An alternative mechanism of flood basalt formation

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Abstract

All large continental igneous provinces and most high-temperature magmas (picrites, komatiites) are found on the margins of cratonic lithosphere. The standard plume model of flood basalt formation offers no explanation for this observation. We propose that thick lithosphere (usually Archean) adjacent to thinner lithosphere may control the locations of flood basalt provinces. The boundary between thick and thin lithosphere focuses both the strain in the lithosphere and the upwelling convection. In addition, the non-uniform boundary condition actually induces a small-scale form of convection that is not present in simple convection and plume models. Whereas plumes are a form of convective instability rising from the base of a convecting system heated from below, the form of convection we are discussing is triggered from above. Unlike other lithospheric mechanisms, the asymmetric lithosphere does not require convective thinning or heating of the plate in order to produce melting. This eliminates time delay between the arrival of the plume head and the onset of volcanism in the stretching model. We consider a series of calculations with a step-function change in thickness of the boundary layer and an externally imposed pull-apart. The flow in our models is shallow and sub-horizontal, and brings hot material from under the thicker (cratonic) boundary layer towards the pull-apart. A simple estimate of the amount of melt generated by this mechanism suggests that it is capable of producing a large igneous province, even for a dry mantle.

1. Introduction

The mechanism of generation of large igneous provinces, such as the Deccan Traps and the Columbia River flood basalts, has puzzled geologists for many years. In the current literature, continental flood basalts (CFB) and other large igneous provinces (LIPs) are related to deep mantle plumes [1]. The idealized plume has two components: a plume head, supposedly responsible for very short-lived, massive igneous events, and a narrow plume tail which generates long-lived hotspot tracks. Mantle plumes are described as local, secondary convective features that are more or less independent of the larger, plate scale flow in the mantle [2]. The source of mantle plumes is a thermal boundary layer deep in the mantle, perhaps the core–mantle boundary [1], although some have argued for a shallower source [3].

The envisioned structure of plumes comes primarily from laboratory tank experiments [4,1], yet plumes are difficult to generate in the laboratory. Artificial plumes are sometimes generated by injection of a narrow jet of low viscosity, superheated fluid into the base of a static tank of fluid [5]. These plumes begin with a large head that assimilates the host fluid as it rises, and a long, thin tail. The head is initially composed of the hottest material from the base of the thermal boundary layer (TBL) but it entrains the overlying material and cools as it rises. As the plume head grows it pushes the overlying material out of
The tail remains hotter, relatively pure D' material. Nataf [6] showed that, in a naturally convecting system, the TBL cannot reach the high temperatures and low viscosities necessary to form the boundary layer instabilities for these features within the convecting system.

2. The plume-head hypothesis and flood basalts

Plume hypotheses attribute massive volcanism to localized, and deep fluid dynamic processes. The interaction of the plume head with the lithosphere is considered to be responsible for short-lived, massive igneous events [7,5]. Yet some LIPs have erupted over a fairly long period of time. The interaction of the hot, buoyant plume head with the lithosphere should also generate uplift and stretching prior to, or concurrent with volcanism [8]. In many large igneous provinces, the evidence for uplift, heating, or stretching is lacking. In fact, some CFBs occur on top of deep sedimentary sequences and in depressions. Cratonic lithosphere adjacent to CFB provinces shows no evidence for heating or thinning [9].

The large plume head, even after it flattens into a pancake shape beneath the lithosphere, should be readily resolved by seismic tomography but, to date, there is no evidence for either plume heads or damaged lithosphere beneath CFB provinces [9,10]. Flood basalts are found on the edges of thick, high-velocity lithosphere that correlates with Archean cratons (Archons) and, when backtracked to their eruption sites, occur over large low-velocity regions of the asthenosphere [11]. In general, these large mantle domains have been insulated by a supercontinent or have not been cooled by subduction since the breakup of Pangea [12,13]. The geoid, topographic, and heat flow anomalies associated with hotspots can be explained by anomalies shallower than 250 km [14] and surface anomalies are not sensitive to the structure of the deep, narrow tail. High-resolution studies of the Yellowstone hotspot show low velocities down to about 200 km depth, about the thickness of the adjacent Wyoming cratonic lithosphere [15,16]. Reports of a deeper anomaly under the Bowie "hotspot" [17] are suspect [18].

One of the important links between CFBs and mantle plumes is the suggestion that many hotspot tracks can be traced back to CFBs [19]. Associations of CFB provinces with specific hotspots often fail when tested with precise plate reconstruction models. For example, the Kerguelen hotspot was 1000 km distant from the Bunbury basalts in Australia and the Rajmahal basalts in India at the time that they formed [20]. On the other hand, Kerguelen appears to have formed close to several Archons, at a time of regional extension and global plate reorganization.

The large North Atlantic igneous province is a region of elevated topography, high-geoid and massive magmatism in east and west Greenland, Iceland, Rejkeyanes Peninsula and the Faeroes. The Greenland basalts are often attributed to the Iceland plume head but there are difficulties with the dimensions, timing and geochemistry [21,22]. In this case, a hotspot track predates the Greenland magmatism [23], being traceable into northern Canada, prior to the "plume-head event" in southern Greenland. The hottest magmas (picrites) occur at some distance from the conjectured plume tail and, in addition, are depleted (MORB-like) compared to the lower temperature, more evolved basalts. Multiple plumes need to be invoked to explain the Baffin Bay and west and east Greenland picrites [22]. The Greenland LIP seem to be related to extension associated with a cratonic lithospheric discontinuity rather than the initial impact site of a plume head.

There are numerous examples of LIPs that have no associated hotspot track (Keweenawan, Karoo, Siberian Traps, Caribbenean, Pacific plateaus, Ethiopia, continental margin basalt sequences) and numerous hotspot tracks that have no associated LIP (Hawaii-Emperor, Line Islands, Cameroon Islands). The LIP on the North American Atlantic margin does not appear to have a hotspot origin [24]. It is a rather remarkable fact that few of the numerous hotspot tracks in the Pacific can be related to the large plateaus (presumed plume-head products), either geometrically or geochemically, or to any other LIP. The large oceanic plateaus formed mostly at new ridges, or triple junctions, or on very young crust [25] and have left no trail. They also formed at times of major plate reorganizations on the edges of a growing Pacific plate [18]. In contrast, several age-progressive seamount chains in the Atlantic (Walvis Ridge) and Indian Oceans (Chagos-Laccadive Ridge,
90°E ridge, 85°E ridge) may be associated with CFBs. These CFBs (Paraná–Etendeka, Deccan, Rajmahal) initiated at Archon boundaries, and pre-existing fractures and the associated tracks were largely built on very young oceanic lithosphere, or at ridges. The initiation and evolution of hotspot features appear, on average, not to be midplate phenomena but strongly favor plate boundaries, lithospheric discontinuities and pre-existing weak areas.

Plumes are thought to have two important implications for the geochemistry of LIPs. First, large volumes of basalt are thought to be a result of hotter than average mantle. High-temperature magmas, such as komatiites and picrites are a more direct manifestation of hot mantle, unless their source regions are wet [26]. Large volumes of basalt can also result from high fertility or high volatile content of the source region, rapid plate extension or small-scale convection. Some CFB provinces are in back-arc basin environments (Columbia River, Patagonia, Antarctic Peninsula and, possibly, Siberian and Keeweenawan CFB) and sites of ridge–trench collisions (central Mexico, Antarctic Peninsula), or above slab-windows. The shallow mantle in these regions has been fluxed with volatiles and sediments, so a high volatile content, and the unusual stress regime could be responsible for the generation of large amounts of melt in these areas. The second geochemical constraint is the “enriched” or primitive component observed in trace element studies of CFBs. It should be noted, however, that sediments and fluids from subduction could also provide the enriched, undepleted, or “primitive” component observed in geochemical studies of CFBs [27], and that these components need not have been recycled deeply into the mantle. The presence of depleted, presumed upper-mantle components, in many hotspot magmas is difficult to reconcile with the plume-head-entainment hypothesis because shallow entrainment in the plume head is specifically excluded by Campbell and Griffiths [5]. The combination of shallow fractionation and contamination (FRAC) can explain many of the geochemical characteristics of CFBs [27].

The growing number of observations requiring modifications to the simple plume model, or a series of special circumstances to fit the plume model into the geological setting of many CFBs, suggests that it is worthwhile considering alternative mechanisms to the strictly fluid dynamical plume models of CFB formation. In fact, the correlation of LIPs to cratons and lithospheric discontinuities is much stronger than the correlation of CFBs and hotspot tracks (see fig. 1 in [11]). Every LIP is adjacent to an Archon or is on an ancient shear zone [11]. Pre-existing fracture zones or sutures are implicated in Deccan, North Atlantic, Ascension, Cape Verde, Columbia River and Paraná volcanism.

3. The alternate model

In this paper, we explore a model where the flood basalt volcanism is controlled, or even initiated, by physical and geometric properties of the lithosphere, primarily the asymmetry of the craton/non-craton boundary and its affect on the surface heat flux and mechanical properties. This model provides a natural explanation for the strong correlation of CFBs and craton boundaries. In our model, the difference in heat flux between the thicker, older, cratonic lithosphere and the thinner, younger, lithosphere drives small-scale convection, drawing mantle from under the deep cratonic lithosphere to the boundary. At the lateral discontinuity, mantle is able to rise freely from below the craton to the depth of the thinner non-craton boundary without requiring large stretching factors. This enables material to cross into the melting zone relatively easily. The small-scale flow drives a large amount of mantle through the melting zone, explaining the large volume of basalt at CFB provinces. Because these regions are often the sites of previous subduction, the shallow mantle may already be volatile rich, giving the magmas an enriched signature without requiring any deep mantle or continental lithosphere component. The rising material may also interact with the continental crust, continental lithosphere and various thermal boundary layers and with the boundary zone which is a natural conduit for previous metasomatism and a site common for sedimentary basins.

We require an external (i.e., not provided by the small-scale flow) pull-apart environment on the craton/non-craton boundary. In contrast to the stretching model of McKenzie and Bickle [28], we do not require stretching to thin the lithosphere and allow mantle from below the lithosphere to melt exten-
sively by adiabatic decompression. Because of the lithospheric asymmetry, this zone is accessible to the fluid on the non-craton boundary side of the boundary. The small-scale flow continuously delivers fresh mantle from beneath the thicker craton to the shallow mantle and the asymmetry of the lithosphere allows the deeper fertile mantle to reach the melt zone without significant lithosphere stretching. The lithosphere on one side of the suture is already thin. This considerably reduces the delay between plate tectonic-induced extension and massive magmatism.

The small-scale flow may also contribute to lithospheric pull-apart but this is not included in our simple model. We are assuming an initially passive mechanism with tensile stresses provided by, for example, slab pull, trench roll-back, slab detachment or ridge–trench annihilation. Other mechanisms, operating at large distances, are slab-induced upwellings [29]. These may be pertinent to the Paraná, Karoo, Ferrar, and Tasman CFBs, which are parallel to Gondwana subduction zones. These mechanisms are all unrelated to thermal instabilities in a deep boundary layer, the conventional source for active upwelling plumes. Our “plumes” also experience adiabatic melting at shallow depths, but they are triggered from above, by conventional plate tectonic forces and lateral temperature gradients, rather than from the bottom of the convecting system. Our mechanism is more closely related to that first discussed by Pekeris [30] than to Rayleigh-Benard convection.

4. Model description

We use the two-dimensional, Cartesian, finite-element convection code ConMan, described in King et al. [31]. The calculations are performed in a 2-by-1 box with a free-slip bottom and imposed, or free-slip velocities at the surface. The depth of the box is scaled to be 600 km. The velocities along the sides of the box are imposed, using the solution for viscous flow under a uniform velocity plate (Couette flow). A grid of 120-by-60 uniformly-spaced elements is used. The temperature boundary condition along the top is \( \theta = 0.0 \) and along the bottom is \( \theta = 1.0 \). The viscosity of the plates is 100 times the background viscosity and the initial viscosity structure does not evolve with the flow. The thinner (oceanic) plate is on the right-hand side of the box \( (1.0 < X < 2.0) \) and is always 50 km thick. The thicker (continental) plate is on the left-hand side of the box \( (0.0 < X < 1.0) \) and its thickness varies from 50 to 200 km. The thermal and chemical properties of the plates are identical to the background fluid, except for the viscosities. Because the isotherms do not evolve far from the initial configuration, the constant viscosity structure that we use to model a more realistic temperature-dependent solution seems justified.

Because the relevant length scale in the problem is the thickness of the upper thermal boundary layer (TBL), we choose this as the depth scale for the Rayleigh number:

\[
R_a = \frac{\rho g \alpha \Delta T (\delta)^2}{\kappa \eta_{\text{plate}}}
\]

where \( \rho \) is the density, \( g \) is the acceleration due to gravity, \( \alpha \) is the coefficient of thermal expansion, \( \Delta T \) is the temperature drop across the TBL, \( \kappa \) is the thermal diffusivity, \( \eta_{\text{plate}} \) is the viscosity of the boundary layer, and \( \delta \) is the thickness of the TBL (we choose the 50 km TBL to scale our problem). This boundary layer Rayleigh number represents the small-scale flow induced by the TBL, not the Rayleigh number of the mantle flow. Note that if the boundary-layer Rayleigh number exceeded the critical value, the boundary layer becomes unstable and sinks into the mantle, forming a drip.

To estimate the region of partial melt, we follow a simple parameterized melting formulation [28]. In this formulation, the pressure and temperature at a given point are compared with an analytic function representing the solidus and liquidus as a function of pressure and temperature for garnet peridotite (Fig. 1a). The fraction of melt is calculated by fitting a polynomial to the experimental data (see [28], eqs. 18–22 for details; Fig. 1b). While this formulation ignores the feedback between the latent heat of melting and the temperature field, it gives a rough estimate of the region of melting. It does not take into account the volatile content of the underlying mantle on the melting curves. As we have argued above, the regions we are considering may have high concentrations of volatiles. Because the effect of
volatiles is to reduce the solidus and liquidus temperatures, our melting results represent a lower bound. Prior to the extraction of the CFB, the mantle may be more fertile than mantle which has not experienced previous hotspot or ridge magmatism.

5. Numerical results

There are two competing effects in the calculations: pull-apart of the plates, and flow driven by the horizontal variation in the surface heat flux due to the difference in the thickness of the thermal boundary layers. Fig. 2 illustrates the imposed pull-apart problem. In this case the Rayleigh number is zero and the solution is the steady-state response of the viscous fluid to the imposed velocity boundary condition at the surface. A velocity of $-1.0$ is used for $0.0 < X < 1.0$ and $1.0$ for $1.0 < X < 2.0$. The thickness of the left-side plate is varied from 50 to 200 km while the thickness of the right-side plate is held constant at 50 km. Fig. 2A is essentially a well known corner flow solution with a plane of symmetry at $X = 1.0$. Once the symmetry is broken (Fig. 2B–D) by increasing the thickness of the "craton" plate, the flow of material toward the boundary between the two plates (the center of the spreading) is primarily from underneath the thinner plate. In the case of equal-thickness lithospheres, the passive response is well studied; however, when the thick-
Fig. 3. Flow driven by non-uniform heat flux due to a 200 km/50 km lithosphere (Fig. 2D) with no imposed surface pull-apart boundary conditions. The shading represents the zone of partial melt, the darkest shade is 2% melt. The boundary layer Rayleigh number (based on the 50 km thick lithosphere) is (A) 0.06; (B) 0.6; and (C) 6.0.

nesses are not equal, the flow is no longer symmetric and the passive upwelling is shifted and the shallow (oceanic or non-cratonic) mantle is preferentially sampled.

Fig. 3 illustrates the thermal convection problem with no imposed pull-apart. In this case the plates are held constant (200 km thick on the left-hand side and 50 km thick on the right-hand side) and the boundary layer Rayleigh number is increased from 0.06 to 6.0. In this case, we use reflecting side-wall boundary conditions; however, experiments with wider aspect ratios and periodic boundary conditions yielded identical conclusions. The calculations evolve from an initial step function increase in the thermal boundary layer thickness at the boundary of the two plates to the state shown in Fig. 3. In these calculations, the flow is driven by the lateral variation in heat flux. Material from underneath the thicker plate flows laterally, in the shallow mantle, toward the thinner plate. The deep subcratonic mantle is sampled. In the three cases, the maximum velocity of the flow is different, but the arrows in each plot are scaled differently to illustrate the pattern of the flow. The maximum dimensionless velocity for each of the flows is Fig. 3A ($Ra = 0.06$) 5.3; Fig. 3B ($Ra = 0.6$) 300.0; Fig. 3C ($Ra = 6.0$) 650.0. In dimensional terms, using $d = 600$ km and diffusivity of $10^{-6}$, these are 0.03, 1.7, and 3.2 cm/yr, respectively. In contrast to current plume theories, the amount of upwelling and melting depends on the thickness contrast between the two plates, the rate of plate separation, and the temperature.

Fig. 3 illustrates the effect of the thickness contrast between the two plates on the amount of melt produced. The dark region outlines the zone of greater than 2% partial melt using the formulation of McKenzie and Bickle [28]. In cases A and B, the boundary-layer Rayleigh number is so low that the oceanic plate significantly thickens by diffusive cooling. The thick oceanic plate keeps the potential temperature in the depth where melting could occur below the minimum temperature required to induce melting. The amount of cooling in cases A and B are unrealistically high for the Earth; the oceanic plate grows from 50 to 100 km thick in 30 Myr. However, even in this case the asymmetry in the lithospheric heat-flow still drives small-scale flow. In case C, the boundary layer is not thickening significantly over the length of this calculation, and a significant zone of melt extends almost across the entire length of the thin plate. Without pull-apart, however, there is no mechanism for this melt to reach the surface. The same is true, of course, for plume models.

In Fig. 4, the imposed extension and thermally driven flows are considered together. In these cases, the plates are held constant (200 km thick on the left-hand side and 50 km thick on the right-hand side), and the boundary-layer Rayleigh number is fixed at 0.6 and the velocity of the imposed pull-apart is varied from 100 to 10. Note in Fig. 3 the velocity of the flow driven by the variation in the thermal boundary layer is 300, so the imposed extension is
Fig. 4. The pull-apart surface boundary conditions and the small-scale heat-flux flow for the 200 km/50 km lithosphere model with a boundary layer Rayleigh number of 57.8 (compare with Fig. 3B). The half spreading rate is (A) 100 units (6 mm/yr); (B) 50 units (3 mm/yr); (C) 25 units (1.5 mm/yr); and (D) 10 units (0.6 mm/yr).

always less than the convective flow. In the case where the pull-apart velocity is comparable to the convective flow velocity (Fig. 3A), the flow comes from deep in the box and slightly favors the thicker plate. For cases where the pull-apart velocity is smaller, a major portion of the flow comes from beneath the thicker plate.

In case 3B, with no imposed pull-apart, there was no melt zone. In case 4A, B, and C, there is small zone of partial melt concentrated beneath the pull-apart boundary. In the case where the external pull-apart was 1/3 of the small-scale flow (Fig. 4A), the zone of partial melting is comparable to the size of a large igneous province (50 × 100 km). In all three models, this mechanism is capable of generating a 50 km by 2 km thick region of melt (assuming 2% average partial melt and complete withdrawal). This is all accomplished with a dry mantle and no excess temperature. Both the presence of volatiles and anomalously warm mantle would increase the amount of melt produced by this mechanism. Also, it is important to keep in mind that this mechanism is continuously fluxing previously unmelted mantle into this zone.

6. Discussion

Although our calculations are for a simplified situation, we stress that variable thermal and mechanical properties (i.e., cratons and ocean plates) at the surface are more realistic than the uniform properties treated by most investigators. Our model is still highly simplistic; in the Earth, not only are the lithospheres of cratonic and non-cratonic provinces of different thicknesses, but the physical properties (strength, thermal conductivity, radioactivity) differ as well. The resulting lateral gradients in temperature and heat flow must drive small-scale convection. Our modeling suggests that it is the large-scale change in heat flux from the thicker craton to the thinner non-craton that drives the small-scale flow in this problem, not the smaller-scale temperature gradient at the asymmetry discontinuity. Lateral temperature gradients in the shallow mantle can exceed vertical temperature gradients in thermal boundary layers, and no critical Rayleigh number need be exceeded before the onset of small-scale convection [32].

Mutter et al. [33] pointed out that large-volume igneous provinces form where the transition from thick to thin lithosphere is abrupt, setting up strong lateral temperature gradients which induce small-scale convection and rapid movement of material...
through the melting zone. Volcanic margins occur at such transitions [24]. Uplift and volcanism along the margins of the North Atlantic have been attributed to the effects of hot asthenosphere juxtaposed next to deep cratonic lithosphere [24,34]. Our calculations support these suggestions.

Our model suggests that this small-scale flow will be present, at some level, at every craton/non-craton boundary (Fig. 3). The Bermuda rise has been suggested to be a result of such small-scale flow at the boundary of the North American continent and the old Atlantic Ocean [35]. Some of the eastern Atlantic hotspots are at similar distances from Europe and Africa and may be similarly explained. In addition, these hotspots were ridge-centered prior to westward migration of the mid-Atlantic ridge, and may have been initially induced by passive plate divergence. Some of the hotspots in the central Atlantic may be related to the convergence of the African and Eurasian plates.

One of the most immediate questions our model raises is how this initial asymmetry forms. It is important to recognize that extension is necessary to focus the flow at the suture between the craton and non-craton (Figs. 3 and 4). This means that not every craton boundary will be the site of an LIP. We note that LIPs often seem to coincide with the time of plate reorganization. The Ontong–Java plateau, Falkland plateau, Etendeka CFB, Manihiki plateau, and Marcus Wake and Wallaby plateaus occurred between 115 and 125 Ma, a time of continental breakup, rapid plate motions, plate reorganization and plutonic activity in eastern and western North America and opening of the Canada (Arctic) basin. The period 80–90 Ma was also one of global plate reorganization. Associated with it was further activity on Ontong–Java, Kerguelen, Mid-Pacific mountains, Cruiser plateau, Line Islands, Reunion, Caribbean, Gorgona and Madagascar. Other examples are given in Anderson [36].

One can speculate that the tectonic plates are normally under compression, but large-scale plate reorganizations can cause transient extension at plate boundaries, fracture zones, and pre-existing sutures and mobile belts. Most LIPs, including ridges and triple junctions, in fact, occur at such boundaries. More local effects, such as ridge–trench annihilation and slab-window formation are also implicated in some CFB provinces. The reorganization may well be important not only in providing the external pull-apart force, but in juxtaposing craton and non-craton, or thick and thin lithosphere. The reorganization could also change a subducting environment, which has recently fluxed the upper mantle with volatiles, to a passive or extensional environment, setting the stage for this CFB mechanism. The southern margin of Gondwana, for example, was a long-lived subduction zone before the parallel belt of CFBs (Paraná, Etendeka, Karoo, Tasman, Ferrar) formed. The western margins of North and South America were subduction zones for a long time prior to the formation of Antarctic peninsula, Patagonia, Paraná, Central Mexican, Basin and Range, Columbia River and Crescent basalts. The Keweenawan and Siberian basalts also formed in a convergent environment. Insofar as the plateaus in the Pacific formed at or near ridges or triple junctions, they were in extensional environments. Tomography shows that they, as well as all the other Mesozoic LIPs, formed when their underlying lithosphere was moving over what are now vast low-velocity (hot) regions of the upper mantle [36].

An important assumption of our imposed pull-apart is that stresses are concentrated at the asymmetric lithosphere boundary. Because this boundary is a chemical/mechanical boundary, or a pre-existing suture, it seems reasonable to assume that this might be a region of weakness or, at least, of stress concentration. Thus, rather than a broad stretching, the deformation will be localized. The phenomenon of reactivated lithospheric fractures is well known. For example, the Deccan, Paraná and Columbia River basalts all occur along reactivated ancient sutures or fractures.

7. Conclusions

Two contrasting plume models have been proposed for the development of Large Igneous Provinces (LIPs):

1. The impact, on the base of the lithosphere, of a transient broad plume head originating at the core–mantle boundary [7,1].

2. Lithosphere extension over a pre-existing steady-state mantle plume which has developed a large thermal head [3,37].
In model 1, the plume head provides the force for plate rupture and the material for LIP. A deep source is required in order for the plume head to reach 2000 km in diameter, the presumed radius of influence of the plume. In model 2, it is stretching and thinning of the lithosphere that triggers magmatism.

The plume impaction model [5] offers an inadequate explanation of how voluminous melt can be produced in such a short time (<10 Myr) below lithosphere that was initially 150 km thick. It appears that thin lithosphere must be available shortly after melts from under thick lithosphere are delivered to the surface, and no plume or lithosphere stretching or erosion model is adequate [37]. Our model addresses and apparently resolves these problems. A range of depths, source location and melting column lengths is available, and magmatism is transient and time progressive rather than steady. Although the continental lithosphere is an important element in our model, its role is completely different from that proposed by geochemists who envisage early melts coming from the subcontinental lithosphere and later melts from the asthenosphere (e.g., [38]). In some models (e.g., [39]), the cratonic lithosphere focuses the upwelling. In our model, the cratonic root is an essential, rather than an accidental element, and it is not an accident that CFB provinces occur next to Archons, the thickest lithospheres on Earth.

One mechanism for generating melt in previously unmelted mantle is to induce stretching and thinning of the lithosphere until asthenosphere is brought across the solidus. White and McKenzie [3] argued that melting from normal mantle is inadequate to form CFB provinces and, therefore, they imposed a hot plume so that melting can commence at a greater depth. They assumed that temperature and stretching are the only parameters and that high temperatures require a plume from the deep mantle. They ignored the contributions that water content and fertility may make to magma production volumes and ignore small-scale convection. They assumed that the mantle is chemically homogeneous. Their mechanism cannot explain the rapid turn-on and turn-off of magmatism nor can it explain the thick volcanic margins that are unrelated to hotspots [24,40]. Our mechanism is consistent with these observations and, furthermore, does not require abnormally hot mantle, although it does not preclude it. Our mechanism is turned on and off by the local stress regime. During extensional periods, LIP building is possible, during compressional periods, it is not. Hot plumes are only required to explain LIPs if temperature is treated as the only parameter controlling melt volume. In addition, asymmetric lithosphere provides a natural method for bringing up deeper and hotter material and provides the space for melting.

Combining plate reconstructions with tomography we believe that most LIP did form over hot regions of the mantle but we envisage these as vast domains, uncooled by cold slabs, rather than areaally restricted plume heads bringing heat up from the lower mantle. Convective cells, or domains, in the upper mantle can differ in temperature, volatile content and fertility depending on their previous history of subduction, continental insulation and melt extraction. We attribute the variety of basalt types, degree of evolution, and the time sequence of eruption, to the various depths swept out by the upwellings at the lithospheric asymmetry boundary, rather than to entrainment by plumes from deep in the mantle, or structure inside a plume head. The mechanism we propose will certainly be more effective over hot mantle because viscosity decreases exponentially with temperature and because less adiabatic ascent is required for melting. The same is true for wet mantle (e.g., low melting temperature, low viscosity) or fertile mantle (high basalt content). Continents experience uplift as they drift over hot regions of the mantle or geoid highs. Differential uplift, and reactivation of suture zones and mobile belts can occur as variable thickness lithosphere drifts about on a lateral inhomogeneous mantle. In addition, changing boundary conditions at the edges of plates can also change stress conditions in plate interiors. These are alternate considerations to long-lived (>150 Myr) plume incubation periods [41], or fossil plume heads [42]. The main points are: the upper mantle is not isothermal or uniform in composition and history, plates are not uniform in thickness and deep upwelling plumes are not the only source of stress, uplift, or temperature variation.

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