the Karoo and Deccan the source is highly enriched OIB-type mantle.

We have suggested elsewhere [45] that Archean high-Mg komatiites were produced by melting in the high-temperature axis of a starting plume. However, komatiites have depleted geochemical and isotopic characteristics, implying that Archaean plumes originate from a MORB-type mantle source. Hence the nature of the source for mantle hotspots appear to have changed from depleted MORB-source mantle in the Archaean to enriched OIB-source mantle for modern hotspots. The origin of the OIB source is controversial but it is widely believed to represent a component of subducted oceanic slab [46]. If this is the case, OIB-type material may be absent from Archaean mantle plumes simply because the first convective cycle involving the formation, subduction and return of oceanic crust was not completed until after the end of the Archaean.

Our calculations show that the diameters of starting plume heads before their collapse beneath the lithosphere lie in the range 800–1200 km if they come from the core–mantle boundary. The introduction of such a large volume of anomalously warm mantle into the upper mantle elevates the lithosphere by about 1000 m and places it under tension. The significance of the tension, induced by the plume head, is that it can lead to rifting of the lithosphere without splitting the continent and producing new oceanic crust. As Richards et al. [5] point out, the mid-continental rifting associated with flood basalts is the consequence of plume initiation, not its cause. At the same time, plumes are not the fundamental force that drives the motion of plates and rifts associated with flood basalts are not failed attempts at continental drift.

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