purú was then calculated from combining this regression with the rating curve (*10*) to obtain the flux in cubic meters per second $Q = 30{,}683 - 9.72\ S_{mc} + .016\ S_{mc}^2$, which was reported as monthly mean discharge.
12. A deseasonalized hydrograph was calculated by subtracting the 83-year mean discharge for each month (Q^*) from the respective monthly mean discharge in each year (Q) for the ith month of the jth year to form the deviation from the long-term monthly mean discharge $Q'_{i,j} = Q_{i,j} - Q_i^*$.
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Flood Basalts and Hot-Spot Tracks: Plume Heads and Tails

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Continental flood basalt eruptions have resulted in sudden and massive accumulations of basaltic lavas in excess of any contemporary volcanic processes. The largest flood basalt events mark the earliest volcanic activity of many major hot spots, which are thought to result from deep mantle plumes. The relative volumes of melt and eruption rates of flood basalts and hot spots as well as their temporal and spatial relations can be explained by a model of mantle plume initiation: Flood basalts represent plume "heads" and hot spots represent continuing magmatism associated with the remaining plume conduit or "tail." Continental rifting is not required, although it commonly follows flood basalt volcanism, and flood basalt provinces may occur as a natural consequence of the initiation of hot-spot activity in ocean basins as well as on continents.

CONTINENTAL FLOOD BASALTS OCCUR worldwide (Fig. 1), and because of their wide age distribution and enormous volume, they are of fundamental importance in the recent evolution of the continental crust and lithosphere. However, the basic cause (or causes) of flood basalt eruptions has remained a mystery. In developing his theory that fixed hot spots (such as Hawaii and Iceland) were formed from convective mantle plumes, Morgan (*1, 2*) noted that hot spots were associated with flood basalts and suggested that flood basalts occurred at the initiation of mantle plume activity. Indeed, laboratory experiments (*3*) have shown that a large initial diapir or "head" is necessary in order to establish a mantle plume conduit (Fig. 2). Recently, several workers have suggested that the eruption of the Deccan flood basalts in India marked the onset of activity of the Reunion hot spot (*4–6*). In this report we show that many continental flood basalt events and, perhaps, large suboceanic plateaus could be the result of mantle plume initiation. Further, we show that the plume initiation model provides a plausible explanation for the enormous volcanic eruption rates for flood basalts compared to those for the associated hot-spot tracks.

Many flood basalt provinces occur along continental margins and were associated with initiation and early development of continental rifting (*7–12*). The best preserved provinces are predominantly Mesozoic and younger in age. The largest of these events (Deccan, North Atlantic, Parana, and Karoo) were coincident in space and time with the earliest activity of major hot spots (Reunion, Iceland, Tristan da Cunha, and Marion, Fig. 1). If these flood basalts and hot-spot tracks resulted from a common mantle-plume formation mechanism, there should be a fairly consistent relation between the size of the initial plume head or diapir (represented by flood basalt volume) and the rate of flow through the remaining deep-mantle conduit (represented by hot spot–track volcanism). To test this idea, we have estimated volumes and eruption rates (Table 1) for several well-studied flood basalt provinces and associated hot spots.

The Deccan Traps, a thick sequence of flat-lying basalt flows covering nearly 500,000 km^2 of west-central India (*10*), is associated with the Reunion hot spot. At its greatest thickness (in the Western Ghats) the lava flow succession is over 2000 m (*12*). When correlative basalts identified offshore (Arabian Sea) are included and an estimate is made for eroded lavas, the province may have covered more than 1.5×10^6 km^2 (*5, 13, 14*). Several lines of evidence indicate that most of the basalt was erupted rapidly. Interflow sedimentary beds, evidence of extensive weathering, and erosional unconformities are scarce in much of the province (*10*). The flows are predominantly reversely magnetized and only two polarity reversals (normal to reverse to normal) have been identified in the entire sequence (*5*). Recent ^{40}Ar-^{39}Ar incremental heating ages show that volcanism occurred between 65 and 69 million years ago (Ma) (*15*) and that the 2000-m-thick Western Ghats section was erupted in less than 2 million years (*16*). In consideration of all evidence for the duration of volcanism, most of the Deccan basalts may have accumulated in as little as 0.5 million years (*17*). The average eruption rate would then have been greater than 1 km^3 per year; more likely, episodes of more rapid eruption were separated by inactive periods (*14*). A series of southward-younging submarine volcanic lineaments (the Laccadive, Chagos, and Mascarene ridges) link the Deccan flood basalts to the Reunion hot

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spot. These lineaments record the northward path of the Indian plate away from the hot spot through Tertiary time. Reunion Island, the present site of hot-spot activity, has been built in the past ~2 million years (*18*) and has a volume of about 75,000 km^3. The present eruption rate is thus close to 0.04 km^3 per year. This rate appears roughly consistent with activity along most of the hot-spot track.

In the North Atlantic Tertiary Province basaltic rocks crop out along correlative sections of the continental margins of the North Atlantic (eastern Greenland and northwestern Europe), opposite the Iceland hot spot. Igneous activity in the province occurred during a short period, probably less than 2 million years (*11, 19*), at about 60 Ma and preceded the onset of seafloor spreading in the North Atlantic (*19*). The estimated total volume of extrusive rocks is $>2 \times 10^6$ km^3 (*20*), which implies an average eruption rate of >1.0 km^3 per year. These flood basalts are linked to the current hot spot through the paired Greenland-Iceland and Iceland-Faeroe ridges. The output from the hot spot (0.023 km^3 per year) is estimated from the volume of basalt forming the island of Iceland (1.0×10^6 $km^2 \times 3.5$ km) that is in excess of the production of the mid-Atlantic spreading ridge and the 15-million-year age range of exposed rocks (*21*). This flux is slightly less than an average eruption rate (0.038 km^3 per year) calculated from direct observation during the last 11 centuries (*22*); both estimates are similar to the output of the Reunion hot spot.

The Parana (Serral Geral) basalts in eastern central South America (Brazil, Paraguay, Uruguay, and Argentina) and the Etendeka basalts of coastal southwest Africa (Namibia) were once joined in a large flood basalt province erupted just before continental breakup in the South Atlantic (*2, 19, 23*). Seafloor spreading has separated the two outcrop areas leaving volcanic trails on the African plate (the Walvis ridge) and the South American plate (the Rio Grande rise) that become younger toward the active Tristan da Cunha hot spot (*23*). The present outcrop of the Parana basalts is 1.2×10^6 km^2, but abundant dikes and outlying basalts suggest that the original area covered was about 2.0×10^6 km^2 (*7*). The present average thickness is 650 m, thus the likely volume is 1.3×10^6 km^3, uncorrected for erosion. The volume of the smaller Etendeka basalt is $\sim 2.5 \times 10^5$ km^3, which yields an approximate original volume for the total province in excess of $\sim 1.5 \times 10^6$ km^3. Radiometric ages (K-Ar) for the Parana basalts show wide scatter but most are between 120 and 130 Ma (*24*). The Etendeka basalts have also been dated (K-Ar) at 120 to 130 Ma (*25*). Paleomagnetic measurements of the Parana basalts have revealed that there are only a few reversals in the section (*9*), which, during this period of the magnetostratigraphic time scale, implies that the duration of volcanism was only a few million years. Hence, average minimum eruption rates for this province may be 0.5 to 1.0 km^3 per year, comparable with that for the Deccan flood basalts. The island of Tristan da Cunha has formed in the last 0.2 million years (*26*) and has a volume of about 6000 km^3; thus, the estimated present eruption rate over the hot spot is 0.03 km per year, similar to that for the Reunion hot spot.

The Karoo basaltic province includes a wide variety of basalt types related to several volcanic episodes of early Jurassic to early Cretaceous age distributed unevenly across southern Africa (*27*). We consider only the early Jurassic phase, which has been related to the Marion–Prince Edward hot spot (*2*). Thick sequences of these basalt flows were erupted throughout the stable continental Kaapvaal craton. Because of discontinuous exposures and variable volcanic stratigraphies (*28*), correlation of units between the major outcrop areas has been difficult. Volumes are also more uncertain than in the younger provinces. The original area covered by Karoo flows may have been as much as 2.0×10^6 km^2 (*9*), and thickness may have averaged 1 km (*7*); these values give an original volume of about 2.0×10^6 km^3. Age relations are also less well known, but the majority of the eruptions of the Early Jurassic phase of volcanism occurred about 195 ± 5 Ma (*29–32*). At present the duration of volcanism is not well constrained. The Karoo basalts of eastern South Africa are similar to the Deccan and Parana-Etendeka basalts in eruptive style (especially in the Lesotho area), and thus they, too, may have been erupted in a few million years at an average eruption rate of ~1 km^3 per year. Cretaceous volcanism in Madagascar and Tertiary volcanism along the Madagascar ridge mark the path of the African plate over the hot spot now beneath Marion and Prince Edward islands. The present hot-spot eruption rate, about 0.02 km^3 per year, is estimated from the volume (10,000 km^3) and age [0.5 Ma (*33*)] of Marion and Prince Edward islands.

The four major flood basalt provinces considered above are the earliest manifestations of long-lived hot spots. Estimates for the initial flood basalt volumes, duration of volcanism, and eruption rates (Table 1) are all similar, as are eruption rates for the present hot spots. There is considerable uncertainty in the flood basalt volume estimates, especially for the two older provinces. However, the apparent similarities among the different provinces suggests that there is a common mechanism for rapidly generating enormous quantities of flood basalts from mantle plumes. For example, during the Deccan event, at least 1.5×10^6 km^3 of basalt was erupted in perhaps as little as 0.5 million years (*14*). If this basalt resulted from a cumulative 20% partial melting of mantle peridotite (*34*), a minimum parent mantle diapir volume of $\sim 7.5 \times 10^6$ km^3 or a spherical blob (plume head or cavity plume; Fig. 2) at least 240 km in diameter would be required. The associated Reunion hot-spot track was formed with a much slower eruptive rate of about 0.04 km^3 per year (compared with 1.5 km^3 per year for the Deccan Traps). We ascribe this smaller,

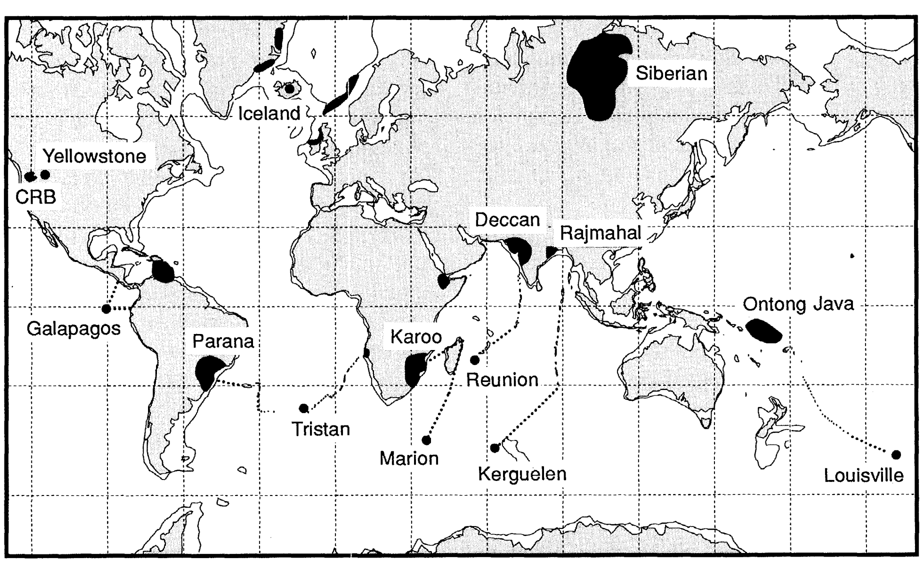

Fig. 1. Distribution of flood basalt provinces erupted in the last 250 million years and associated active hot spots and volcanic traces; CRB, Columbia River basalts.

Table 1. Estimated age, original volume, duration of volcanism, and eruption rate for flood basalt provinces and associated hot spots (marked with an asterisk). All hot spots are currently active. It is difficult to assign proper error estimates to the flood basalt volumes. As much as a factor of 2 uncertainty is possible, although other authors have given similar volume estimates (*10, 19*); m.y., million years. Reported analytical uncertainties are given for ages.

Province and age (Ma) or hot spot	Original volume (km^3)	Duration (m.y.)	Rate (km^3 per year)
Deccan (67 ± 1)	>1,500,000	<1.0	>1.5
*Reunion	75,000	2.0	0.04
North Atlantic Tertiary (60 ± 1)	>2,000,000	2.0	>1.0
*Iceland	3,500,000	15	0.02
Parana-Etendeka (125 ± 5)	1,500,000	2.0 ?	~0.75
*Tristan da Cunha	6,000	0.2	0.03
Karoo (early phase) (195 ± 5)	2,000,000	2.0 ?	~1.0
*Marion–Prince Edward (Crozet)	10,000	0.5	0.02

more steady hot-spot source to the remaining feeder conduit trailing the initial cavity plume.

As a model for the initiation of a mantle plume associated with a hot spot, consider a low viscosity plume being fed at a constant volume rate of flow R at the base of a fluid layer of depth d and viscosity η. As shown by Whitehead and Luther (*3*) a large plume head is formed, and the rise of buoyant plume material is controlled to a good approximation by the (instantaneous) Stokes rise velocity of the plume head. A thin conduit connects the plume with its source. The key assumption is that the plume has much lower viscosity than the surrounding fluid, and this situation gives rise to the configuration in Fig. 2. Such a viscosity contrast is expected for mantle plume material that is perhaps several hundred degrees hotter than normal mantle (*35*). For simplicity, we also assume that the plume buoyancy is nondiffusive—either the relevant time scales are too short for significant conduction of heat out of the plume, or the plume is chemically buoyant (for example, because of higher volatile content). Thermal diffusion and resulting thermal entrainment of surrounding fluid would cause enlargement of the plume (*36*) but would not qualitatively alter our results.

If the plume source is turned on at time $t = 0$, then the final volume V of the plume head is given by

$$V(t) = Rt \qquad (1)$$

where

$$t = (5\eta d/\Delta\rho g)^{3/5}\,(4\pi/3R)^{2/5} \qquad (2)$$

is the time required (*3*) for a plume of density contrast $\Delta\rho$ to rise through the fluid (viscosity η and depth d); g is the gravitational acceleration. The slowdown of the plume as it nears the free surface has been ignored. In our calculations, we used the more accurate, experimentally determined relation (*31*)

$$t = (d/0.16)^{5/7}(\eta/g\Delta\rho)^{4/7}R^{-3/7} \qquad (3)$$

although the difference is not significant given the uncertainties in the values of the parameters.

As an example, we adopt the reasonable values $\Delta\rho = 0.1$ g/cm^3, $d = 3000$ km (mantle depth), and $\eta = 10^{22}$ Poise (average mantle viscosity). Also, we use the observed eruptive rate along the Reunion hot spot track as a constraint on R: If 10% partial melting of the mantle plume produces primitive Reunion magmas (*34*), $R = 0.04/0.10 = 0.4$ km^3 per year. Using Eq. 3, we obtain $t = 28$ million years for the cavity plume rise time. The final volume of the plume head is 11.2×10^6 km^3 (so $V \times 0.20 = 2.2 \times 10^6$ km^3 is the total flood basalt volume for 20% melting), and the final diameter $a(t)$ as it encounters the upper surface is given by

$$a(t) = 2(3V/4\pi)^{1/3} = 280 \text{ km}$$

The approximate match with our estimate for the diameter of the Deccan blob of 240 km is a result of our particular choices for mantle viscosity, melt fraction, and plume density contrast. This calculation is obviously not robust, but it does demonstrate that the simple model is compatible with the observed volume and rate estimates for the Deccan Traps and Reunion hot spot. Most likely, a large part of the plume head is formed as the initial boundary layer instability detaches so that only a fraction of the head is actually fed through the trailing conduit. This would imply a much shorter rise time for the diapir, perhaps ~15 million years (*6*). Also, the mantle is certainly more complicated than a simple, passive fluid layer with uniform viscosity. However, the comparisons above show that the general concept is plausible: A feeder conduit flow rate sufficient to produce a hot spot is capable of producing an initial plume head sufficient to cause a flood basalt event.

The relation between final plume head volume V and conduit feed rate R from Eqs. 1 and 3 is $V \approx R^{4/7}$. Thus, the flood basalt event size should be a fairly weak function of feed rate. Also, a greater contribution to the plume head from the initial boundary layer instability will substantially reduce the total rise time through the mantle but not the final volume. Therefore we conclude that, for a given mantle viscosity structure, there will be a roughly characteristic plume head size (in the above case ~280-km diameter) required to traverse the mantle to the earth's surface and hence, similar total flood basalt volumes.

The important observations are that (i) the eruptive rates for flood basalt events are enormous (at least 20 to 50 times as large) compared with those of hot-spot tracks and

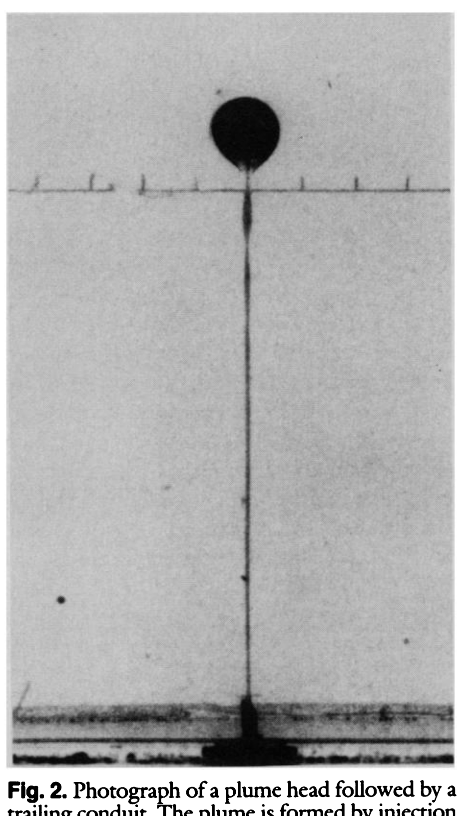

Fig. 2. Photograph of a plume head followed by a trailing conduit. The plume is formed by injection of water (colored with red dye) at a constant rate into the bottom of a large tank of glucose syrup. The water is buoyant and has a viscosity about four orders of magnitude less than that of the glucose. The horizontal scale just below the plume head has 1-cm tick marks. The photograph was taken 58 s after the beginning of injection. Similar experiments (applied in another context) are shown by Olson and Nam (*53*).

that (ii) the total flood basalt volumes and associated hot-spot eruptive rates are similar for the four events described above. Any detailed model must account for these basic relations. We have not attempted to explain how the initial plume instability forms or why the conduit might be maintained—various scenarios have been developed by others from models of core-mantle boundary instability (*6, 38, 39*). Observationally, we maintain that such instabilities are formed (most probably at the core-mantle boundary) and that the conduits remain for long periods of time (~100 million years, as implied by the persistence of hot spots).

In addition to the flood basalts we have discussed in detail, other continental flood basalt provinces have been linked to currently active hot spots (Fig. 1): the Columbia River basalts (~15 Ma) to Yellowstone (*1*), the Ethiopian basalts to Afar (*2*), and the Rajamahal traps to Kerguelen (*40*). Morgan (*2*) speculated that the Siberian traps may mark the beginning of the Jan Mayen hot spot, although the intervening hot-spot track has not been identified. Large, submerged oceanic plateaus may also be largely volcanic, of the flood basalt type, and formed at the initiation of hot-spot activity. For example, the Ontong Java Plateau, which formed ~120 Ma (*41*) and has an area of $\sim 0.75 \times 10^6$ km^2, might be associated with the initiation of the hot spot that produced the Louisville Ridge (*42*). Much of the anomalously shallow Caribbean plate (area 0.5×10^6 km^2) is thought to be a 1- to 2-km-thick accumulation of basalt of age ~80 to 90 Ma (*43*) and has been linked to the initiation of the Galápagos hot spot (*44*). Deep ocean drilling and sampling is needed to define better the age, composition, and extent of these oceanic plateaus so as to verify their relation with hot-spot tracks.

An alternative mechanism for rapid melt generation in flood basalt provinces that has been proposed recently is decompression of abnormally hot asthenosphere as a passive response to continental stretching and rifting (*11, 19*). This model differs from ours in that no plume initiation event is involved and large amounts of extension and rifting of continental lithosphere must precede the flood basalt eruptions. This important difference in the predicted location, timing, and tectonic style for flood basalt events (active versus passive rifting) suggests a number of tests for the two models.

Several lines of geological evidence favor the plume initiation model. (i) Some flood basalts such as the Siberian traps and Columbia River basalts are not associated with rifting events. (ii) Dike swarms that fed the Deccan and Karoo eruptions are randomly oriented and are distributed over broad cratonic areas (*32, 45*); this distribution is evidence against a purely extensional mechanism. (iii) Where rifting occurred, it was preceded by large flood basalt eruptions. The ~195-Ma Karoo eruptions predate rifting by at least several tens of millions of years, although later, less voluminous flows at about 178 Ma may have occurred just before rifting of Antarctica and Africa (*27, 32*). The Deccan eruptions preceded an intraoceanic ridge jump (and minor rifting event) that carried the Seychelles bank southwestward from the Indian subcontinent; the main tholeiitic eruptions of the Deccan preceded significant east-west extension (*45*). The Parana-Etendeka eruptions occurred before the opening of the South Atlantic Ocean in that area (*46*), and archtype structures (and associated feeder dikes) are distributed over a broad area of continental lithosphere. (iv) If the Caribbean and Ontong-Java plateaus are flood basalt events, they would not fit the passive rifting model because there was no associated continental rifting. (v) Similar flood basalt volumes (Table 1) are expected for different plume initiation events but not necessarily for rifting events where melting is linked to the extent of upper mantle heating. In particular, we note that India was moving northward rapidly (~15 cm per year) at the time of the Deccan eruptions. This motion suggests that there was not any long-term heating of the asthenosphere beneath India by the Reunion plume and that there must have been a large plume initiation event.

Our model suggests that in the cases where continental rifting follows flood basalt volcanism, both are the result of a large, hot diapir. Processes of melt generation and of lithospheric thinning and extension are easily explained in this context. An excess plume temperature of ~250° to 300°C would generate 10 to 20% melt at the base of the lithosphere (*47*), especially if coupled with probable higher volatile contents (CO_2, H_2O) of lower mantle material (*48*). Even small melt fractions at the base of the lithosphere may accelerate heating, thinning, and, perhaps, destabilization of the lithosphere and result in further melting as hot plume material intrudes to shallower levels. If rifting and extension do occur in response to the buoyant plume head, decompression of plume material will promote further melting. The plume head will spread out beneath the lithosphere rapidly (compared to total rise time through the mantle) because the viscosity of the uppermost mantle is probably significantly less than that of the deep mantle. (For a reasonable upper mantle viscosity of 10^{21} Poise, the Stokes rise velocity of our Deccan blob is about 1.5 m per year.) Rapid spreading might account for the geologically brief (<1-million-year) duration of the Deccan eruptions, for example. Because many flood basalt events (Deccan, North Atlantic, Parana, Siberia) were preceded by thermal doming events (*19, 46, 49, 50*), analysis of the extent and timing of precursory uplift (and subsequent subsidence) could place additional constraints on initial blob size and rise time through the upper mantle (*51*).

The plume initiation model for flood basalts also has important implications for understanding earth history and the evolution of the mantle. We discuss three points in particular:

1) Rampino and Stothers (*24*) showed that flood basalt events are similar in age to mass faunal and floral extinction events and have suggested that both are caused by periodic meteorite impacts. However, this appears to be implausible if hot spots and flood basalts are closely related; meteorites would not coincidentally land on mantle plumes and would most likely not produce hot-spot sources that remain relatively stationary with respect to plate motions for periods of ~100 million years. Rather we suggest that if there is a relation between flood basalts and extinctions, both are the result of plume initiation (*4*) following, perhaps, core-mantle boundary instability (*6, 52*).

2) We have inferred initial diapir volumes based only on extrusive (basalt) volumes and uniform 20% partial melting. Clearly our estimate of ~240-km-diameter blobs is a lower bound because there may be large bodies of intrusive rocks underlying the flood basalt plateaus (*11*). Also, the entire plume head will probably not undergo extensive partial melting; the lower portion may not melt if the plume cools significantly before it rises to the base of the lithosphere. Therefore, even larger initial diapirs may be required to account for the flood basalt volumes. We see no obvious way of forming such large diapirs in a region as shallow as the upper mantle, that is, from instability of a boundary layer at ~650-km depth (or shallower); the core-mantle boundary is the most plausible source.

3) The compositions of some flood basalts (for example, the Deccan traps) exhibit an enriched mantle component similar to ocean island (hot spot) basalts as well as variable contributions from the continental lithosphere and shallow, depleted suboceanic mantle (*7, 10, 14*). This composition is consistent with our model if plumes transport less depleted material to the surface from the deep mantle and assimilate upper-mantle and lithospheric material as they rise and melt. Therefore, much geochemical het-

erogeneity in and variation among flood basalt provinces is expected, depending on the plate tectonic setting in each case.

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Giant Radiation-Induced Color Halos in Quartz: Solution to a Riddle

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The radii of radiation-induced color halos (RICHs) surrounding radioactive mineral inclusions in mica generally correspond closely to the calculated range of common uranogenic and thorogenic alpha particles in mica. Many exceptions are known, however, and these variants have led investigators to some rather exotic interpretations. Three RICHs found in quartz are identified as aluminum hole-trapping centers. Whereas the inner radii of these RICHs closely match the predicted range of the most energetic common alphas (39 micrometers), the color centers observed extend to 100 micrometers. Migration of valence-band holes down a radiation-induced charge potential might account for enigmatic RICHs. Such RICHs provide natural experiments in ultraslow charge diffusion.

IN 1907 JOLY (*1*) POINTED OUT THAT microscopic color halos commonly observed surrounding small inclusions of radioactive minerals were caused by damage produced by alpha particles emanating from the inclusions. Shortly afterwards, Rutherford (*2*) noted a close correspondence between the radial size of halos and the energies of the alpha particles. A number of workers have described and measured these radiation-induced color halos (RICHs) and, from their sizes, have tried to match them with specific radionuclides in the inclusions [see also (*3*)]. Although it seems possible to relate the sizes of most of the described halos to alpha emitters in the U and Th decay chains, there are many exceptions. Particularly controversial have been two (perhaps artificial) classes of RICHs referred to as Po halos (*3–7*) and giant halos (*3, 8, 9*).

The Po halos are RICHs that have a size and ring structure apparently comparable with the range in silicate minerals of alpha particles emitted by uranogenic Po radioisotopes of mass 210, 214, and 218 (*4–7*), although this interpretation has been challenged (*10*). Significantly, rings that can be attributed to the other five alpha decays in the ^{238}U series seem to be lacking (*4–7*).

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That the half-life of ^{218}Po is 3 min has not deterred some investigators from proposing separation of Po from its radioactive progenitors before its inclusion in minerals (*4, 5*). Indeed, Po halos have even been offered as possible evidence of an instantaneous creation (*11*).

Giant halos are anomalous RICHs that have radii extending more than approximately 47 μm from edge of the inclusion. Suggested possible mechanisms of formation for these RICHs have included (i) high-energy alphas emitted by rare and undocumented isomers (*9*); (ii) postulated and unknown superheavy elements (*9, 12*); (iii) diffusion of radiolytic atomic H (*13*) and radioactive nuclides through what are structurally and chemically highly anisotropic mineral lattices; (iv) fissionogenic alphas from extinct ^{244}Pu (*14*); and (v) channeling of alphas through open cleavage cracks (*15*).

We have found three giant RICHs surrounding monazite inclusions in quartz grains extracted from two different granites (*16*). These RICHs are seen in reflected light as zones of smoky quartz beginning at a nearly spherical surface 39 μm from the outer edge of the inclusions. Smoky quartz is a common variety of quartz whose gray color is induced by ionizing radiation; the color center is $[AlO_4]^0$, and it is formed as a result of a hole in the valence band being