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Continental flood basalts: episodic magmatism above long-lived hotspots

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Abstract

The eruption of continental flood basalt (CFB) may reflect episodic magmatism above long-lived mantle plumes. The Iceland and Yellowstone hotspots have generated successive CFB provinces, large intrusive complexes, anomalous uplift, basin formation, and rifting events, and linear volcanic chains dating back >120 and >70 Ma, respectively. Amagmatic intervals occurred: (1) when ascent of plumes to shallow levels was impeded by impenetrable lithosphere, resulting in sub-lithospheric ponding of plume mantle; and (2) in response to dispersion by subducting oceanic lithosphere in convergent margin settings. By comparison with the plume eruptive potential of typical oceanic hotspots, it is apparent that preservation of only a small portion of plume mantle ponded during an amagmatic interval is necessary to account for large volume of CFBs. Thermal erosion, lithospheric attenuation, translation of ponded hotspot mantle to the base of thinner penetrable lithosphere, and passage of plume mantle through slab windows in subducting oceanic lithosphere led to subsequent breakthrough and eruption of CFB. Since both mantle plume and plate tectonic processes have been operating since the Archean, it seems likely that the migration of continents over hotspots, with attendant magmatic and tectonic consequences, is a common occurrence in the geological record. © 2000 Elsevier Science B.V. All rights reserved.

Keywords: flood basalts; mantle plumes; magmatism; hot spots

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1. Introduction

Continental flood basalts (CFB) result from short-lived (~ 2 My), highly productive magmatic events (eruptive rates of 0.1 to >1.0 km^3/yr) and are characterized by basalts derived, in part, from an anomalously hot, enriched mantle reservoir [1]. CFB magmatism has been attributed to the ascent of starting mantle plume heads from near the core–mantle boundary [2–4]. This model is consistent with experimental studies suggesting that initiation of a mantle plume requires the development of a large (diameter >400 km and probably ~ 1000 km) bulbous head – the starting plume head – in order to generate enough buoyancy to displace the overlying mantle [2,4]. Steady-state hotspots (e.g. Hawaii) are considered

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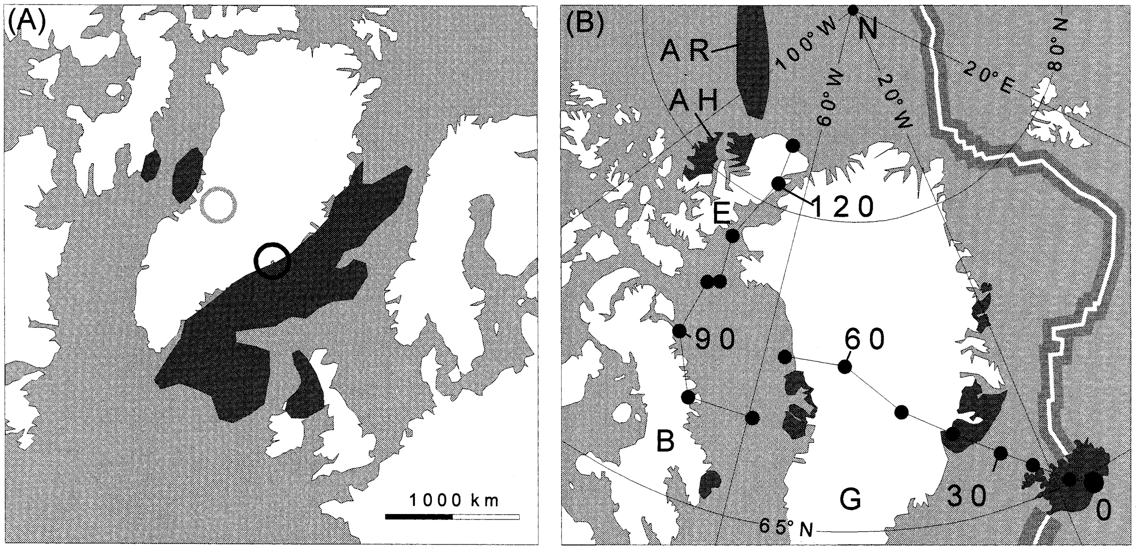

Fig. 1. (A) Paleogeography of the northern Atlantic at about 63 Ma showing the distribution of magmatic rocks forming the North Atlantic Igneous Province, modified from White and McKenzie [1]. Also indicated are the interpreted sites of suggested plume head impact of Saunders et al. [6] (grey circle) and White and McKenzie [1] (black circle). (B) The trajectory of the Iceland hotspot back to 130 Ma, modified from Müller et al. [8] and Lawver and Müller [9]. Dots indicate the hotspot location in 10 Ma increments (numbered every 30 Ma). Breaks in track indicate post-hotspot rifting. Active mid-oceanic ridge is indicated by white line. Dark grey indicates CFB and hotspot basalts attributable to the Iceland hotspot. G, Greenland; B, Baffin Island; E, Ellesmere Island; AH, Axel Heiberg Island; AR, Alpha Ridge.

to be the result of plume 'tails' which continue to rise up along the conduits established during ascent of the starting heads [2,5]. An alternative hypothesis is that continental flood basalts result from ponding of slowly accumulated hot plume material beneath the lithosphere [1], with only subsequent break-up of the overlying lithosphere allowing for voluminous melt production.

In this paper, we present two examples of CFB provinces attributable to long-lived plumes rather than by starting mantle plume heads. To illustrate this view, the Northern Atlantic and Columbia River CFB provinces, commonly cited as products of starting plume heads, are re-examined. In both cases, the plumes responsible for these basaltic outpourings appear to have been active for tens of millions of years prior to CFB magmatism. A model of migration of continents over pre-existing plumes, and of ponding of plume mantle and episodic lithospheric breakthrough, more plausibly accounts for the protracted histories of continental plume-related magmatic activity.

2. Northern Atlantic igneous province

Flood basalts on Baffin Island, western and eastern Greenland, the northern coast of Scandinavia, and northwestern Great Britain (Fig. 1) were previously attributed to the arrival of a starting plume head at 63 Ma [2,6,7], prior to rifting of the northern Atlantic and the Labrador Sea. In this interpretation, the plume head rose beneath Greenland [1,6] and, on encountering the lithosphere, flattened into a disc 1200 km across (Fig. 1). Anomalously hot mantle continued to rise as a plume tail, producing the Iceland hotspot track on the sea floor created during subsequent rifting. Explanation of the flood basalt province as the product of continental break-up above a steady-state hotspot that rose beneath east Greenland [1] was rejected on the grounds that the volume of magma produced was too great to be the product of a steady-state plume beneath a stationary plate [2].

The starting mantle plume head model is called into question by plate kinematic studies [8,9] which have established the plate motion relative to a fixed Iceland hotspot back to 130 Ma, and by the prolonged history of crustal thinning and rifting present along the hotspot track. The geometrically determined Iceland hotspot track (Fig. 1) coincides spatially and temporally with the distribution of mantle plume-derived volcanic rocks, and with areas affected by anomalous uplift [9]. The hotspot is currently beneath eastern Iceland. Between 50 and 40 Ma, eastern Greenland passed over the hotspot during which time the Kangerlussuaq, Skaergaard and Kap Deichman intrusive complexes were emplaced. 2–6 km of denudation of the coast in and south of the Kangerlussuaq area occurred at this time, consistent with the development of a thermally supported topographic high above a hotspot [10,11]. Between 60 and 50 Ma when the hotspot was beneath central Greenland, voluminous flood basalts were erupted in eastern Greenland. The volume of high-temperature picrites in this succession is small [12], consistent with magmatism above the distal portion of hotspot-influenced mantle. Between 70 and 60 Ma when the hotspot was beneath western Greenland and Baffin Bay, voluminous basaltic flows, including large volumes of picrite [12,13], were erupted immediately above the hotspot. Rapid uplift, giving rise to regional unconformities, immediately preceded volcanism in western Greenland [14] and in eastern Greenland and Scotland [14,15].

The hotspot track prior to 80 Ma is poorly constrained [8,9]. A thick sequence of $\sim$100 Ma basalt flows on northern Ellesmere Island (the Strand Fiord formation) appear to have erupted above the hotspot. Interpretation of these basalts as being attributable to a hotspot is consistent with the intra-plate setting of the basalt flows; their continental tholeiitic chemistry (subalkaline with relatively low MgO); and their large volume (up to 789 m thick and covering much of northern Axel Heiberg Island) and high eruptive rates (giving rise to basalt flows >60 m thick) [16]. In addition, the volcanic sequence is spatially associated with, and lies near the focal point of a coeval radial dyke swarm ([17]; R.E. Ernst, personal communication). Prior to 120 Ma the hotspot lay north of Ellesmere Island beneath the oceanic Canada Basin, and may have formed the Alpha Ridge, a linear volcanic edifice that terminates against Ellesmere Island [9].

The slow southward migration of the Iceland hotspot into the Baffin Bay, Davis Strait, and finally Greenland area resulted in gradual thermal erosion of the overriding lithosphere. This thermal erosion is recorded as intermittent thinning and rifting of the lithosphere beginning in the Baffin Bay region prior to 99 Ma, the end of the Early Cretaceous. Immature Early Cretaceous clastic sediments attributable to rifting crop out along the margins of the Davis Strait and Baffin Bay regions [18,19]. Onshore unconformities that have been related to rift geometries mapped on seismic sections in Baffin Bay indicate rifting in the Lower and mid-Cretaceous, as well as younger base Tertiary and mid-Tertiary events [20–22]. Magnetic anomalies indicate movement of Greenland away from Baffin island as early as anomaly 34 time (pre-83 Ma, the end of chron 34), the oldest identifiable anomaly in the Labrador Sea [22]. Long duration uplift, giving rise to the anomalously high elevations of eastern Baffin Island, may be a consequence of plume-related lithospheric thinning.

The rapid uplift prior to volcanism in western and eastern Greenland has been interpreted as providing a record of the instantaneous emplacement of a starting mantle plume head at the base of the lithosphere [14,15], or of a plume head rising along the conduit of pre-existing plume following a period of weak or non-existent plume activity [6]. However, alternative interpretations involving the presence of a steady-state plume are permissible. Rapid uplift events may simply mark the propagation of active rifts into regions already overlying ponded hot plume mantle (cf. [1]). Synchronous migration of crust in these areas over the plume axis may have further enhanced uplift of the overlying crust. Finally, rapid lateral migration of ponded plume material, perhaps in response to overflow of ‘topographic’ features on the base of the lithosphere (cf. [23]) may have given rise to the uplift events. In all three cases, rapid uplift is attributable to lithospheric causes; arrival of a starting mantle plume head is not required.

The Iceland hotspot was apparently active for at least 70 My prior to the Tertiary eruption of CFB of the North Atlantic igneous province. CFB magmatism and uplift along the hotspot track resulted from the migration of the Laurentian continent over a steady-state hotspot which previously was beneath the Arctic basin. Heating and thinning of continental lithosphere above the hotspot probably facilitated subsequent rifting and rupturing of Laurentia.

3. Columbia River flood basalts

The Columbia River flood basalts have been attributed to a starting plume head centered near the Nevada–Oregon–Idaho border at 17 Ma [24] (Fig. 2). On impact with the lithosphere, this hypothetical plume head flattened into a disc >1000 km across. 175 000 km^3 of flood basalt and minor rhyolite erupted above the northern half of the plume head, while Basin and Range extension occurred above the southern half [24,25]. Southwestward motion of North America of 3 cm/yr in the hotspot reference frame places the present plume tail beneath Yellowstone, Wyoming, an area characterized by elevated heat flow, recent magmatism, anomalously high elevations, and active faulting and uplift [24,25].

We argue that the Columbia River flood basalt province was not generated by a Miocene starting plume head, but rather by sub-crustal accumulation and subsequent break-through of basalt generated above a long-lived steady-state hotspot. Supporting this view are plate motion models [8,26,27] of North America relative to a fixed Yellowstone hotspot back to 130 Ma [26,27]. These plate motion models result in similar hotspot tracks, although in detail they differ in the exact time and latitude at which the Yellowstone hotspot crossed beneath North America. In Fig. 2 we employ the plate motion model of Johnston et al. [27], a modified version of the plate motion model proposed by Engebretson et al. [26]. This model, like the others, places the hotspot south of the main body of the Columbia River flood basalts at 17 Ma, indicating northward deflection of the plume basalt.

Northward deflection of the plume was prob-

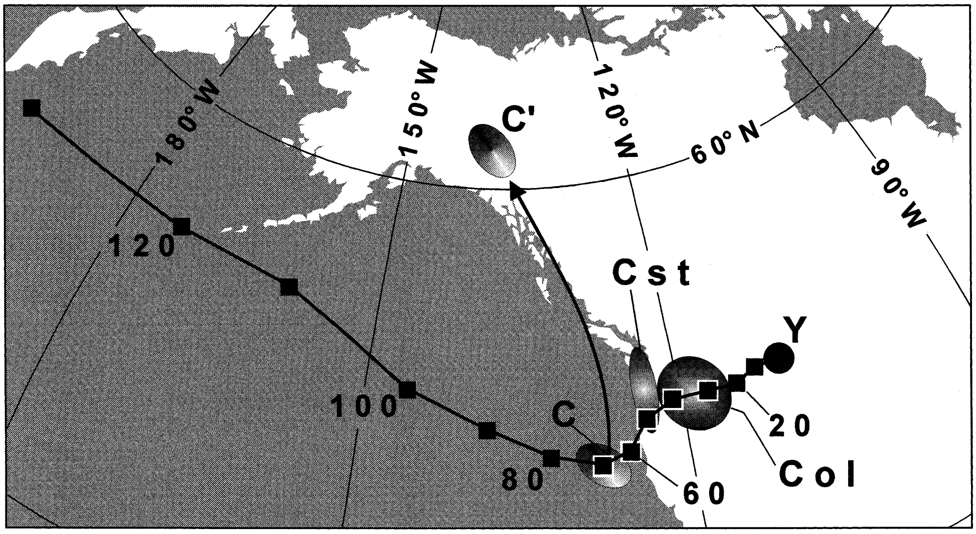

Fig. 2. Yellowstone hotspot trajectory back to 130 Ma (modified from [11]). Small squares indicate the paleolocations of the hotspot [26,27] in 10 Ma increments (numbered every 20 Ma). Patterned areas indicate CFB attributable to the Yellowstone hotspot, including the Columbia River (Col), and Coast Range (Cst) basalts, and the Carmacks Group (C, paleomagnetically determined paleolocation at 70 Ma; and C′, current location). Arrow lies along the C–C′ great circle and indicates approximate post-depositional coastwise displacement path of the Carmacks Group. Translation resulted from coupling with the Kula plate [26].

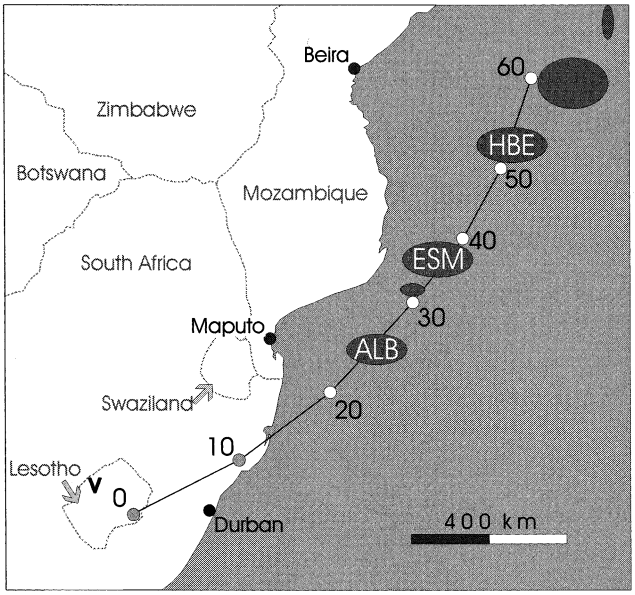

Fig. 3. Interpreted Quathlamba hotspot trajectory back to 60 Ma, modified after Hartnady [47]. Small white and grey circles indicate paleolocations of the hotspot in 10 Ma increments. Dark grey ellipses schematically indicate seamounts attributable to the hotspot including the Hall bank–Bassas da India–Europa Island (HBE), Eggert (ESM), and Almirante Leite Bank (ALB). **V** indicates the location of a 1983 basalt eruption in Lesotho.

ably a result of interaction with the subducting Farallon plate [28]. Migration of the leading edge of the Farallon–Juan de Fuca plate to a position west of the plume enabled the plume to rise in a normal, more vertical orientation, giving rise to volcanic rocks of the Snake River plain [28]. Deflection of plume mantle was also probably in part the product of ponding against, and lateral flow along the north-trending, western margin of thick Precambrian lithosphere of the North American craton [29]. Northward lateral flow of the ponded plume mantle gave rise to coeval, northward migrating uplift of the overlying crust [29].

The CRB were preceded by a 33 My amagmatic interval during which the plume lay beneath the continental lithosphere. Immature Eocene and Miocene clastic sedimentary rocks of the Pasco Basin [30] are preserved beneath the CRB and provide a record of crustal thinning and basin formation above the Yellowstone plume during this amagmatic interval.

At 50 Ma, the Yellowstone hotspot was located beneath the southern Oregon coast. The Kula–Farallon ridge intersected the convergent margin near the hotspot at this time, and the resulting slab window may have focused upwelling plume mantle [31]. The voluminous Coast Range basalt, erupted near the hotspot at this time, consists of accreted paleo-seamounts [32] and basalts erupted in rifts along the continental margin [31]. The Coast Range basalt includes the Crescent Formation, a >16 km thick accumulation of basalts [33] exposed on the Olympic Peninsula, incorporating ocean-island tholeiites derived from anomalously hot enriched mantle [31]. The volume, chemistry, tectonic setting and location of the Coast Range assemblage suggest magmatism above the Yellowstone hotspot [31]. The coeval Metchosin Igneous Complex of southern Vancouver Island, a mafic assemblage of gabbros, sheeted dykes, and submarine and subaerial basalt [34], is a more northerly portion of this igneous province. The presence of subaerial sequences within these juvenile, largely oceanic successions [31,34] may be the result of syn-magmatic thermal inflation of the lithosphere.

Basalts of the 70 Ma Carmacks Group in central southwest Yukon [35] are an earlier manifestation of the Yellowstone plume. This voluminous CFB sequence includes potassic basalts with 15 wt% MgO. The volume and composition of the basalts requires anomalously high liquidus temperatures (1400°C at 1 bar, dry) consistent with a hotspot origin [27]. Magmatism was associated with dextral strike-slip faulting, and with epi- and meso-thermal gold mineralization and hydrothermal alteration. Pre- to syn-magmatic uplift is indicated by deposition of the volcanic succession over a deeply incised topographic surface; by the presence of thick, coarse, locally derived conglomeratic sequences interbedded with the lowermost volcanic layers; and by basal breccias interpreted as paleo-landslide deposits [35,36]. Paleomagnetic data indicate eruption 1900 ± 700 km south of their current location, near the 70 Ma position of the Yellowstone hotspot [27,37] (Fig. 2). The coincident paleolocations of the Carmacks Group and the Yellowstone plume are remarkable. As with the Coast Range basalt, the Kula–Farallon ridge intersected the convergent margin near the hotspot at this time and probably served to focus upwelling of plume mantle [38]. Subsequent northward displacement of the crust on which the Carmacks Group was emplaced resulted from coupling with the adjacent, rapidly north-moving Kula plate [27,39]. Pre-Carmacks Group Yellowstone hotspot-related volcanic rocks erupted onto oceanic crust of the Pacific basin would have soon entered into nearby subduction zones dipping beneath North America. These young, buoyant and hard-to-subduct edifices may have accreted and been displaced along the margin. Identification of these older portions of the hotspot track awaits further research.

In summary, hotspot-related volcanic successions indicate 70 My of Yellowstone hotspot activity. Continental magmatism along the hotspot track, including the Columbia River flood basalt, resulted from the westward migration of North America over the Yellowstone hotspot, a pre-existing steady-state plume which was previously located within the Pacific basin. Like the Iceland hotspot track, magmatism above the Yellowstone hotspot appears to have been episodic; a 30 My amagmatic interval separates eruption of the Coast Range basalts and Columbia River flood basalts.

4. Discussion

Starting mantle plume heads are popularly associated with the eruption of CFB and the onset of hotspot activity. However, in two important cases CFB resulted from mantle plumes that had already been active for tens of millions of years. The North Atlantic igneous province has been attributed to a starting plume head at 63 Ma (Iceland), and the Columbia River flood basalts to a starting plume (Yellowstone) at 17 Ma. In each case, a discontinuous igneous welt extends away from the CFB province along plate motion trajectories toward older projected positions of the mantle plume. We suggest that these welts are continental hotspot tracks. Magmatic activity above the Iceland and Yellowstone mantle plumes extends back for at least 120 Ma and 70 Ma, respectively. CFB magmatism may thus be considered as a punctuated response to the migration of lithosphere of varying thickness and age across pre-existing steady-state hotspots.

Unlike the relatively constant eruption of basalt that characterizes hotspots above mantle plumes that rise beneath oceanic lithosphere, the continental portions of the Iceland and Yellowstone plume tracks are characterized by episodic magmatism: short-lived voluminous magmatic events are separated by extended amagmatic intervals. The Iceland hotspot track includes prolonged amagmatic intervals from 120 Ma to 105 Ma, and 100 Ma to 70 Ma, corresponding with the migration of thick Archean continental lithosphere of Greenland and the Canadian Archipelago over the mantle plume. Magmatic events coincided with migration of oceanic lithosphere and thinned, rifting continental lithosphere over the plume. Amagmatic intervals above the Yellowstone hotspot occurred from 70 Ma to 55 Ma, and from 50 to 17 Ma, and were in part attributable to dispersion, deflection and cooling of the ascending plume mantle by the subducting slab of the Farallon plate. Magmatism occurred in environments progressing from an accreted terrane undergoing strike-slip deformation, to oceanic and transitional crust of the North American continental margin, to continental lithosphere of increasing maturity. The spatial association of the Kula–Farallon ridge–trench intersection with the Yellowstone hotspot during Carmacks and Coast Range magmatism suggests that mantle upwelling may have been focused by the resulting slab windows. The modest amounts of post-15 Ma Yellowstone hotspot magmatism, restricted to small volumes of silicic and basaltic volcanic rocks tracking toward the current hotspot location in Wyoming, reflect the easterly-increasing thickness and stability of the North American lithosphere.

To summarize, the episodic nature of the continental portions of the Iceland and Yellowstone hotspot tracks is attributable to the migration of thick impenetrable Archean lithosphere over the plumes; and, in the case of the Yellowstone hotspot, to dispersion of, and interference with, the plume mantle by subducting slabs. Magmatic events correspond with the migration of ‘thinspots’ [40], such as attenuated or thinning lithosphere, over the plumes, thereby permitting plume asthenosphere to rise to sufficiently shallow levels to permit decompressive melting [23].

The large volumes of rapidly erupted basalt present in the North Atlantic igneous province and the Columbia River flood basalt requires the partial melting of very large volumes of hot plume mantle [24,41]. The necessity of such large volumes of plume mantle has been the most significant factor in interpretation of these large igneous provinces as resulting from starting mantle plume heads [2,24], or, in the case of the North Atlantic Igneous Complex, from a pulsing plume [6]. However, a model of steady-state upwelling and ponding of plume mantle beneath a continent is sufficient to account for the volumes of these igneous provinces. In this model, the steady-state ascent of plume mantle continues unabated during the migration of continental lithosphere over a mantle plume. Ponding of the hot plume mantle beneath thick lithosphere gives rise to sub-continental reservoirs of anomalously hot plume mantle – mantle that retains the potential, given the opportunity to rise to shallow enough levels, to melt in appreciable volumes.

To illustrate this point, we define *plume eruptive potential* (PEP) as the lava erupted above a plume when it rises beneath oceanic lithosphere, i.e. 30 000–95 000 km^3/My for a typical oceanic hotspot [4]. Where continental lithosphere overrides a plume, its lava production will drop below its PEP because melt production will be restricted to greater depths, thereby reducing the volume of melt produced. The increased depth of melting is expected because plume ascent will be inhibited by the reduced density contrast between plume and continental lithosphere, and because continental lithosphere is typically thicker than oceanic lithosphere. The low density and greater thickness of continental crust will further impede the ascent of plume-related magma. The reduction in lava production may be partially offset by increased crustal contribution through anatexis and assimilation. In addition to a reduction in melt production, it is unlikely that all the hot plume mantle that ascends beneath a continent during an amagmatic interval will remain available for melting; some will cool and sink, and some will be left behind by movement of the overriding plate. Subduction of slabs, as in the case of the Yellowstone plume, is also likely to confound the ascent and ponding of plume mantle. Thus, subcontinental plume mantle is expected to be characterized by decreased magma production, and ponded reservoirs of plume mantle to be continually eroded by thermal and physical processes. Despite these factors, preservation of only a small portion of the plume mantle ponded during an amagmatic interval is necessary to explain large volume igneous provinces.

For example, eruption of the Columbia River flood basalts terminated a largely amagmatic 33 My interval above the Yellowstone mantle plume, extending from 50 Ma, the end of Coast Range magmatism, to 17 Ma, the age of the Columbia River flood basalts. If the Yellowstone plume has a typical PEP (30 000–95 000 km^3/My) then, under oceanic conditions, the plume would have been responsible for the eruption of 9.9–31.4 $\times 10^{11}$ km^3 of basalt during this interval. Therefore, Columbia River flood basalts, with a volume of 1.75 $\times 10^5$ km^3 [42], equate to 5.5–17.6% of the Yellowstone plume’s PEP during this 33 Ma interval. Put another way, the Columbia River flood basalts represent 1.8–5.8 My of lava production of a typical oceanic hotspot. Even if (for the reasons discussed above) the eruptive potential of the sub-continental Yellowstone plume was 75% of a sub-oceanic hotspot, then Columbia River flood basalts would still amount to only 21.8–70.3% of the eruptive potential of the Yellowstone hotspot. Thus, despite the fact that migration of the Yellowstone plume beneath continental lithosphere would have resulted in lava production dropping below PEP, and that ponded plume mantle would have been subjected to thermal and physical erosion, steady-state upwelling and ponding of plume mantle beneath a continent is sufficient to account for the volumes of the CRB.

Sub-continental ponding of plume mantle during amagmatic intervals may be responsible for thermal erosion and melting of the overriding lithosphere. Previous studies [2,3], based largely on thermal modelling, indicate that continental mantle above a mantle plume would have been heated only briefly, if at all, above its solidus. From this observation it has been concluded that mantle lithosphere is unlikely to contribute significantly to CFBs [3]. These models assumed instantaneous emplacement of a starting mantle plume head, coeval with flood basalt magmatism. However, continental lithosphere overriding the Iceland and Yellowstone mantle plumes was locally thinned, giving rise to long duration uplift, basin formation and rifting. Isotopic and geochemical data, including large ion lithophile element and light rare earth element enrichment, and high field strength element depletion [13,43–46], commonly indicate a significant (20%) lithospheric contribution to CFB (e.g. [44,47]). Prolonged ponding of plume mantle beneath the lithosphere, and constant recharge of these reservoirs by continued upwelling within a steady-state plume, explains the continental lithospheric contribution to flood basalts, and the thinning of the lithosphere above the ponded plume mantle.

Overriding of hotspots by continents is inevitable given the number of hotspots, their longevity, and the rate at which continents move. At present there are >30 active sub-oceanic hotspots. Assuming a random global distribution of hotspots (a simplification – see [48]) >10 active hotspots may reside beneath continents. Given an average continent-bearing plate velocity of 3 cm/yr relative to the hotspot reference frame, today's continents will move across 70–105% of the Earth's surface over the next 150–350 Ma. Assuming that the average hotspot has a lifespan of 150 My, most hotspots will, as a result of plate movement, reside for part of their lifespan beneath a continent. Continents will, on average, migrate over ~ 5 hotspots every 150 My.

For example, the African continent appears to have recently migrated over the Quathlamba hotspot (Fig. 3). This hotspot was responsible for a linear seamount chain through the Mozambique Basin which youngs toward, and terminates against, the east coast of South Africa [49]. The youngest seamount adjacent to the continental margin indicates overriding of the hotspot at 10 Ma. Extrapolation of the seamount trend into the African continent places the hotspot presently beneath east central South Africa. Although it has yet to result in a CFB, the hotspot is thought to be responsible for the thin lithosphere (~ 70 km), elevated heat flow, and high topography of southern Africa, and for recent (1983) small amounts of mafic volcanism in Lesotho. Because mantle plumes and their related hotspots are thought to have played a small but significant role in the dissipation of heat within the earth since Archean time, it seems likely that the migration of continents over pre-existing plumes is a common process responsible for significant continental tectonism and magmatism (cf. [50]).

5. Conclusions

Migration of a continent over a pre-existing steady-state hotspot can result in the punctuated eruption of lava to form CFB. Thus, CFB provinces do not necessarily mark the instantaneous emplacement of a starting plume head and the initiation of a new hotspot. Continental magmatism above a hotspot is episodic: amagmatic intervals are attributable to overriding of the plume by impenetrable continental lithosphere, and to dispersion and deflection of the plume by subducting slabs within convergent margin settings. Preservation of only a small portion of the mantle ponded during an amagmatic interval is necessary to account for the large volumes of CFBs. Ponded mantle is responsible for thermal erosion, thinning and melting, of the overlying continental lithosphere. CFB eruptions are restricted to 'thinspots' where thinned or actively thinning and rifting lithosphere provides the hotspot mantle access to shallow levels, and, in convergent margin settings, to slab windows through which plume material can gain access to the overriding lithosphere. The longevity of hotspots together with the movement of plates makes hotspot–continent interaction inevitable.

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