Mantle Plumes and Continental Tectonics

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Mantle plumes and plate tectonics, the result of two distinct modes of convection within the Earth, operate largely independently. Although plumes are secondary in terms of heat transport, they have probably played an important role in continental geology. A new plume starts with a large spherical head that can cause uplift and flood basalt volcanism, and may be responsible for regional-scale metamorphism or crustal melting and varying amounts of crustal extension. Plume heads are followed by narrow tails that give rise to the familiar hot-spot tracks. The cumulative effect of processes associated with tail volcanism may also significantly affect continental crust.

Plate tectonics, the "quantitative description of the kinematics of the lithosphere" (*1*), has provided a conceptual framework of the processes that operate at the margins of the tectonic plates that cover the Earth's surface. This framework has, however, proven less useful when it comes to processes operating away from plate boundaries (*1*).

Two prominent features that have defied a plate tectonic explanation are the oceanic volcanic (or hot-spot) chains and continental flood basalt (CFB) provinces. Both may contain basalts derived through the melting of mantle that is unusually hot relative to the mid-ocean ridge source and thus are likely to result in large part from temperature rather than compositional variations in the mantle (*2*). The CFB provinces and their oceanic equivalents, the basaltic oceanic plateaus, which were erupted at rates possibly as high as 22 km^3/year (*3*), are arguably the most spectacular magmatic features on the planet. A model of mantle plumes (*4*) appears to explain successfully a wide range of observations relating to both ocean island chains and flood basalt provinces (*5*). Mantle plumes most plausibly arise from a hot thermal boundary layer at the base of the mantle (*6–8*), whereas plates are part of the cool thermal boundary layer at the top of the convecting mantle (*9*). Plates and plumes are thus complementary, each driving a distinct form of mantle convection (*8*), and seem to operate largely independently (*10*).

In this paper we assess the importance of mantle plumes in initiating thermal and structural reworking of the continents and use the geological record to infer the extent of any interactions between plumes and plates. We then use an example from continental geology to illustrate how geological observations can provide an important test of the model.

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Historical Development

Wilson (*11*) was the first to suggest that lines of ocean islands such as the Hawaii-Emperor chain formed when the Earth's mobile surface passed over a fixed region of relatively hot mantle (a hot spot) where large amounts of magma were produced. Morgan (*6*) showed that hot spots on several plates had not moved discernibly relative to each other and suggested that these were regions where plumes of hot material ascended from the core-mantle boundary (CMB). He noted also that some hot spots could be traced back along volcanic chains to often distant flood basalt provinces, a number of which appeared immediately to precede continental rifting. Early fluid dynamics analog experiments (*12*) showed that a low-viscosity plume initiating in Earth's mantle would ascend as a spherical pocket of fluid fed by a pipe, that is, that the earliest stage of a plume would be the development and rise of a large-volume head, which would be trailed by a narrow and possibly long-lived conduit (or tail) through which buoyant source material could be added continuously to the ascending head (see Fig. 1A).

Morgan (*13*) suggested that flood basalts formed through melting of the heads of new (starting) plumes whereas the hot-spot chains were derived through melting of hot material rising in the long-lived tails that trailed and ultimately superseded the starting-plume phase. This suggestion, which was based on the results of experiments on plumes driven by buoyancy resulting from compositional (and thus density) differences between the plume source layer and overlying fluid (*12*), was taken up by Richards *et al.* (*14*), who identified ten couples of flood basalt provinces and hot-spot tracks that may have formed from plume heads and tails during the past 250 million years. However, extrapolation (*14*) from the fluid dynamics experiments on composition-driven plumes showed that the volume of the head of such a plume formed in the mantle will be relatively small [diameter of ~300 km for a plume having a density contrast $\Delta\rho$ of 0.1 g cm^{-3} and rising from the CMB (*14*)] compared with the extent of CFB provinces (diameters of ~2000 km).

Plumes can result also when a layer of fluid becomes buoyant upon being heated, and experiments have shown that the heads of thermally driven plumes enlarge considerably more during ascent than those of compositionally driven plumes. Enlargement occurs because the plume entrains a boundary layer heated by its passage (Fig. 1A); the heads of thermal plumes are estimated to reach diameters of about 1000 km if they start at the CMB (*4, 15*). These results have been used to argue that the plumes that give rise to both the CFBs and ocean islands arise from a thermal boundary layer at the base of the mantle (*5*). Further support for this suggestion comes from the correspondence between the measured heat content transported by plumes and the inferred heat flux across the core-mantle boundary (*8*).

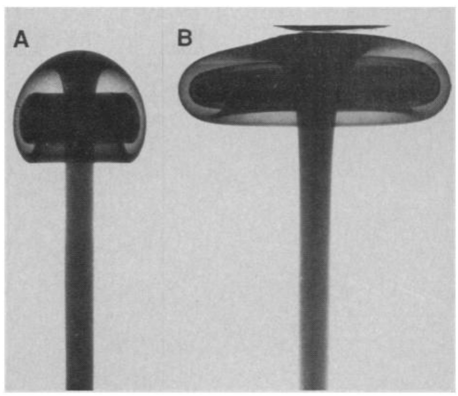

Fig. 1. Photographs of a starting plume in glucose syrup. (**A**) effects of entrainment into the rising vortex immediately before the onset of near-surface spreading; (**B**) the plume after lateral spreading has approximately doubled the head diameter. In these experiments hot source material was dyed blue, and appears black or dark gray in the photographs, while entrained surrounding material remains light-colored (*4*).

Dynamical Framework

Fluid dynamics experiments can be used (*4*) to show that at the time of separation from a hot low-viscosity source layer the diameter (D) of a new plume head varies as $D \sim Q^{2/5}(v/g\alpha\kappa\Delta T_s)^{1/5}$, where Q is the volume flux from the source layer into the ascending plume head, v is the kinematic viscosity of the overlying mantle, g is the acceleration due to gravity, α is the coefficient of thermal expansion, κ is the thermal conductivity of the mantle, and ΔT_s is the temperature excess of plume source material. Similarly, once separated from the source layer the diameter of the ascending head increases as $D(z) \sim Q^{1/5}(v/g\alpha\Delta T_s)^{1/5}\kappa^{2/5}z^{3/5}$, where z is the distance traveled. For conditions appropriate for the modern Earth, including buoyancy fluxes comparable with those feeding modern hot-spot tracks, and for constant material properties throughout ascent, plume heads are predicted to attain diameters of 800 to 1200 km during ascent through the mantle. Changing the assumed viscosity of the lower mantle from 10^{22} Pa s to 10^{21} Pa s, a value considered more appropriate for a hotter Late Archean Earth, decreases this estimate to 600 to 800 km (*4*). Varying the values assumed for source flux (Q) or source temperature (ΔT_s) over likely ranges (10^3 to 10^5 N s^{-1} and 200° to 800°C, respectively) likewise results in changes of the estimated head diameter of only a few hundred kilometers (*4*). An increase in the thermal expansion coefficient (α) with height in the mantle (*16*) will result in larger initial plume head volumes (*4*) but in less growth by entrainment during ascent. The final diameter of a plume head is rather insensitive to the precise values of properties such as α, and it is difficult to avoid the conclusion that the heads of plumes originating at the CMB will be large features.

Upon nearing the surface, a plume head originating at the CMB spreads (Fig. 1B) to give a disc of hot material predicted to be 1500 to 2500 km across and 100 to 200 km thick (*4, 17*). This is approximately the size of a number of CFB provinces, supporting the validity of the model (*5*). By way of contrast, a thermally driven plume head originating at the 650-km seismic discontinuity and driven by a typical hot-spot buoyancy flux is calculated to reach a diameter of only 600 km after near-surface spreading (*4, 17*). Slow-moving plume heads fed by fluxes much less than the smallest fluxes observed (*8*) are not expected to reach the surface because they become sheared out by the flow associated with plate movement (*5*).

For the modern Earth, the maximum temperature variation within basalt sources appears to be ~250°C, equivalent to a range in mantle potential temperatures (T_P) of 1300° to 1550°C (*2*). [(T_P is defined by McKenzie and Bickle (*2*) as the temperature that mantle material would have if raised to the Earth's surface along an adiabat.] On the basis of the model, average T_P's in plume heads are estimated to be of intermediate values, ~1350° to 1400°C, whereas hot source material rising in the plume conduit can have a T_P as high as 1550°C. Surface uplift due to the positive buoyancy of the hot plume material has been quantified [(*4, 17, 18*) Fig. 2] at as much as ~1000 m. This amount of uplift can lead to extension of overlying crust (*19*) and is predicted to precede initiation of voluminous basaltic volcanism by 3 to 30 million years (*5, 17*); the time interval is particularly dependent on the viscosity of the upper mantle assumed in the model. Lateral variations in the rheology of the uppermost mantle, especially those associated with different lithospheric thicknesses, are predicted to play an important role in determining both the minimum depth reached by rising plume heads, as well as many of the details of plume-lithosphere interaction (*4, 17*).

Continental Plume Provinces

A number of observations on CFB provinces, including their diameters of ~2000 km, evidence for uplift before eruption of the basalts, and the compositional and isotopic characteristics of erupted basalts, are explained satisfactorily by the starting plume model (*5*). Acceptance of the plume model for the origin of ocean island chains and CFB provinces (*5, 14*), and for the origin of oceanic volcanic plateaus (*20*), raises the question: "What else happens when a plume rises beneath a continent?" The experiments allow several inferences to be made. Two types of plume provinces should be recognizable, depending on whether the crust passes over a plume head or a plume tail. Plume head provinces should be equant and 1500 to 2500 km across, whereas volcanism and tectonism in plume tail provinces should be restricted to a narrow (<300 km wide) linear belt, that is a hot-spot track.

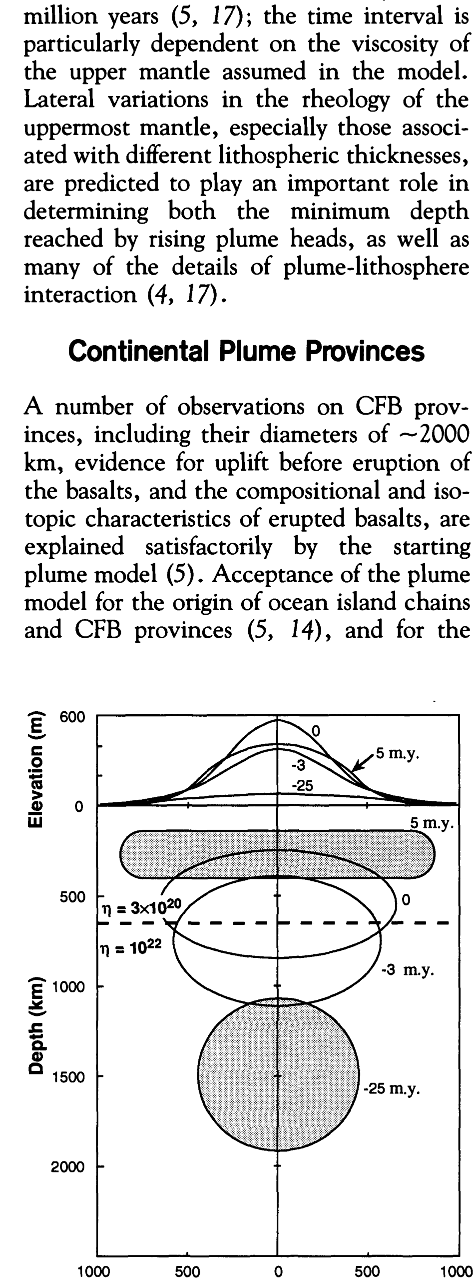

Fig. 2. Diagram showing the calculated surface uplift above a rising plume head. Modified from (*17*); m.y., million years.

Both types of plume province may contain basaltic magmatism, including high-Mg varieties (picrites, komatiites) that result from melting of high-T_P material ascending within the plume conduit (*5, 21*). Both may be associated with extension. Uplift is expected to predate both extension and initiation of basaltic magmatism in both types of plume province. Upward transfer of heat from the plume may initiate partial melting of overlying continental crust (*22*). Conductive heat transfer into overlying mantle and crust may play an important role in the development of both types of plume province over time scales greater than ~10 million years because it can lead to weakening (and thus easier extension) of the crust and uppermost mantle.

Finally, because of the thickness (100 to 200 km) and depth beneath the surface (50 to >100 km) of the hot plume layer, the time scale for cooling is considerable (of order hundreds of millions of years). The slow subsidence resulting from the cooling of a plume emplaced beneath continental crust could result in the formation of long-lived sedimentary basins (*5*). We term the entire process of thermal and structural reworking above mantle plumes "plume tectonics."

The most definitive evidence for plume involvement in continental reworking is the presence in the crust of magmas derived through melting of anomalously hot mantle (*21*). Modern-day picrites, which derive from mantle with a potential temperature up to 250°C hotter than that which yields mid-ocean ridge basalts (*2*), and komatiites, which were derived from even hotter regions in the Archean mantle (*21*), have been interpreted as the products of decompression melting of hot material ascending in plume conduits (*4, 21*). Thus, in searching the geological record for examples of continental plume provinces, we have focused initially on areas where picrites or komatiites are found interbedded with basalts produced from much cooler mantle and where there has been more or less synchronous eruption of crust-derived melts.

There are examples of both types of plume province. Regions that we infer to have resulted from impingement of a plume head include the flood basalt provinces of the Karoo of southern Africa, the Deccan of western India, and the North Atlantic Tertiary province (*5, 13, 14*), and the Late Archean granite-greenstone terranes of the Yilgarn Block of Western Australia (*21, 22*); plume tail provinces equate with the familiar hot-spot tracks such as that associated with the Yellowstone plume or with lines of volcanic islands such as the Hawaii-Emperor chain.

Plume head provinces. Of the plume head provinces, the Yilgarn and Karoo, in particular, have extended magmatic histories consistent with conductive heat transfer upward from the top of plume heads that had presumably ascended to near the base of the crust. Both are clearly plume-related because of the high mantle temperatures implied by the presence of either komatiites (Yilgarn) or picrites (Karoo).

In the southeastern Yilgarn province, eruption of a thick (to 8 km) sequence of basalts that contains komatiite and komatiitic components began at ~2715 million years ago (Ma). The basalts are underlain by felsic volcanic rocks that were erupted at 2940 Ma. These rocks provide direct evidence that the basalts were erupted onto continental crust (*22*). Further evidence for the presence of continental basement comes from the presence of xenocrystic zircon in felsic volcanic rocks, granites and basalts; and from lead isotope compositions of granites, and from ores from the basaltic sequences (*22, 23*). The basaltic rocks are overlain by felsic volcanic rocks, which began erupting at 2688 ± 8 Ma, and volcanogenic sedimentary rocks. Both basaltic and felsic volcanic sequences are intruded by granites of similar (2685 to 2690 Ma) age to the felsic volcanic rocks; felsic magmatism was followed by minor mafic magmatism. This history provides direct evidence that there was a thermal anomaly in the mantle that overlapped in both time and space with the production of large volumes of felsic melts. More granite was emplaced during a second major period of felsic magmatic activity 2660 to 2665 Ma, and small, isolated granitic intrusions have ages as young as 2600 ± 10 Ma. The overall time scale for magmatism is thus ~100 million years. Derivation of the granitic rocks through partial melting of preexisting continental crust now appears to be accepted by most workers (*22, 24*). Evidence for this includes lead isotope compositions, the difficulty of producing large volumes of K-rich granite directly from the mantle, and the presence of old xenocrystic zircons in a number of samples (*22*).

These observations have been interpreted as follows (*21, 22*). Partial melting in the bulk of a rising and spreading plume head resulted in an early period of (often voluminous) basaltic volcanism, whereas melting of the much hotter material that ascended in the axial conduit produced the komatiites. Conduction of heat upward resulted (after a time delay of ~25 million years) in the production of crustally derived melts. This first anatectic episode, which produced both the early (2685 to 2690 Ma) granites and volcanic rocks, was voluminous but apparently of brief duration, perhaps only several million years. Continued conduction of heat into the crust resulted in the production of later anatectic melts, including a second major episode having high SiO_2, K_2O, and Na_2O contents, low mafic mineral abundances, and negative Sr and Eu anomalies, all interpreted as indicating derivation from originally structurally higher material that underwent partial melting at lower temperatures and pressures than those prevailing during the first episode. The youngest granites have characteristics (they are alkali-rich, water-poor, and may contain fluorite or alkali pyroxenes) often interpreted as indicating a high-temperature origin (*25*). The end result of Late Archean reworking has been the production of typical internally differentiated, layered stable crust (*26*). Later episodes of basaltic magmatism (for example, during emplacement of a widespread suite of gabbroic dykes at ~2400 Ma) are not associated with important crustally derived magmatism, presumably because components with low melting points had been removed from the crust in Late Archean time (*22, 27*).

The presence of Late Archean granites throughout the Yilgarn craton shows that this crustal reworking event (*22*) affected profoundly an approximately equant area ~800 km across. If this was the result of a plume event, and if plumes commonly play an important part in initiating continental breakup (*13, 28*), then this is likely to be a fragment of an originally larger area.

The much younger Karoo sequence of southern Africa has many similarities. A brief period of basaltic volcanism at 193 ± 5 Ma (*29, 30*) produced thick sequences of picrites near the inferred position of the plume track (*5*). In the Lebombo area, the basalts are overlain by felsic volcanic rocks of crustal derivation; these have ages of ~177 Ma (*29*) and are interbedded with, and overlain by, basaltic rocks. Small syenite intrusions are as young as 130 to 135 Ma (*29, 30*). Felsic volcanism is only important off the craton, presumably because the lower part of the relatively young crust that encircled the craton had not been stripped of anatectic components and also possibly because the thinner lithosphere beneath this younger crust permitted easier ascent of the hot plume head (*28*).

Felsic magmatism is much less important in many other areas that have been interpreted in terms of the starting plume model, such as the Deccan and the North Atlantic Tertiary Province (resulting respectively from initiation of the Reunion and Iceland hot spots (*5, 13, 14*). We infer that this difference is a consequence of the earlier extraction of low-melting components from the lower and middle crust of these regions.

The magmatic records of the Yilgarn and Karoo differ fundamentally from those of consumptive margins (that is, island and continental arcs) (Fig. 3). The key features of the Yilgarn craton are its equant form; scale (>800 km); presence of komatiites; magmatic sequence of basalt, granite, and later diverse magmatism; the occurrence of brief but voluminous episodes of magmatism separated by periods of quiescence; and the evidence that the felsic rocks were derived through partial melting of continental crust rather than from subducted oceanic crust.

These features contrast with the situation in magmatic arcs clearly associated with subduction, which usually have an elongate form; lengths of thousands but widths of only a few hundreds of kilometers; no clear evidence for the presence of high-temperature mafic magmas; a magmatic continuum from gabbro through granite, with more or less simultaneous eruption of compositionally diverse rocks for tens of millions of years; and an inferred origin for the more mafic rocks through dehydration melting of amphibolite [Fig. 3 (*31, 32*)]. Although some rocks that may have formed through dehydration melting of amphibolite (the so-called tonalite-trondjemite-granite suite) do occur in Yilgarn craton, they are of only minor importance and may have formed through partial melting of buried greenstones rather than through melting of subducted oceanic crust (*22*).

Other continental magmatic terrains such as the Mid-Proterozoic magmatism of the United States midcontinent and of northern Australia (*33*) and the Late Caledonian granites of Scotland and Australia (*34*) (Fig. 3), which are difficult to interpret in plate tectonic terms, have similarities with the Yilgarn (in particular, geographic form, scale, and derivation through partial melting of older continental crust). We suggest that reworking above plume heads deserves consideration as a potentially important process in the evolution of the observed continental crust.

Plume tail provinces. The most obvious feature of plume tail provinces is the presence of a time-progressive chain of volcanic centers that may be many thousands of kilometers long, corresponding to plume tails that may have persisted for as long as 200 million years (*13*). The longevity of the main volcanic interval at any given site on the chain is predicted to be related to the rate at which the overlying plate passes across the hot-spot conduit. The province resulting from the Yellowstone plume (the Snake River Plain) provides one on-land example. The Plain is a volcanic-filled graben atop a broad arch (35). Early rhyolitic volcanism is followed by basaltic volcanism, a main phase of crust-derived magmatism, and a final episode of plain-flooding basaltic volcanism. The entire volcanic history takes ~8 million years at any one locality (36), consistent with passage of North America at 25 km per million years over a plume conduit 200 km across.

Although lateral spreading of material from the top of the conduit may result in the production of a strip of hot plume material ~1000 km wide (8, 37), a striking characteristic of continental plume tail provinces (as well as of oceanic hot-spot chains) is the restriction of major tectonism and magmatism to a narrow (commonly <200 km) volcanic- and sediment-filled trough atop a broad arch. Thermal subsidence of the arch may result in later and much more widespread sedimentation. Formation of the graben associated with the Snake River Plain has been suggested to result from removal of previously underlying crust through melting and eruptive dispersal (35).

Plumes and Crustal Reworking

How important is the reworking of continental crust above a plume head, and how frequently might such reworking occur? Both questions can be addressed with the geological record. The first question has been answered above. The model developed for the Yilgarn craton carries the implication that thermal and structural reworking of continental crust above a plume head can play an important part in the development of that crust. Answering the second question requires an estimate of the frequency with which a plume head might rise beneath a particular piece of continental crust.

Rate of plume-initiated crustal reworking. Estimation of the average time (the repeat time) between plume events affecting a given piece of crust is complicated because: (i) two-thirds of the Earth is covered by oceanic crust, which has a mean age of only 53 Ma (38); (ii) the effects of plumes (particularly weak plumes) may be difficult to distinguish in the geological record; (iii) the easily observable record (the past 200 million years) may not be representative of longer time periods; (iv) plume heads may overlap; and (v) the very processes expected above a plume head (uplift and erosion followed by slow subsidence) tend to obliterate obvious plume-related features such as volcanism.

An assessment of likely plume head–plume tail couples indicates that at least eight major plumes resulting in CFB provinces (Yellowstone, Afar, Iceland, Reunion, Tristan, Tasmania-Ferrar, Karoo, Siberian traps) have risen beneath predominantly continental crust during the past 250 million years; another eight, probably smaller, examples (East Africa, Raton, Eastern Australia, Trindade or Martin, New England, Meteor-Discovery-Bouvet, Australian Northwest Shelf, Fernando) have also been suggested (39). The actual number of new plumes could be considerably greater than this. Our assessment is meant to be indicative rather than definitive; we use these best examples to show the likely bounds that might be placed on the repeat time (40). Eight large (radius R = 1200 km following near-surface spreading) plume heads in 250 million years implies a repeat time for the rise of a plume head beneath a particular spot on the Earth's surface of 1300 million years; addition of a further eight medium-sized (R = 1000 km) plume heads brings the estimated repeat time down to 800 million years [see (40) for details]. For comparison, inclusion of an extra 16 small (R = 800 km) plumes decreases the estimated repeat time to 500 million years. Sixteen new plumes rising beneath continents in 250 million years implies a global total of 43 (Fig. 4) if we assume that plume initiation is independent of near-surface processes (10). Although the values for plume head radius used in these calculations derive from the fluid dynamics experiments, the close correspondence between predictions and the observed dimensions of one of the inferred surface manifestations of plume heads (the CFB provinces) is taken as indicating that the major uncertainties in estimating the repeat time result not from poor estimates of plume-head radii but from incomplete knowledge of the rate of plume initiation and of the size-frequency distribution of plume heads.

A reasonable estimate is thus that a piece of continental crust will be affected by a plume head event on average every 500 to

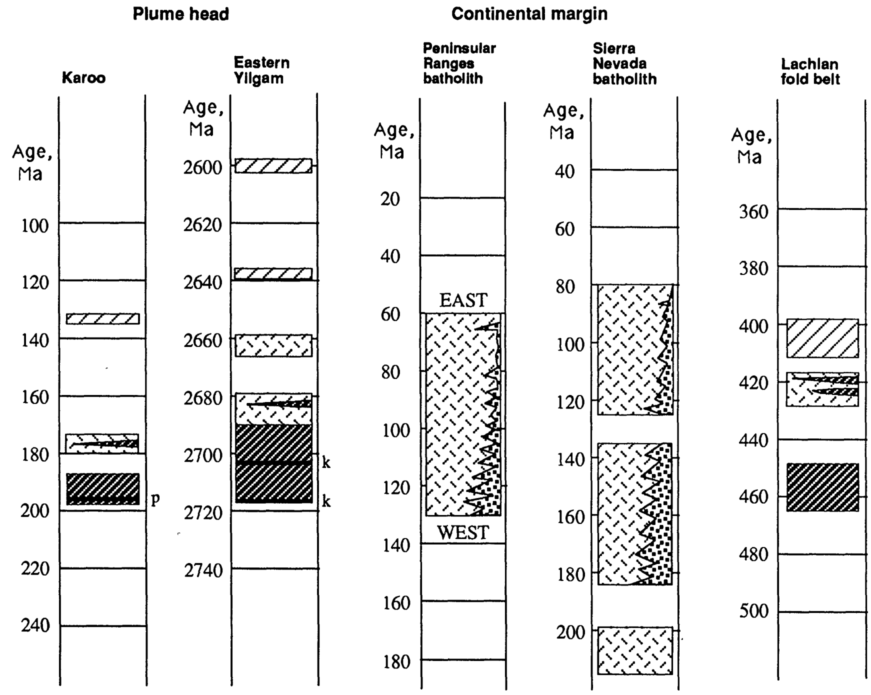

Fig. 3. Stratigraphic columns summarizing the magmatic history of the Yilgarn and Karoo, two inferred plume head provinces, and comparing them with two continental margin arcs, the Peninsular Ranges (PRB) and Sierra Nevada (SNB) batholiths of western North America, and with the Lachlan Fold Belt of southeastern Australia. Extent of gabbroic and intermediate magmatism shown for the PRB and SNB is diagramatic only, and is intended to show that the proportion of more mafic magmatism decreases through time (54). The magmatic record of the plume head provinces, which show the sequence mafic, felsic, mafic, felsic (Yilgarn only), and late high-temperature felsic magmatism, is in marked contrast to the record of continuous, compositionally variable magmatism characteristic of the continental margin arcs, but is similar to that of the Lachlan, an area that has proven difficult to interpret in terms of plate tectonics. Data from (22, 29, 30, 54).

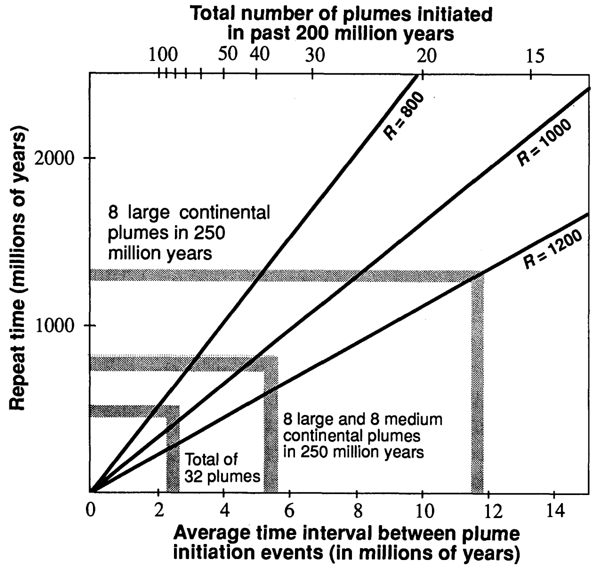

Fig. 4. Diagram showing the relation between frequency of plume initiation and plume repeat time (*40*) for a range of inferred plume head diameters. 200 Ma is chosen for comparison because it is approximately the age of the longest lived present-day plumes (*13, 28*). Radii (*R*) are for plume heads before initiation of lateral spreading, which approximately doubles the plume head diameter. Details of the calculation are given in (*40*); see text for discussion. The repeat times shown here correspond to the time required for plumes to affect 63% of the area under consideration.

800 million years (Fig. 4). The geochronological record of the continents shows apparent episodicity, with periods 100 to 200 million years containing many dated samples separated by much longer periods of magmatic quiescence (*41*). Although these data are usually interpreted as measuring the times of continental growth, we suggest that the geochronology of the continents records the combined effects of crustal growth (which occurs predominantly at consumptive plate margins) and crustal reworking, including that above plume heads.

Crustal reworking above plume tails is likely to be much less important. The methodology of Davies (*8*) can be used to estimate that plume-related features are produced at a rate of about 10^6 km^2 per million years. The mean age of oceanic crust (area: 310×10^6 km^2) is 53 million years (*38*), so that ~15% of the present ocean floor has been affected by hot-spot activity (or, 1% of the oceans is affected by hot-spot activity every 3 million years). If the preserved record is representative, and if continents are as likely to be affected as are oceans, then the calculated repeat time of 300 million years is in broad agreement with earlier estimates (*42*). However, these values are calculated for the entire area inferred to be underlain by plume material, the width of which is much greater [up to about 1000 km (*8, 37*)] than the width (commonly <200 km) of any major geological feature clearly related to the underlying hot spot. Because of this, the repeat time for significant reworking above a plume tail is probably of order 1000 to 2000 million years; plume tail provinces are thus predicted to be a less important feature in the geological record than plume head provinces.

The estimates of repeat time can also be used to calculate (*43*) the flux of volume (V), mass (M), and heat (Q) carried into the upper mantle by plumes. For plume heads these are $V = 2 \times 10^3$ m^3/s, $M = 6.6 \times 10^6$ kg/s, and $Q = 0.5$ to 1.0×10^{12} W (1.0 to 2.0 mW/m^2) for a repeat time of 500 million years; and $V = 1.4 \times 10^3$ m^3/s, $M = 4.6 \times 10^6$ kg/s, and $Q = 0.4$ to 0.7×10^{12} W (0.7 to 1.4 mW/m^2) for a repeat time of 800 million years. This compares with estimates of $V = 2.2 \times 10^3$ m^3/s, $M = 7.3 \times 10^6$ kg/s and $Q = 2.3 \times 10^{12}$ W (4.5 mW/m^2) for plume tails (*8, 37*). The repeat times for heads are less than for tails; total heat flux is also less, and volume and mass fluxes are comparable. Much of the difference in estimated heat fluxes between heads and tails results from a lower estimated average temperature for plume heads.

If the total mass flux (heads plus tails) for plumes estimated on the basis of the record of the past 250 million years (1.2 to 1.4×10^7 kg/s) is representative, then the plume flux alone is capable of turning over the upper mantle ($mass = 1.27 \times 10^{24}$ kg) on a time scale of 3000 million years.

Plumes and Plates

One further question is the extent and nature (if any) of the interaction of plumes with the convective system driving the tectonic plates. Although there is still much controversy regarding the origin of the forces that move the plates, models of whole-mantle convection driven by the negative buoyancy of a cold, rigid surface boundary layer (principally the oceanic lithosphere) are consistent with a wide variety of observations (*44*). Regardless of the details of the driving mechanism, this plate-scale convective mode is responsible for most of the heat loss from the mantle, and >80% of the heat loss from the Earth (*9, 44*). The plumes that give rise to hot-spot chains (plume tails) are inferred to be derived from a thermal boundary layer at the CMB (*5, 8, 44, 45*), although this is also controversial, and to carry through the mantle ~6% of the heat ultimately lost at the Earth's surface (*8, 37*). When the plume head plus tail fluxes are combined (2.8 to 3.3×10^{12} W, or 5.2 to 6.5 mW/m^2), the total amount of heat transported through the mantle by plumes is equivalent to 7 to 9% of the mantle-derived surface heat loss [38×10^{12} W; 74 mW/m^2 (*8, 37*)].

The ultimate driving force for plate tectonics is loss of heat from the mantle, whereas plumes appear to be driven by heat loss from the core (*8, 9, 37, 44, 45*). The areally averaged heat fluxes estimated for the two boundary layers [18 to 22 mW/m^2 for the CMB; 74 mW/m^2 for the mantle component of surface heat flow (*8, 37*)] differ by only a factor of 3 or 4. Their tectonic manifestations are so different because of the temperature dependence of mantle rheology. The hot boundary layer above the core has reduced viscosity, which gives rise to plume heads and tails, whereas the cool surface boundary layer forms stiff plates that subduct as sheets and drive the plate-scale convective flow. The poor spatial correlation between hot spots and plate boundaries (*10*) indicates that the two modes of convection operate largely independently.

However, the geological record indicates that there is at least some interaction between plumes and plates. The clearest example is the observation that continental breakup is often immediately preceded by the development of flood basalt provinces (*6, 13, 46*). Data from a number of such provinces shows that there is a continuum (Fig. 5), from those where initiation of a new spreading ridge followed soon after basalt eruption (for example, North Atlantic Tertiary Province; Deccan; Paraná-Etendeka) through those where a new ocean basin formed only after a significant (15 to 40 million year) time lag (for example, Afar, Karoo, Newark-Palisades) to situations where no new ocean formed (for example, the Siberian traps).

Observations suggest that for the modern Earth the rise of a plume by itself cannot initiate continental breakup. But in situations where the plate-scale motions are suitably arranged, the extra horizontal deviatoric stresses generated by uplift above a rising plume may at times be of sufficient magnitude to allow rifting to either proceed more rapidly or result in transfer of the axis of spreading to as near as is possible to the plume center. For example, spreading within the South Atlantic propagated rapidly northward following the rise of the plume head which gave rise to the Paraná-Etendeka CFB province. Examples of plume-initiated ridge jumps (*28*) include the opening of the North Atlantic following ascent of the Iceland plume, and the rifting of the Seychelles away from India following arrival of the Reunion plume, the head of which is inferred to be the source of the Deccan basalts (*5, 14*).

When a plume rises beneath a continent far from an active spreading ridge, spreading may be delayed or not occur at all. The sequence of events recorded along the southeastern margin of the United States before the opening of the Central Atlantic (*28, 47*) may provide an example of delayed spreading. Here subsidence (and slow extension) was under way by 230 Ma, presumably in response to plate-scale motions. Cessation of sedimentation in many areas 210 to 215 Ma may be interpreted (*28*) as resulting from uplift presaging the arrival of the Fernando (Newark-Palisades) plume head at 201 ± 2 Ma; oceanic crust began forming about 175 Ma. The present position of this plume has been suggested to be the weak Fernando de Noronha hot spot off the northeastern coast of South America (*28*). The time lag between plume arrival and continental breakup is explicable in terms of the numerical experiments of Houseman and England (*19*), who showed that breakup (or "runaway extension") occurs when an appropriate combination of rheology and elevation is attained (*28*).

Extrapolation from the fluid dynamics experiments indicates that the maximum elevation likely to be attained above a large plume head is 500 to 1000 m, unless lateral displacement of lithosphere becomes important (*5, 17*); continued ascent of hot source material up a strong axial conduit may increase the ultimate amount of uplift associated with plume emplacement. However, it appears that even in situations where slow extension is under way before arrival of the plume, the magnitude of the extra forces generated above the plume head are incapable of forcing breakup unless the crust is already hot, or unless sufficient heat can be conducted upward from the plume head to alter substantially the rheology of the overlying mantle and crust. The time scale for this appears to be comparable to the time taken to raise the lower crust to its melting point, that is, 15 to 40 Ma (*22*). This analysis also accounts for why breakup, when it does occur, is often localized along relatively young orogenic zones: the thinner and weaker lithosphere typically associated with younger crust allows the plume to ascend farther (*28*); thus increasing uplift as well as shortening conductive distances.

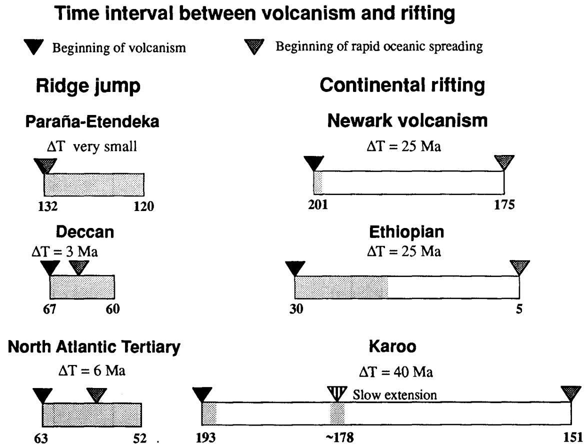

Fig. 5. Diagram showing the length of the interval between plume arrival (as shown by the first record of mafic volcanism) and initiation of oceanic spreading for six Mesozoic or Tertiary examples. Stipple shows approximate extent of volcanic activity. Although the beginning of slow extension between southern Africa and Antarctica began at 180 ± 5 Ma (*56*), coincident with a period of felsic volcanism (and thus rheologically weak lower crust) in the eastern Karoo at ~177 Ma, rapid oceanic spreading was not initiated until much later, 150 to 152 Ma. Data from (*13, 28–30, 46, 47, 55*).

Crustal Production

Do plumes contribute directly to continental growth? For example, Hill *et al.* (*22*) estimated that as much as 15% of the Late Archean crust of the southeastern Yilgarn craton resulted from contemporary additions of extrusive basalt and intrusive gabbro. We attempted to evaluate the possibility that basalts derived through melting of ascending plume material may be a significant contributor to continental growth. It has been suggested also that obduction of the oceanic equivalent of the CFB provinces, the oceanic plateaus, may be an important mechanism for growth of the continental crust (*20, 48*). Sandwell and Schubert estimated an upper bound to this mechanism of about 4 km^3/yr (*48*); actual rates most likely are much lower than this and extremely difficult to evaluate.

We suggest that although neither CFB provinces nor obducted oceanic plateaus such as Wrangellia (*20*) appear to form more than a minor component of the continental mass, their relative areal importance may underestimate the real contribution they make to crustal growth. Both CFBs and obducted oceanic plateaus undergo rapid weathering and erosion. The dominant products of weathering are clay minerals, which are light enough to resist subduction and thus will remain as part of the continental mass. Crustal addition rates from eruption of CFBs may be 0.1 to 0.3 km^3/year for short time periods (~5 million years); long-term rates are 0.02 to 0.05 km^3/year (or 0.6 to 1.5 $\times 10^{11}$ kg/year) for the eight CFB provinces since 250 Ma. At this rate, plume volcanism could contribute a few percent of the present continental mass (2.3 $\times 10^{22}$ kg) in 3800 million years. Although CFB extrusion and contemporaneous intrusion has probably played no more than a minor role in the growth of the continental crust, it could be an important contributor of elements enriched in basalts, such as nickel and chromium.

Looking Ahead

A quantitative dynamically-based model such as that now being developed for mantle plumes is not only testable, but may be useful also in providing a framework for the interpretation of complex observations. For example, the sequence of events associated with the opening of the Central Atlantic (early subsidence, uplift, basaltic volcanism, and formation of a new ocean after a lag of 25 million years) is consistent with that predicted for the rise of a starting plume beneath a slowly extending continent; this type of success serves to focus attention on key predictions of the plume model (such as prevolcanism uplift), which are not made by other models [such as the passive rifting model (*46*), for which the predicted time scales are inappropriate (*5, 28*)].

An example that illustrates the advantages of a predictive model concerns the later Tertiary evolution of the western United States. If much of the magmatic and tectonic complexity of this region results from the rise of two starting plumes (Raton ~34 Ma; Yellowstone 17.5 Ma (*35*), then the model predicts that the areas now occupied by, respectively, the southern and central Rocky Mountains should have been uplifted before eruption of the first plume-related basaltic rocks. The record of sedimentation in the Gulf of Mexico (*49*) provides evidence that this was the case—uplift of the southern Rocky Mountains was under way by 37 Ma, whereas uplift of the central and northern Rocky Mountains began a little later, at 23 to 24 Ma. Other evidence points to uplift of the southern Rocky Mountains commencing at about 38 to 40 Ma (*50*). The time intervals of ~6 million years between initiation of uplift and the first appearance of basaltic magmatism in both the southern and central Rocky Mountains may be related to the rise of two new plume heads. If the plume model is appropriate, for R = 500 km before spreading and ΔT = 100°C, a time interval of 6 million years implies that the upper mantle had a viscosity of 1 to 2 $\times 10^{20}$ Pa s (*17*). We infer from the presence of (probably subduction-related) volcanism that predates the inferred times of plume arrival that this area was already underlain by unusually hot (and thus low-viscosity) uppermost mantle. The low value of 1 to 2 $\times 10^{20}$ Pa s calculated from the observed time interval is one commonly associated with unusually hot mantle (*51*). This accord between the calculated viscosity and the likely low value inferred from the geological observations provides support for the starting plume hypothesis, and suggests that further application of the model in an attempt to unravel the forces responsible for the past 40 million years of geological evolution of this region may be productive.

An alternate mechanism for large-scale continental reworking is that of Himalayan-type continent-continent collisions. Although this may result in the structural and thermal reworking of broad areas many thousands of kilometers across (*1, 52*), it might not be expected to include the substantial and early basaltic volcanism that is a common characteristic of plume head provinces. Other features (*52*) of Himalayan-type events (for example, maximum temperature attained within the crust and relative timing of major structural and magmatic events) are also likely to differ significantly from those associated with plumes. In summary, each mechanism for initiating large-scale thermal reworking of continental crust is likely to have a characteristic combination of time scales and magmatic evolution that could allow identification.

Three areas of uncertainty limit detailed application (or evaluation) of the plume model. First, improved resolution of the predicted time interval between uplift and initial magmatism requires better knowledge of the values and variability of upper mantle viscosity, as well as a better understanding of the process by which plumes penetrate the lithosphere [see (*17*) for a discussion]. The latter point in particular is at present poorly understood. Plume penetration may involve thermomechanical erosion, heat transport by magma, and the development of gravitational instabilities in the lower part of the lithosphere. Indeed, although both the compositions of inferred plume-derived basalts and the observed time intervals between initial basaltic magmatism and the beginning of voluminous crust-derived magmatism appear to require that the top of the plume head ascends to within a short distance (~10 km) of the base of the crust (*22*), the mechanism by which this is accomplished is unknown.

Second, calculation of the uplift and extension history of crust above a rising plume head is limited by knowledge of the temperature-dependence of rheology (*19*). Improvement in our knowledge of the material properties of upper mantle and lower crustal rocks, and of the processes involved in plume-lithosphere interaction, is needed for more precise modeling.

Finally, we have as yet only a poor understanding of the likely extent of the interaction between a rising plume and subduction. The geological record of the continental margin of the northwestern margin of the United States does give some idea of one possible situation where a plume head rose through a subduction zone, one expression of which is the Cascades magmatic arc, developed for at least the past 50 Ma as the Juan de Fuca Plate has been subducted beneath western North America (*53*). A postulated Yellowstone plume head nearing the top of its ascent 17 to 18 million years ago must have reached the upper part of the descending slab. At about 18 Ma the rate of magmatic activity within this arc slowed considerably, and when it increased again 4 million years later, magmatism had shifted to more mafic compositions (*53*), consistent with the introduction of hot material from the plume head into the mantle wedge overlying the downgoing slab. Geometric considerations suggest that part of the plume head may have spread westward as far as the Juan de Fuca ridge. This example suggests that a better knowledge of plume-slab interactions may lead to a better understanding of the tectonic and magmatic complexities of some arcs.

The apparently successful application of the mantle plume model to the origin of continental flood basalt provinces, oceanic plateaus, and hot-spot tracks carries the implication that mantle plumes may play an important role in continental geology. We suggest that further candidates for plume head provinces include the Trans-Hudson orogen of the North American Proterozoic, the Proterozoic anorogenic magmatism of the American midcontinent and of northern Australia, and the 'Late Caledonian' granite terrains of Scotland and of southeastern Australia. The Midcontinent Rift of North America may be a candidate for a Proterozoic hot-spot track. Mantle plumes may also exercise considerable control over the properties of the top half of the subcontinental upper mantle.

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Sci. 75, 259 (1988)] commented that "The relationship of these late granitoids to the collision process is obscure, and hitherto their ages and compositional variations have not accorded with those predicted by various plate models." B. W. Chappell and W. E. Stephens (*ibid.*, p.71), comparing granites of similar (Late Caledonian) age in Scotland and southeastern Australia with the Mesozoic continental margin batholiths of western North America, concluded: "The Paleozoic granites show little evidence of such a direct relationship to subduction." See also A. N. Halliday and W. E. Stephens, *Phys. Earth Planet. Inter.* **35**, 89 (1984); M. Owen and D. Wyborn, *Austral. Bur. Min. Res. Bull. No. 204* (1979).
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40. The simple repeat time (T_R) is calculated from the relationship $T_R = S/NA$, where S is the area of the continents (1.9×10^8 km^2), N is the number of plumes per unit time, and A is the cross-sectional area of the plume head after near-surface spreading. This estimate is used because it is more readily compared with the observed episodicity in the geochronologic record of the continents. An alternate method is to use a probability calculation of the time required for plumes to affect a specified percentage of the surface area; the definition of T_R corresponds to the time for 63% of the surface to be affected.
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43. For plume tails, the volume flux (V) = $B/\rho\alpha\Delta T$, where B = differential mass flux [termed "buoyancy flux" by Sleep (*37*)] = 54.9×10^3 kg/s^{-1} (*37*); heat flux (Q) = $\rho C\Delta T$. The factor of 2 variation in the estimates of Q for plume heads arises from uncertainty in estimates of the amount of entrainment; Q is calculated for source per total ratios of 0.25 and 0.5. $\rho = 3300$ kg/m^{-3}; $\alpha = 3 \times 10^{-5}$ °C^{-1}; $C = 1.25 \times 10^3$ J/kg^{-1}/°C^{-1}. Increasing α as a function of decreasing pressure (*16*) will have the effect of increasing the source per total ratio.
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Hybridization of Bird Species

Peter R. Grant and B. Rosemary Grant

Hybridization, the interbreeding of species, provides favorable conditions for major and rapid evolution to occur. In birds it is widespread. Approximately one in ten species is known to hybridize, and the true global incidence is likely to be much higher. A longitudinal study of Darwin's finch populations on a Galápagos island shows that hybrids exhibit higher fitness than the parental species over several years. Hybrids may be at an occasional disadvantage for ecological rather than genetic reasons in this climatically fluctuating environment. Hybridization presents challenges to the reconstruction of phylogenies, formulation of biological species concepts and definitions, and the practice of biological conservation.

Species of sexually reproducing organisms are "groups of actually or potentially interbreeding natural populations which are reproductively isolated from other such groups" (*1*). Periodically attempts have been made to improve on this definition by dealing inter alia with the awkward fact that for some populations the criterion of demarcation is not absolute (*2–5*). Some populations occasionally interbreed, and then the question becomes one of determining the fates of the offspring (*1, 6*). Therefore, hybridization, which strictly is the interbreeding of species, is of pivotal importance in two respects: in framing ideas about the nature of taxonomic judgments to be made about particular populations (*7*) and more generally for understanding biological processes of evolution including speciation (*1, 2, 8*).

Traditional approaches to the study of hybridization have included the crossing of lines in the laboratory or greenhouse for genetical analysis, and the estimation of frequencies of phenotypic or genotypic classes in nature, their mating pattern, and their reproductive success. By themselves each is incomplete. In this article, we describe the desired but rarely achieved direct study of hybridization in nature through pedigree analysis. The study populations are birds. We present new information on the consequences of hybridization in populations of Darwin's finches over several generations.

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The Broad Patterns

Mayr and Short (*9*) estimated that approximately 10% of 516 nonmarine species of birds regularly hybridize. Meise (*10*) made a broader survey and concluded that 2% of all recent bird species hybridize regularly, and an additional 3% hybridize occasionally. A more definitive estimate can now be made. In the last 2 years the total number of