Basaltic magmatism and the geodynamics of the East African Rift System

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Abstract: The major and trace element and radiogenic isotope compositions of basalts from throughout the East African rift system are reviewed in the context of constraints from previous geophysical studies. The data indicate the presence of two mantle plumes, the East African and Afar plumes, which dynamically support the East African and Ethiopian plateaus. Rifting across the plateaus is accompanied by the generation of large volumes of basaltic magma and associated evolved derivatives. Relatively few mafic magmas have an unambiguous Afar mantle plume signature, notably the MgO-rich picrites and ankaramites from the 29–31 Ma Ethiopian traps, and the most recent basalts (<5 Ma) from Afar. The Eocene Amaro basalts from southern Ethiopia also have a plume source but their lower source temperatures and isotopic characteristics are distinct from those of Afar. The remaining basalts from the Ethiopian rift, and throughout the Kenya and Western rifts, have a lithospheric source region as reflected in both radiogenic isotope and trace element characteristics. The Amaro basalts are suggested as the first manifestations of magmatism from the East African plume; subsequent magmatic activity being represented by progressively younger episodes further south through Turkana, Kenya and into Northern. Tanzania, as the African plate migrated north. Despite their clear lithospheric characteristics, U-series data on geologically recent basalts from the axis of the Kenya rift show that they were generated in a dynamic melting regime. Melting is effected when lithospheric mantle heats up and becomes incorporated into the convecting mantle, hence leading to greater degrees of lithospheric thinning than are indicated by extension across individual rift basins.

The frequent association in the geological record of continental break-up, large igneous provinces (LIP), and, by inference, mantle plumes is now well established (White & McKenzie 1989, 1995; Storey 1995). In most cases, all we can study is the end result of break-up, namely the rifted margins of the two continents and the developing ocean basin between, with an aseismic ridge marking the trace of the mantle plume. By contrast, the East African rift and the associated nascent oceans of the Red Sea and the Gulf of Aden, present examples of the various stages of continental rifting and break-up that are still in close proximity to the underlying mantle plumes. Within the region, there is a record of igneous activity stretching from the Eocene to the present day which offers an unrivalled opportunity to investigate the role of mantle plumes in this important geodynamic process.

Compared with other continents, Africa has a unique topographic character. Instead of linear mountain chains related to continental collision and tectonic processes at destructive plate margins, it is dominated by basins and plateaus which in some cases, most notably in East Africa, are associated with the development of rift basins (Fig. 1). This association is now established as causative, the topography at least in part driving extension (e.g. Coblenz & Sandiford 1994), and the topography has been further linked to upwelling in the mantle beneath the lithosphere (Ebinger *et al.* 1989; Nyblade & Brazier 2002). The relationship between topography and mantle convection, as reflected in tomographic images of the deep mantle, has been developed further to include the broad plateau of southern Africa, now sometimes referred to as the 'African superswell'. This feature is related to upwelling in the lower mantle, whereas the Ethiopian and East African plateaus, and possibly the smaller plateaus of North Africa (e.g. Tibesti, Hoggar and Darfur), are a product of upper mantle dynamics (Lithgow-Bertelloni & Silver 1998). In many respects, the topography of the African continent is mapping out major features of the convective regime in the underlying mantle.

The East and North African plateaus are also characterized by recent volcanic activity, much of which is basaltic and has a mantle origin (e.g. Ashwal & Burke 1989). Given the limited extension associated with the rifts across these plateaus, the presence of any magmatism implies elevated mantle potential temperatures (e.g. McKenzie & Bickle 1984), and the eruption of 10^5–10^6 km^3 of mafic and differentiated magmas along the length of the Kenya rift (King 1978; Williams 1982;

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Baker 1987) throughout the Neogene, and similar volumes broadly associated with the Ethiopian rift, further implies that temperatures commensurate with those of mantle plumes have existed beneath this region for tens of millions of years. Although, in detail, most of the Ethiopian magmatism predates the onset of rifting in the Main Ethiopian Rift and is synchronous with or postdates extension across the Red Sea and Gulf of Aden, the eruption of such volumes of mafic lavas still requires the presence of elevated mantle potential temperatures. Aspects of the compositions of basalts from Ethiopia, notably their high ^{3}He contents (Marty *et al.* 1996), lend further strength to the argument for the presence of a mantle plume beneath the Ethiopian plateau.

The geology and composition of the basalts can therefore be used to infer properties of the underlying mantle. The volumes and rates of eruption combined with estimates of the depth, temperature and extent of melting can be exploited to develop models of the physical properties of the melting regime. The more detailed aspects of the basalt composition, notably trace elements and radiogenic isotopes, allow the distinction of different mantle source regions, such as asthenosphere versus mantle lithosphere, and possibly between different mantle plumes. Finally, analysis of U-series isotopes has the added potential to provide insights into the melting processes that have led to the production of the most recent basalts. All of these data can be used to test and complement physical models

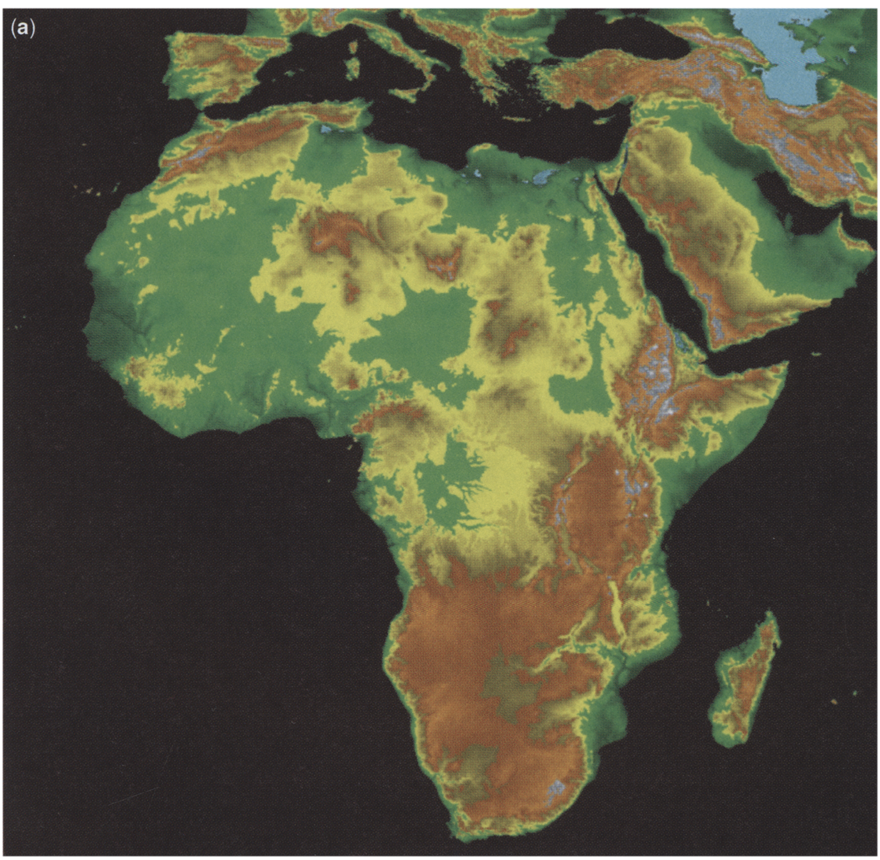

Fig. 1. (a) Digital elevation model of continental Africa and the Arabian peninsula with a larger image of the East African and Ethiopian plateaus. Note the broad plateau region of southern Africa (African superswell), and the roughly circular platforms of the East African and Ethiopian plateaus. Key: green <1000 m; yellow 1000–2000 m; brown 2000–3000 m; grey >3000 m. Image prepared from GTOPO30 digital elevation data, US National Intelligence Mapping Agency, by S. Drury.

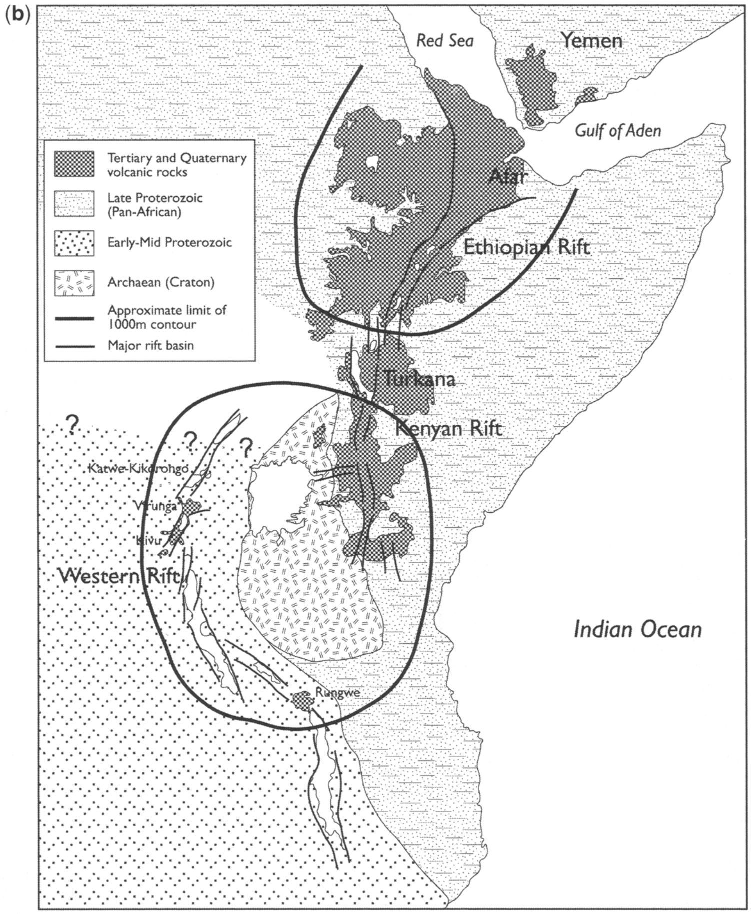

Fig. 1. **(b)** Sketch map of East Africa showing the location of the Main Ethiopian Rift and the Kenya and Western rifts, the distribution of Tertiary and Quaternary volcanic rocks and the disposition of basement of contrasting ages. The bold lines indicate the Ethiopian and East African plateaus with elevations above 1000 m (modified after Rogers *et al.* 1998).

of the mantle beneath the rift and associated plateaus.

This paper reviews critical aspects of the geochemical data from various studies of basaltic rocks from the Kenya, Western and Ethiopian rifts, the three major branches of the modern African rift system, and uses them to draw conclusions about the composition and structure of the mantle beneath this important example of continental rifting and break-up. It focuses on the interaction between mantle convection, lithosphere structure and tectonic processes along the length of the African rift, and in the process seeks to inform models of the rifting process. Specific questions that will be posed in this paper address the influence and involvement of the mantle lithosphere in the development of the rift, and the nature, number and longevity of underlying mantle plumes.

Rifting and magmatism on the East African Plateau

Lithosphere control on the location of rifting

One of the primary distinctions between the Ethiopian and East African plateaus is that whereas the first is cut by a single rift system, the latter is cut by two rifts, the Kenya or Gregory rift and the Western rift (Fig. 1). It is now well established that the development of the two rifts across the East African plateau is a consequence of lithospheric anisotropy (Nyblade & Brazier 2002). The presence of the old, cold and mechanically strong Tanzanian craton diverts rifting to the surrounding mobile belts, the Late Proterozoic Mozambique belt to the east and the Mid-Proterozoic Kibaran belt to the west. By contrast, the lack of a cratonic core underlying the Ethiopian plateau allows extension to propagate in a more uniform manner across the plateau (Fig. 1).

The Kenya and Western rifts have contrasting magmatic characteristics. The former rift is associated with large volumes ($>10^5$ km^3) of basaltic magma and evolved derivatives that have erupted throughout the Neogene (King 1978; Williams 1982; Baker 1987). As the rift has developed from wide half-graben basins into narrower full graben, so the magmatic activity has become increasingly focused on the active central rift valley. By contrast, volcanism along the length of the Western rift is sporadic and limited in extent and volume, and appears to be associated with accommodation zones between adjacent rift basins (Rosendahl 1987). The evolution of the Kenya rift from a broad depression, through half-graben to a narrow zone with the development of full-graben basins (Smith & Moseley 1993) is consistent with extension of thick lithosphere above hot mantle (Buck 1991). Extension facilitates magmatic intrusion, weakening the lithosphere which, when subject to further extensional stresses, continues to fail along the line of the original fracture; the rift thus acts as a focus for future extensional failure.

Compositional differences between the Western and Kenya rifts

In addition to the differences in volumes of basaltic volcanism in the two rifts, there are clear compositional differences between the volcanic products of the Western and Kenya rifts. This is expressed in a plot of K_2O versus SiO_2 (Fig. 2) in which mafic rocks from the Western rift at a given silica content have a higher potassium content than those from the Kenya rift. This difference is generally attributable to the presence of a potassic phase, either amphibole or phlogopite, in the mantle source region of at least the K-rich magmas (e.g. Edgar *et al.* 1976). The enrichment of potassium is accompanied by an enrichment of other incompatible elements, such as the LREE, Th, U and Ta, while the lack of fractionation between high-field-strength elements—HFSE such as Zr, Hf, Ti, Ta and Nb—and the large ion lithophile elements—LILE such as Rb, Cs, K, Ba and Sr—implies that the source underwent enrichment by addition of an alkaline mafic silicate melt (Rogers *et al.* 1992, 1998). Moreover, the radiogenic isotope characteristics of the Virunga potassic basalts (Fig. 3) further imply that trace element enrichment of the source region occurred about 1000 Ma ago (Davies & Lloyd 1989; Rogers *et al.* 1992, 1998; Furman 1995; Furman & Graham 1999), similar

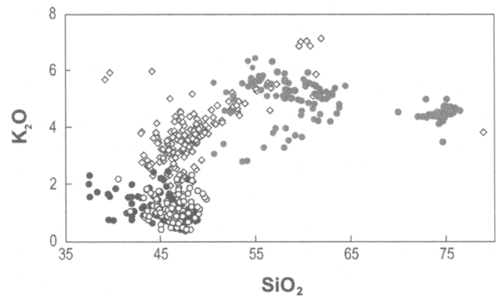

Fig. 2. Potassium–silica diagram for volcanic rocks from the Virunga province in the Western rift (open diamonds) and the Kenya rift (circles). Black circles: basalts from those parts of the Kenya rift underlain by the craton and its remobilized margins; white circles: basalts from those parts of the Kenya rift underlain by the late Proterozoic mobile belt. Grey circles represent analyses of intermediate and evolved rock types, including trachytes, comendites and phonolites. Data from Rogers *et al.* (1992, 1998) and Macdonald *et al.* (2001) and references therein.

to the age of lithosphere stabilization after the Kibaran orogeny in this part of Africa. Thus, the majority of lavas from the Western rift are attributed to a lithospheric source region with probably no input from the underlying asthenosphere, although a sub-lithospheric component has been invoked to explain the Sr and Nd isotope characteristics of the Kivu lavas (Furman & Graham 1999).

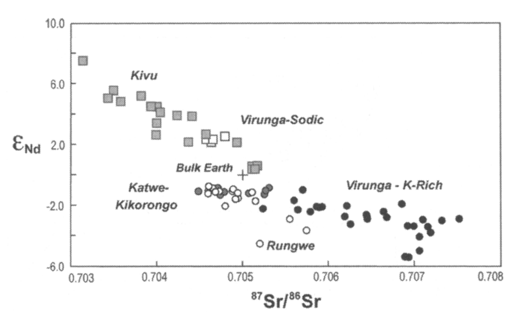

Fig. 3. Nd and Sr isotope analyses of mafic rocks from the Rungwe (open circles), Kivu (grey squares), Virunga potassic (black circles) and sodic (open squares), and Katwe–Kikorongo (grey circles) volcanic provinces in the Western rift. With the exception of the Kivu and Virunga-sodic data (from Nyiragongo), all analyses plot in the so-called enriched quadrant, signifying derivation from an old- trace element-enriched source region. (Data from Vollmer & Norry 1983; Davies & Lloyd 1989; Rogers *et al.* 1992, 1998; Furman 1995; and Furman & Graham 1999).

In contrast to the Western rift mafic lavas, basalts from the Kenya rift, while still alkaline, are not potassic and are superficially comparable with ocean island basalts (OIB) (Latin *et al.* 1993; Macdonald *et al.* 2001). They range from transitional tholeiites through alkali basalts to basanites and nephelinites, representing a spectrum of compositions that can be produced by decreasing melt fractions and increasing depths of melt segregation in a typical mantle melting regime (Macdonald *et al.* 2001). However, their radiogenic isotope ratios show a clear control exerted by the underlying lithosphere. The Kenya rift cuts across the major lithospheric boundary between the Tanzania Craton and the Mozambique Late Proterozoic mobile belt (Smith & Moseley 1993), and the basalts erupted through these contrasting basement types have distinct Sr and Nd isotope characteristics (Fig. 4). The major boundary appears to be between the mobile belt and the remobilized craton margin (Rogers *et al.* 2000).

Further differences in the trace elements of the Western and Kenya rift basalts are evident in the fractionation of the REE. All rift basalts are LREE-enriched, with LREE abundances ranging from a few tens times the chondritic abundance to many hundreds. However, significant differences are apparent in the fractionation of the HREE compared with overall LREE enrichment. These are summarized on plots of Tb/Yb against La/Yb which allow the effects of both melt fraction and residual mineralogy to be distinguished, most significantly whether or not a melt was generated in the presence of residual garnet.

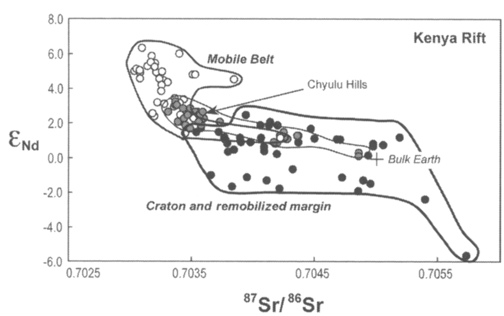

Fig. 4. Nd and Sr isotope analyses of basaltic rocks from the Kenya rift. Open symbols: basalts from the northern sector of the rift underlain by the late Proterozoic mobile belt. Black symbols: basalts from the southern sector of the rift underlain by the Tanzanian craton and its remobilized margins. Grey symbols: basalts from the Chyulu Hills (Spath *et al.* 2001), east of the main rift valley and straddling the craton–mobile belt boundary. All other data from Furman *et al.* (2004) and Rogers *et al.* (2000) and references therein.

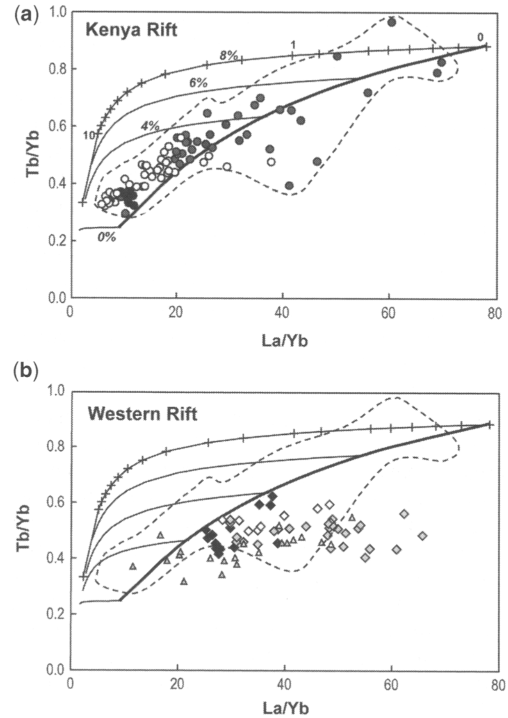

Fig. 5. Plots of La/Yb against Tb/Yb for basaltic rocks from (a) the Kenya rift and (b) the Western rift. The melting models in both diagrams are based on fractional melts derived from a fertile mantle source region with chondritic ratios of La/Yb and Tb/Yb. Partition coefficients are from Johnson (1998) and Blundy *et al.* (1998). Tick marks on uppermost curve illustrate melt fractions (labelled for 0, 1 and 10% melting). Lines labelled in italic numbers give the modal abundance of garnet in the source region (0, 2, 4, 6, 8%).
(a) Open circles represent analyses of basalts from the northern Kenya rift erupted through the Proterozoic mobile belt; and black circles, are basaltic rocks erupted through the craton and remobilized craton margin in the southern rift and northern Tanzania. Data from Macdonald *et al.* (2001) and Rogers (unpublished analyses).
(b) Grey triangles represent samples from the Kivu province (Furman & Graham 1999); white triangles, the Rungwe province (Furman 1995). Diamonds are samples from the Viunga province: black diamonds, K-basanites from Muhavura (Rogers *et al.* 1998); open and grey diamonds, K-basanites and more-evolved rocks from Karisimbi (Rogers *et al.* 1995).

The REE fractionation of basalts from the Kenya and Western rifts are summarized in Fig. 5. Superimposed on this diagram is a set of curves for melting a primitive mantle source region with 0–8% modal garnet. As the amount of garnet in the source region increases, so does the amount of variation in Tb/Yb, and so this diagram can be used to determine possible source mineralogy and depth of melting. The bold curved line at the high La/Yb extremity of the melting curves represents the locus of 0% melts, which is the theoretical maximum REE fractionation that can be produced in one melting stage from a fertile mantle source with unfractionated (i.e. chondritic) REE ratios. All samples that plot to the right of this bold line must have been derived from an enriched, probably lithospheric mantle source region whereas those to the left do not require an enriched source.

The significance of this line becomes apparent for the Western rift lavas (Fig. 5b), for which virtually all of the data plot to the right of the solid line, confirming the need for an enriched mantle source region as implied by the radiogenic isotope ratios. Note that this data field includes basalts from both the Rungwe and Kivu provinces, as well as the Virunga. In more detail, the trend within the data is shallow with limited variation in Tb/Yb for a large range of La/Yb, comparable with the shallow gradients of the melting curves with 2–4% garnet in the source. While some of this variation may be the result of clinopyroxene fractionation, overall it suggests that the melts were generated from a mantle source region with a low and constant garnet mode. Limited mineralogical variation in the mantle source is most consistent with a batch or equilibrium melting process, as is expected for melting the mantle lithosphere.

By contrast, the Kenya basalts define a much steeper trend in Fig. 5a that covers a greater range in source mineralogy, implying a source region with between 0 and 6% garnet, and melt fractions ranging between 2 and 10% (Macdonald *et al.* 2001). The most extreme REE fractionation is exhibited by those basalts erupted through the Tanzanian craton, although a few samples from all basement types plot beyond the limit of melting. However, most of the Kenya basalts plot to the left of the limiting line and so do not require an enriched source, although this does not preclude their origin from a source with fractionated REE. This behaviour is clearly quite distinct from that shown by the magmas of the Western rift and implies a marked difference in the depth and extent of the melting regime beneath Kenya, straddling the garnet–spinel lherzolite transition zone and extending possibly as shallow as the Moho (cf. Macdonald *et al.* 2001). Melting over such a depth range is only possible in a convecting system where the thermal gradient is adiabatic; in other words, where the asthenosphere extends to depths shallower than the normal base of the mantle lithosphere as has been revealed by the seismic profiles across the northern and central Kenya rift (Mechie *et al.* 1997).

Thus, the physical melting process is occurring in mantle that is currently imaged seismically as asthenosphere and so should have the compositional characteristics of ocean island basalts (OIB), yet the radiogenic isotope ratios suggest a lithospheric source region. Moreover, other aspects of the trace element compositions of Kenya basalts, in contrast to the REE, indicate a source region that is distinct from OIB. Macdonald *et al.* (2001) showed that they tend to be characterized by low Zr/Y ratios relative to Nb/Y compared with OIB. Moreover, recent studies based on more precise ICPMS analyses have shown that their Zr/Hf ratios are generally higher than many OIB with similar Zr contents (Le Roex *et al.* 2001; Spath *et al.* 2001; Furman *et al.* 2004). Such variations in the HFSE are unlikely to be a consequence of crustal contamination (cf. Macdonald *et al.* 2001), but are more easily attributed to a metasomatized mantle source probably located in the mantle lithosphere.

In summary, whereas plateau uplift is related to the dynamics of the underlying convecting mantle, the lithosphere exerts a controlling influence on the tectonics in both the Western and Kenya rifts. There is clear evidence for mantle with elevated temperatures, both from the presence of basaltic magma and from seismic attenuation (Weereratne *et al.* 2003). But the compositions of the basalts erupted within both rifts are strongly controlled by lithospheric structure and age. This is particularly apparent in the Western rift where all compositional parameters indicate a mantle lithosphere source region. In the Kenya rift, the involvement of the mantle lithosphere is most apparent in the Sr–Nd isotope systematics and particular aspects of the trace element contents. However, the REE reflect variations in the mineralogy of the mantle source that is most consistent with melting in the convecting asthenosphere, particularly in the central and northern sectors of the Kenya rift underlain by the Proterozoic mobile belt. This apparent contradiction will be discussed again below in relation to U-series isotopes.

Notwithstanding the nature of the melting process operating beneath the Kenya rift, the radiogenic isotopes reflect the age of the underlying mantle lithosphere and suggest that it represents the dominant source of basaltic magma. If the underlying mantle plume does contribute to Kenya basalts, then its isotopic composition will lie at the point common to basalts from both the mobile belt and cratonic regimes, as shown on Fig. 6. Alternatively, it may not contribute to surface magmatism in which case the basalts reveal little of the plume composition. However, a mantle plume remains a requirement as a source of heat to generate basaltic magma from the lithosphere (cf. Turner *et al.* 1996) and as a source of dynamic uplift to generate the plateau topography (Ebinger *et al.* 1989).

Rifting and magmatism on the Ethiopian plateau

The Ethiopian rift and plateau appear superficially to be much simpler than East Africa, with one rift cutting across essentially isotropic lithosphere. However, beneath that deceptive simplicity lies a longer geological history related in part to the eruption of a thick sequence of flood basalts and subsequent continental break-up. Earliest volcanic activity is represented by a sequence of basalts in southern Ethiopia that erupted during the Eocene, between 35 and 45 Ma ago (Ebinger *et al.* 1993; George *et al.* 1998). These were followed by the eruption of the Ethiopian–Yemeni flood basalts, or traps, between 31 and 29 Ma ago (Baker *et al.* 1996; Hofmann *et al.* 1997; Rochette *et al.* 1998) that covered a large area of present-day Ethiopia, Eritrea and the southern Arabian peninsula with outliers occurring west and north into Sudan and east to the Somali border. Present-day volumes exceed 250 000 km^3 (Mohr 1983) and estimates of the original volume are as high as 10^6 km^3 (Courtillot *et al.* 1999). Subsequent activity became more alkaline with the development of shield volcanoes overlying the flood basalt sequences (Kieffer *et al.* 2004) before becoming focused on the Afar depression. To the present-day volcanic and magmatic activity is largely confined to Afar, the tectonically active segments of the Ethiopian rift and the active spreading centres of the Red Sea and the Gulf of Aden (Wolfenden *et al.* 2004).

Compositionally, Ethiopian rift basalts are dominated by transitional tholeiites, falling close to the boundary between alkali basalts and olivine tholeiites (e.g. Hart *et al.* 1989; George & Rogers 2002). This characteristic extends to the Ethiopian flood basalts (Mohr 1983; Pik *et al.* 1998), making them unusual amongst flood basalts generally which are most frequently quartz-normative tholeiites (e.g. Turner & Hawkesworth 1995). Basalts from the Main Ethiopian Rift tend to be more alkalic and, as with the Kenya rift, retain a dominance of sodium over potassium (Hart *et al.* 1989; George & Rogers 2002).

Basalts from the Afar depression

The Miocene to Recent basaltic activity associated with the development of the Afar depression shows a much simpler compositional pattern than that from the Kenya rift, reflecting both the greater homogeneity of the basement lithosphere and the developing tectonic regime (Vidal *et al.* 1991; Deniel *et al.* 1994). Figure 6 illustrates the Nd and Sr isotope composition of basalts from Afar and the Red Sea and Gulf of Aden spreading centres. The latter, as might be expected, are dominated by high ε_{Nd} and low $^{87}Sr/^{86}Sr$, typical of MORB worldwide. Basalts from the Afar, by contrast, have lower ε_{Nd} and higher $^{87}Sr/^{86}Sr$ ratios than MORB, and extend to values beyond that for the bulk Earth. Significantly, there is a secular change in the isotope characteristics, with the older Miocene lavas plotting at the most extreme isotope values, and the most recent trending towards a composition with ε_{Nd} *c.* +6 and $^{87}Sr/^{86}Sr$ 0.7035 (Deniel *et al.* 1994). These most recent lavas are characterized by $^3He/^4He$ ratios up to 17 × atmospheric ($R/R_a = 17$) (Marty *et al.* 1996; Scarsi & Craig 1996) considerably greater than the value of 8 typical of most MORB, and in that respect have

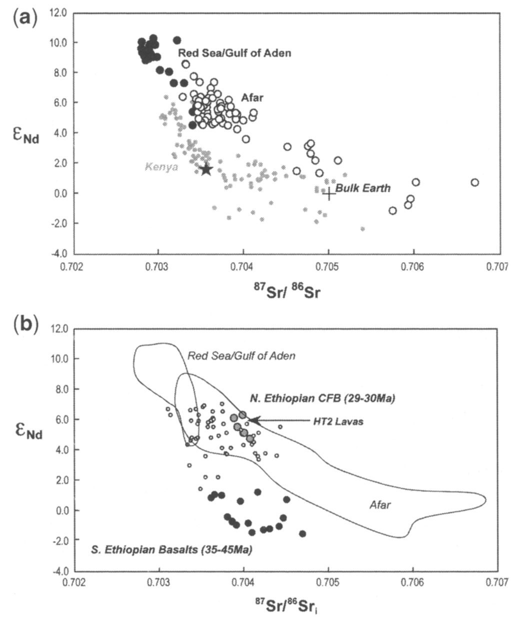

Fig. 6. **(a)** Sr and Nd isotope data from the Afar depression (Deniel *et al.* 1994; Vidal *et al.* 1991), (open circles), the Gulf of Aden (Schilling *et al.* 1992) and the Red Sea (Eissen *et al.* 1989) (black circles) compared with data from the Kenya rift (small grey circles). Note the difference between the probable composition of the Afar mantle plume and the whole array of data from Kenya. A possible composition of the East African mantle plume is marked at the common point between the 'mobile belt' and 'craton and margin' fields for the Kenya rift.
(b) Sr and Nd isotope data for northern Ethiopian flood basalts and shield volcanoes (Pik *et al.* 1998; Kieffer *et al.* 2004) compared with similar data for the southern Ethiopian basalts (George & Rogers 2002). Fields for Afar and Red Sea/Gulf of Aden based on (a). Note the similarity between the southern Ethiopian basalts and the data from Kenya.

much in common with plume-related OIB (e.g. Hannan & Graham 1996; Ellam & Stuart 2005). Hence the compositional variations within the Afar depression have been interpreted to reflect the waning influence of the lithosphere and the increasing influence of the underlying Afar mantle plume following extension as Arabia drifted slowly away from Africa (Vidal *et al.* 1991; Deniel *et al.* 1994).

The secular trends identified by Vidal *et al.* (1991) and Deniel *et al.* (1994) also showed an increase in the $^{206}Pb/^{204}Pb$ isotope ratio, which lead them to conclude that the Afar plume has a so-called HIMU (or high-μ) signature. However, more recent analysis of basalts from across the Arabian plate has shown that the HIMU signature extends beyond the topographic influence of the plume and beyond regions characterized by elevated $^{3}He/^{4}He$ ratios (Bertrand *et al.* 2003; Pik *et al.* in press). Moreover, one of the characteristics of HIMU mantle plumes worldwide is that they have low $^{3}He/^{4}He$ ratios, not high. The alternative suggestion is that the HIMU characteristics are associated with the mantle lithosphere and that the plume is rarely sampled uncontaminated with lithospheric mantle by modern basaltic activity. The preferred composition for the Afar plume is that given by Pik *et al.* (1999) that relates to the composition of magnesian high-titanium basalts from the Ethiopian flood basalts.

The Ethiopian flood basalts (traps)

The 29–30 Ma Ethiopian traps have been subdivided into three distinct magma types, the so-called low titanium (LT) and two high titanium groups, designated HT1 & HT2 (Pik *et al.* 1998). These magma groups are defined on the basis of their variation of TiO_2 with MgO and systematic variations in their immobile trace element contents. While variations within magma groups can be related to magmatic differentiation, both the LT and HT2 magma types include lavas with relatively primitive compositions such that variations between magma groups reflect either differences in source composition, mineralogy or both. In the case of the LT magmas, these may represent the mantle lithosphere or crustal contamination (Pik *et al.* 1999), although a more recent study invokes heterogeneity in the underlying mantle plume or 'Ethiopian superswell' (Kieffer *et al.* 2004).

Notwithstanding the different interpretations of the geochemical signals in the LT magmas, the HT2 magma group includes picritic and ankaramitic compositions, in addition to basalts, that are isotopically similar to modern-day Afar basalts with ε_{Nd} *c.* +6 and elevated $^{3}He/^{4}He$ ratios (18 R/R_a) (Marty *et al.* 1996; Pik *et al.* 1999). Their REE are strongly fractionated and they have unusually low Al_2O_3 contents, reflecting the influence of garnet in their source region and implying melting at pressures >3 GPa and possibly as high as 4–5 GPa (120–150 km) (Pik *et al.* 1999).

The REE fractionation within the Ethiopian flood basalts is summarized in Fig. 7a. The HT2 magmas plot at high Tb/Yb ratios even though their La/Yb ratios are not as extreme as those of the Kenya basalts. The LT basalts by contrast plot at much lower Tb/Yb ratios and concomitantly low La/Yb, emphasizing the lower content of garnet in their source region and hence a lower pressure of melting. As with the Kenya basalts, the broad linear trend defined by the Ethiopian flood basalts in Fig. 6a reflects the development of a melting regime that ranges from depths within the garnet stability field to possibly as shallow as the Moho. The data plot well within the field of melts that can be derived from a fertile mantle source region and so none of the magmas require an enriched source. This supports the contention that all are plume-derived but that the LT magmas owe their characteristic incompatible element characteristics to crustal contamination (Pik *et al.* 1999).

Melting at the depths indicated by the HT2 magmas requires elevated mantle potential temperatures, probably in excess of 1580 °C, but more significantly, the restricted melting depths also imply melting prior to lithospheric thinning. This latter observation is consistent with the tectonic record of thinning along the flanks of the Gulf of Aden and the Red Sea where fission track evidence points to a major phase of extension associated with magmatic margins between 25 and 20 Ma, post-dating CFB eruption by at least 5 Ma (Omar & Steckler 1995; Menzies *et al.* 1997). The high eruption and source temperatures and melting depths of the HT2 magmas suggest that these are the earliest unambiguous manifestations of the Afar mantle plume (e.g. Pik *et al.* 1999). Indeed, the HT2 magmas may even result from melting of the plume head during the initial stages of plume development and probably represent the true isotopic composition of the Afar plume, relatively unaffected by lithosphere interaction.

Southern Ethiopian basalts

Despite the high source temperatures and great melting depths of the HT2 magmas, they are not the earliest volcanic deposits in the Ethiopian province. These are represented by the Eocene basalts and evolved derivatives in southern Ethiopia (Ebinger *et al.* 1993), with ages between 45 and 35 Ma, and so predating the Ethiopian traps by as much as 15 Ma. They have been divided into two magma groups, the older Amaro basalts being overlain by the Gamo basalts. The latter have relatively

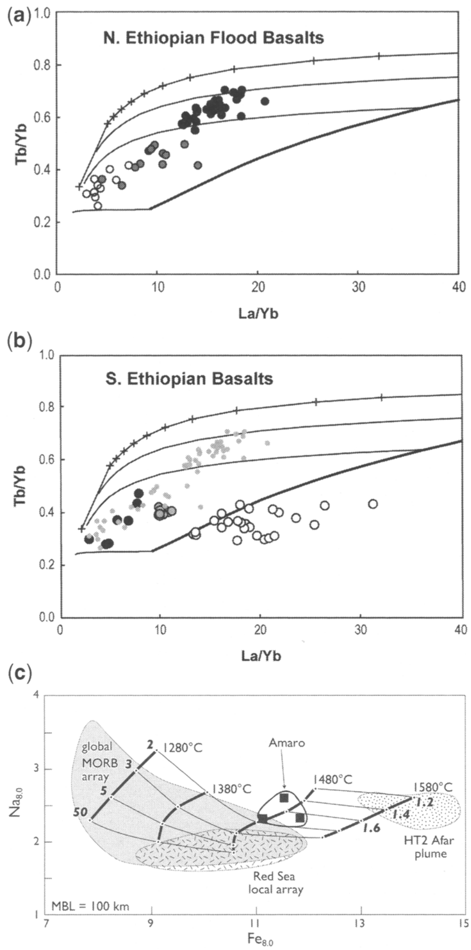

Fig. 7. **(a)** REE fractionation in the Oligocene (29–30 Ma) northern Ethiopian flood basalts. Model curves as in Figure 5 but note expanded La/Yb scale. Black circles = HT2 lavas; grey circles = HT1; open circles = LT lavas. Data from Pik *et al.* (1998) and Rogers (unpub. analyses). **(b)** REE fractionation in basalts from southern Ethiopia. Black circles = Amaro basalts; grey circles = Gamo basalts (both Eocen–Oligocene, 45–35 Ma); open circles = Miocene (19–11 Ma) Getra-Kele and Plio-Pleistocene (<2 Ma) Tosa-Sucha basalts. The latter two are associated with the development of the Main Ethiopian Rift, whereas the Amaro and Gamo basalts predate rifting in Ethiopia. **(c)** Comparison of the fractionation-corrected compositions of the Amaro basalts with the HT2 basalts from northern Ethiopia, MORB and basalts from the Red Sea. Superimposed are point- and depth-averaged melt compositions (McKenzie & Bickle 1988) for mantle of varying potential temperature and degrees of extension. Note that the HT2 basalts require derivation from elevated T_p ($\sim$1580 °C) at modest amounts of extension (β = 1.2–1.4) whereas the Amaro basalts were derived from a cooler mantle source ($T_p \sim$ 1480 °C). (After George & Rogers 2002.)

evolved compositions and appear to have developed their transitional tholeiitic characteristics from the more primitive Amaro basalts as a result of pyroxene fractionation in the mid-crust (George & Rogers 2002). The Amaro basalts are more primitive with a more pronounced tholeiitic affinity, and their major and trace element compositions reflect derivation from a mantle source region with elevated potential temperatures but lower than those typical of the HT2 from the Ethiopian traps. The REE are also less fractionated than those typical of the HT2, suggesting shallower melting depths as summarized in Fig. 7b. Lower mantle potential temperatures are also indicated by the major element compositions of the most primitive Amaro basalts, compared with the HT2 as shown in Fig. 7c. Superimposed on this diagram are point- and depth-averaged melts calculated using the parameterization of Watson & McKenzie (1991). These data show that while the Amaro basalts can be derived from a mantle source with a potential temperature of 1480 °C with moderate extension ($\beta = 1.8 - 1.6$), the HT2 magmas require considerably higher potential temperatures but with less extension ($\beta = 1.2 - 1.4$) (George & Rogers 2002).

The shallower melting regime of the Amaro basalts suggests derivation from beneath thinned continental lithosphere, yet extension in the southern Ethiopian rift began during the Miocene, at about 18 Ma (WoldeGabriel *et al.* 1990), 15–25 Ma after the their eruption. Extension contemporaneous with the Amaro–Gamo basalts occurs to the west and was responsible for the development of, amongst others, the Kaisut–Lokichar basins in NW Kenya and southern Sudan (Bosworth & Morley 1994; Hendrie *et al.* 1994), and the reactivation of the Pibor rift in southern Sudan (Ebinger & Ibrahim 1994). Moreover, the stratigraphic equivalents of the Amaro and Gamo basalts are thicker in the Omo region in SW Ethiopia (Davidson 1983), further suggesting that these Palaeogene rift systems were the main locus for Eocene mafic magmatism.

Geochemically, the Amaro and Gamo basalts were originally attributed to an early phase of Afar-related volcanism (Stewart & Rogers 1996) and their early eruption has provided a key element of the single plume model for the evolution of East African magmatism (Ebinger & Sleep 1998). However, the more detailed assessment of their compositional features summarized above (George & Rogers 2002) has shown that they have characteristics that are distinct from both recent Afar basalts and the HT2 magmas, and more similar to basalts from Kenya. Both the major and the trace element contents of the Amaro basalts imply marked petrological differences compared with the HT2 magmas from the northern Ethiopian CFB. Differences are also apparent in the Sr and Nd isotope systematics: the southern Ethiopian basalts have lower ε_{Nd} values (<2) despite similar $^{87}Sr/^{86}Sr$ ratios to the northern Ethiopian traps (0.7035), compositions which overlap those of modern-day Kenya basalts (Fig. 6b). Together, these contrasts strongly imply that the Eocene basalts from southern Ethiopia were derived from a source that was distinct from the HT2 source region, and if the Amaro source is sub-lithospheric this further implies a different mantle plume. The isotopic affinity with the Kenya basalts also fits well with their geochronology in relation to the evolution of the Kenya rift (see below).

Basalts from the Ethiopian rift

The magmatic evolution of the Ethiopian Rift is less well defined than that of the Kenya rift. In the south, after the Amaro and Gamo basalts, there is a hiatus until alkaline magmatism associated with extension across the Main Ethiopian Rift (MER) begins at 19 Ma, continuing to 11 Ma (Ebinger *et al.* 1993; George *et al.* 1998). These so-called Getra–Kele basalts were derived from an incompatible element-enriched garnet-free source region (George & Rogers 2002). The enriched nature of the source is illustrated by the REE fractionation (Fig. 7b). Both the Getra–Kele and the Plio-Pleistocene basalts from Tosa–Sucha (George & Rogers 1999) plot close to or outside the limit of melts that can be derived from primitive mantle, and their low Tb/Yb ratios indicate a garnet-poor source region. Once again, they appear to have been derived from the mantle lithosphere although they have an ambiguous isotopic signal with ε_{Nd} values intermediate between the Afar plume and that proposed for the Kenya plume. These isotopic characteristics are also apparent in the basalts from other sectors of the rift (e.g. Hart *et al.* 1989) although their REE fractionation is less extreme (Rooney *et al.* 2005), and it is clear that further investigations into the igneous geochemistry of Ethiopian rift basalts are required before their significance can be deduced. The most recent volcanic activity is located in a series of en echelon volcanic segments (sometimes referred to as the Wonji fault belt) comprising extensive mafic dyke injection (Wolfenden *et al.* 2004) and central volcanoes characterized by evolved alkaline magmas (e.g. Gasparon *et al.* 1993; Peccerillo *et al.* 2003).

U-series isotope analyses of basalts from the Kenya rift and Afar

Critical aspects of basalts from the African rift discussed above are contradictory. For example, in the Kenya rift and possibly in the Ethiopian rift, selected trace element (e.g. Zr/Hf ratios) and isotope characteristics indicate a lithospheric source region. By contrast, variations in the REE indicate melting at a range of depths, consistent with geophysical studies that show the presence of asthenospheric partial molten mantle as shallow as the base of the crust in parts of northern Kenya. Analyses of U-series isotopes in recent mantle-derived magmas allows the investigation of different aspects of the melting regime and comparison with physical models of melt generation and segregation (e.g. Bourdon & Sims 2003; Lundstrom 2003). Such studies are as yet in their infancy in the African rift but of particular relevance here is their use in relating geochemical variations to the physical nature of the melting regime beneath the rift axis.

Early studies using α-counting techniques (Black *et al.* 1997, 1998) revealed that Kenya basalts are characterized by an excess of ^{230}Th over ^{238}U, and this has been verified using mass spectrometric analyses. The results of more recent studies are summarized in Fig. 8 where they are compared with similar data from ocean island basalts which are generally considered to be plume-related. The available data from Afar (Vigier *et al.* 1999; Rogers *et al.* unpublished data) plot in a similar position to many OIB. In particular, they define a near-vertical trend within which the U/Th elemental ratio (expressed as the ($^{238}U/^{232}Th$) activity ratio) varies little, whereas the ($^{230}Th/^{232}Th$) ratio deviates from the equiline, resulting in a ^{230}Th excess of up to 40%. Young (<10 ka) basaltic rocks from the Kenya rift

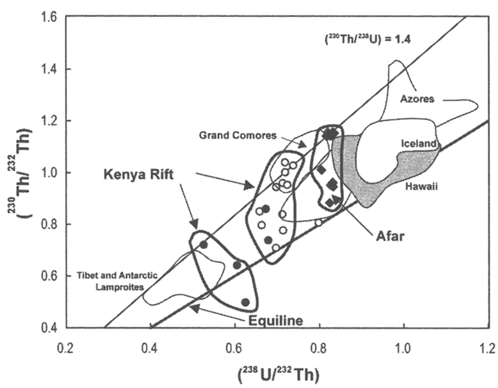

Fig. 8. U-series equiline diagram comparing analyses from the Kenya rift (Rogers *et al.* unpub. data) and Afar (Vigier *et al.* 1999; Rogers *et al.* unpub. data). Note that the Kenya data plot at lower ($^{238}U/^{232}Th$) ratios than Afar and the majority of OIB. The near-vertical arrays in both the Afar and Kenya data imply dynamic melting in the convecting mantle beneath both regions.

also plot above or close to the equiline but define two fields, one at low ($^{238}U/^{232}Th$) ~0.55 – 0.6, which exhibits limited ^{230}Th excess and a second with higher ($^{238}U/^{232}Th$) ~0.7 – 0.75 and more variable ^{230}Th excess, ranging from 0–46%.

The results emphasize the difference in U/Th ratios between Kenya basalts and OIB, in that U/Th ratios are generally lower than most OIB. Th is more incompatible in the mantle than U and so a lower U/Th ratio in the source region of the Kenya basalts is consistent with a trace element-enriched source region. Note also that the three samples that plot at the lowest U/Th ratios are comparable with lithosphere-derived lamproites from Tibet and Antarctica, further emphasizing the link with the mantle lithosphere. Finally, comparison with results from Ardoukoba in the Afar depression (Vigier *et al.* 1999) reveals the distinctive higher U/Th ratios of the latter relative to Kenya and more comparable with OIB. This is further evidence in favour of the distinctiveness of the mantle source regions tapped by basalts from the Kenya rift and those from at least the Afar segment of the Ethiopian rift.

The most significant feature of these data is the variable ($^{230}Th/^{232}Th$) at almost constant ($^{238}U/^{232}Th$) in both the Kenya basalts with the higher ($^{238}U/^{232}Th$) ratios (~0.7) and the Afar basalts. Given that the samples are all historic or recent and that crustal residence times have been negligible, such vertical arrays on the equiline diagram are indicative of a dynamic melting regime in convecting mantle. The 40% ^{230}Th excess in the basalts from both regions is the maximum that can be produced during melting of a garnet-bearing mantle source region (Bourdon & Sims 2003) and this is consistent with the variations in the REE described above. However, it is in conflict with the conclusion from radiogenic isotopes and aspects of the trace element characteristics of the Kenya basalts that they have compositions largely controlled by the mantle lithosphere, because the lithosphere is rigid and does not convect, and hence cannot easily develop a dynamic melting regime.

One possible mechanism that may reconcile this apparent contradiction involves the thermal erosion of the lithosphere, as has been suggested to account for the isotope characteristics of the Kivu volcanic rocks of the Western rift and basanites from the Huri Hills in Kenya (Furman & Graham 1999). Beneath the Kenya rift, the total lithospheric thinning revealed by seismic investigations is greater than that determined from upper crustal extension alone. The lithosphere underlying the rift flanks has a thickness of >120 km but beneath the axis the seismic lithosphere has been almost completely replaced by mantle with low seismic velocities. Part of this reduction in seismic velocity is related to the local presence of up to 6% melt, but significantly at least half of the seismic attenuation is due to temperature increases in the solid mantle (Green *et al.* 1991). As a consequence of this heating, the lithosphere has become more ductile and involved in convective movement, thus contributing to lithospheric thinning and allowing melts with lithospheric characteristics to be generated in a dynamic melting regime.

Discussion

One plume or more beneath East Africa?

The involvement of mantle plumes in the evolution of the present topography and tectonics of East Africa has led to competing models as to the number of distinct plumes that may lie beneath. At one extreme, Ebinger & Sleep (1999) suggest that magmatism across the whole of East and North Africa can be explained as a consequence of the impact of a single plume beneath southern Ethiopia 45 Ma ago. After impact, mobile plume mantle spread beneath the lithosphere, channelled by the inverse topography at its base, generated during Mesozoic and Tertiary extensional events. By contrast, George *et al.* (1998) and Rogers *et al.* (2000) have presented geochronological and geochemical evidence that suggests the migration and composition of basalts from southern Ethiopia and Kenya represent the products of one mantle plume whereas the basalts from the northern Ethiopian plateau and the Afar depression represent the products of another.

A single plume model is difficult to reconcile with geochemical data, especially the He, Sr and Nd isotopes. Comparing the Nd and Sr isotopic compilations from the Kenya rift and Afar for the most recent basaltic lavas (<12 Ma) emphasizes the distinct features of the two systems (Fig. 6). Significantly, there is little compositional overlap between the two, Kenya basalts being displaced to lower values of ε_{Nd} for a given value of $^{87}Sr/^{86}Sr$. The Red Sea and Afar trends focus on isotopic values of $\varepsilon_{Nd} > +6$ and $^{87}Sr/^{86}Sr \approx 0.7035$, those for Kenya have a common composition at $\varepsilon_{Nd} \approx 1 - 2$ and $^{87}Sr/^{86}Sr < 0.7035$ (Fig. 6). While the latter may or may not represent the composition of the sub-lithospheric mantle beneath the East African plateau, there is clearly no contribution to the Kenya basalts from material with an Afar-like composition.

A recent review of the distribution of elevated $^{3}He/^{4}He$ ratios in both ground waters and basaltic rocks (Pik *et al.* 2006) further confirms that this signal is present in the East African rift only in samples derived from within the geographical confines of the Ethiopian plateau. Elevated $^3He/^4He$ ratios have yet to be confirmed beyond the limits of the plateau, suggesting that the compositional effects of the Afar plume do not extend beyond the limit of its physical influence. When high $^3He/^4He$ is present in basaltic rocks, it is a particularly unambiguous indicator of a mantle plume source of probable deep mantle provenance (e.g. Ellam & Stuart 2005; Hannan & Graham 1996). The absence of high $^3He/^4He$ ratios away from the topographic influence of the Afar plume (i.e. the Ethiopian plateau) is strong evidence against the spread of Afar plume material beyond these geographical confines (Pik *et al.* 2006). These observations are strong evidence that the two plateaus are supported by upwelling mantle with distinct compositions, even though the composition of that beneath East Africa cannot be defined with certainty.

Encouraged by the single plume model, Furman *et al.* (2004) related recent basalts from the Turkana region of northern Kenya to a mantle plume source region similar to Afar, chiefly on account of their 'high-μ' Pb isotope ratios. However, as discussed above, the so-called 'high-μ' Pb isotope characteristics are found throughout the region and are not confined to the Ethiopian plateau, and an alternative view places this material within the mantle lithosphere (Bertrand *et al.* 2003).

Moreover, other compositional characteristics of the Turkana basalts are inconsistent with plume derivation. For example, they have high SiO_2 and low FeO_t contents, and lack a garnet signature in their REE fractionation, implying melt generation at pressures of 1.5–2.0 GPa (Furman *et al.* 2004), well within the depth limit of the extended mantle lithosphere. The Turkana depression shows the least topographic elevation of any magmatic region within the African rift (*s.l.*) south of Afar, coupled with the least negative regional Bouguer gravity anomaly (Tessema & Antoine 2004), and therefore shows the least dynamic uplift of any part of the East African rift system. In addition, the depth to the top of the melting column estimated from seismic velocity profiles is also deeper beneath Turkana and the northern Kenya rift than it is further south, which suggests that mantle potential temperatures are lower. Thus, it is unlikely that Turkana is underlain by a physical mantle plume, although this does not preclude the possible incorporation of plume-derived materials in the mantle beneath Turkana (cf. Furman *et al.* 2004).

Magma migration and plate motions

Tracing back through the development of the two systems, it is clear that the most voluminous magmas generated at the greatest depth and the highest temperatures are found in the northern Ethiopian flood basalts, implying that these represent the earliest manifestation of the Afar mantle plume. The Eocene Amaro basalts of southern Ethiopia, by contrast, were produced at shallower depths and lower temperatures and have isotopic characteristics that are more comparable with those of more recent basalts erupted in the Kenya rift. Moreover, they define the oldest end of a time sequence related to the onset of magmatism that migrates from southern Ethiopia in the Eocene, through Turkana in the Oligocene and southwards across the Kenya dome and into northern Tanzania during the Miocene (Fig. 9) (George *et al.* 1998). Note that this age progression refers to the onset of magmatism, much of the length of this trail being characterized by magmatism and geothermal activity to the present day.

The southward migration of magmatism corresponds to $\sim 10°$ latitude in 40 Ma and is equivalent to an average velocity of about 25 mm a^{-1}. This rate compares favourably with current and previous African plate vectors (O'Connor *et al.* 1999) which suggest counterclockwise rotation about a pole close to the Canary Islands with a decrease in velocity from 30 mm a^{-1} to 20 mm a^{-1} occurring 19 Ma ago. These plate motions result in a northward vector for the African plate in the vicinity of the African rift, consistent with the apparent southward migration of magmatism since 45 Ma.

The alternative one- and two-plume models for the African rift have recently been investigated numerically by Lin *et al.* (2005). Their results reveal that the distribution and age progression of magmatism in both the Kenya and Ethiopian rifts are reproduced most closely in a model with two mantle plumes. The distribution of magma predominantly along the Kenya rift is consistent with a plume currently beneath the Tanzanian craton, in

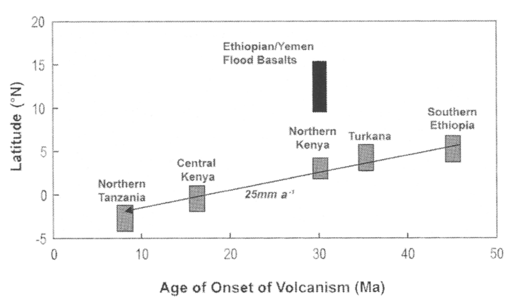

Fig. 9. Age progression in the onset of magmatism from southern Ethiopia south through Kenya into northern Tanzania at a rate of *c.* 25 mm a^{-1} (after George *et al.* 1998). The onset of magmatism in the Ethiopian/Yemen flood basalts does not lie on this trend.

a position similar to that imaged seismically (Weeraratne *et al.* 2003), and requires the craton to approach the plume from the SW, consistent with plate motions. They also show that a stagnation streamline develops between the two plumes, allowing for the generation of basalts with distinct compositions and a limited zone between where hybrid magmas may occur (Lin *et al.* 2005). This may explain the ambiguous characteristics of the Turkana basalts, erupted in the extended region between the two plumes. While these models do not represent an independent test of the 'two-plumes' concept, they serve to demonstrate that a scenario involving two mantle plumes in positions that explain the present-day dynamic topography is consistent with our knowledge of the evolution of the system as a whole and our current understanding of mantle dynamics.

The change in plate velocity vector at 19 Ma reported by O'Connor *et al.* (1999) occurred shortly after major rifting along the flanks of the present-day Red Sea as reflected in fission track dating and other geological evidence. Prior to this time, the African and Arabian plates were moving as a single unit, largely driven by slab pull as the remnants of the Tethys Ocean were subducted beneath the Eurasian plate. Extension across the Red Sea, possibly aided by the earlier emplacement of the Afar mantle plume (Bellahsen *et al.* 2003), cut this driving force from the African plate, which consequently slowed and changed direction.

A major consequence of this important change in the geodynamics of the African plate was the change in the orientation of the major extensional stress. The current stress map of the African plate (Coblenz & Sandiford 1994) shows that the continent is everywhere under extension and that stress is at a maximum over the Ethiopian and East African plateaus, where the topography attains its maximum altitude. The present-day stress orientation across the Ethiopian and East African plateaus is roughly east–west, hence the orientation of the Neogene rift systems, albeit partly controlled by basement fabrics (Smith & Moseley 1993). Given the plate tectonic history of the African plate, it is likely that this regime has dominated since Arabia finally split away from Africa at *c.* 20 Ma (Omar & Steckler 1995; Menzies *et al.* 1997).

Prior to this event, the stress regime is more speculative, but considering that plate motions would then have been dominated by slab pull to the NE, it is likely that the main extensional stress would be orientated in a NE–SW direction resulting in NW–SE-orientated extensional basins. Indeed, this is the general observation and basins ranging in age from the Cretaceous (Anza graben) through the early Tertiary (Bosworth 1992; Bosworth & Morley 1994) and including the Red Sea itself, are oriented roughly in a NW–SE direction, but that during the late Palaeogene and throughout the Neogene, basins become oriented roughly north–south (Kenya rift) or NE–SW (MER) (Ebinger & Ibrahim 1994), corresponding to the modern-day stress orientation.

Thus a pattern emerges of extensional basins controlled in part by basement structure, but developing in response to the stress regime across the whole plate which is in turn a product of far-field, plate-boundary forces and topography. Moreover, given that the major topographic features of continental Africa are controlled by mantle convection and, in particular, mantle plumes, it is inevitable that many of these extensional basins will be accompanied by basaltic magmatism. In any continental environment, extension is an essential element for the development of basaltic volcanism which will consequently be focused on contemporary rift zones. The southward migration of magmatism from southern Ethiopia through Kenya to Tanzania followed the southward propagation of the rift system which was in turn caused by the southward migration of dynamic uplift caused by the underlying mantle plume. The main extensional driving forces in this part of the African plate are derived from variations in the gravitational potential energy across the plate. This contrasts with the situation during the break-up of the Red Sea and Gulf of Aden where extension once initiated, continued to develop into first broad rifted basins and then the young oceanic basins we now see. These well-developed extensional basins subsequently acted and continue to act as a focus for magmatic activity which consequently does not migrate in response to the movement of the overriding plate but exploits the lithospheric weaknesses generated at the extensional plate boundaries.

Conclusions

Combining the evidence from basalt petrology and geochemistry with the large-scale topographic and geophysical features of the East African rift system leads to a system that is driven by plume-related dynamic uplift, the intraplate stress field and plate motions. The dynamic topography contributes significantly to the extensional force that drives rifting across both the East African and Ethiopian rifts, but that force also includes a contribution from far-field forces generated within the African plate. The association of magmatism with rifting in Africa is a direct consequence of the link between topography and extensional stress, simply because the greatest elevations are the result of dynamic support by the underlying mantle plume. In other words, extension only occurs where the lithosphere is underlain by a plume and that is also the region where magmatic activity will occur.

Magma migration from southern Ethiopia to northern Tanzania is a consequence of plate motion over the East African plume while magmatism in northern Ethiopia resulted initially from the impact of the Afar plume *c.* 30 Ma ago. Subsequently, as Arabia split away from the African plate, a rift formed along the line of the Red Sea and Gulf of Aden, prior to full separation and the onset of sea-floor spreading. Somewhat more speculatively, the change in orientation of extensional basins from NW–SE throughout the Cretaceous and Palaeogene, to broadly north–south during the latest Palaeogene throughout the Neogene and into Recent times, may be due to the isolation of the African plate from slab pull once Arabia separated fully and became an independent plate.

Mafic magma throughout the Kenya rift, Western rift and probably in large parts of the Ethiopian rift, were and continue to be derived from shallow and compositionally distinctive source regions within the mantle lithosphere. Melting in the mantle beneath the Kenya rift is achieved by thermal erosion of solid lithospheric material into the convecting asthenosphere, with melt generated as the material heats up and becomes incorporated in convective movements beneath the rift axis.

Notwithstanding the above description, not all of the evidence is consistent with the 'two-plume' model and not the least of these problems concerns the migration of magmatism and extension northwards from southern Ethiopia towards Afar in the development of the Main Ethiopian Rift (Wolfenden *et al.* 2004). This is in the opposite direction from that predicted by plate motions over a stationary hot spot. Moreover, present-day Nd and Sr isotope characteristics of Ethiopian rift basalts are distinct from those considered diagnostic of the Afar plume, yet they are characterized by Afar-like $^3He/^4He$ isotope ratios. Clearly, the details of the evolution of the Ethiopian rift and its magmatic history require further investigation and will no doubt provide further tests of the current model. While such topics represent targets for future research, the 'two-plume' model accounts for many of the large-scale geological, geophysical and petrological observations throughout the East African rift system and remains a viable explanation for the evolution and development of the system throughout the Palaeogene and Neogene.

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