Os – Hf isotopic insight into mantle plume dynamics beneath the East African Rift System

Wendy R. Nelson a,⁎, Tanya Furman a, Peter E. van Keken b, Steven B. Shirey c, Barry B. Hanan d

⁎ Corresponding author at: Department of Earth and Atmospheric Sciences, University of Houston, 510 Science and Research Bldg. 1, Houston, TX 77204, United States. Tel. +1 202 478 8477; fax: +1 202 478 8821.
E-mail addresses: wrnelson2@uh.edu (W.R. Nelson), furman@psu.edu (T. Furman), keken@umich.edu (P.E. van Keken), shirey@dtm.ciw.edu (S.B. Shirey), bhanan@mail.sdsu.edu (B.B. Hanan).

Available online 2 June 2012
Received in revised form 1 May 2012
Received 2 June 2010
Article history:

A R T I C L E  I N F O

Article history:
Received 2 June 2010
Received in revised form 1 May 2012
Accepted 26 May 2012
Available online 2 June 2012
Editor: D.B. Dingwell

Keywords:
Mantle plume
Os
East African Rift
Afar
HIMU

A B S T R A C T
We report new Os and Hf isotopic data on mafic lavas from several key portions of the East African Rift System (EARS) with the goal of determining how contributions from various source domains influence volcanism in the evolving rift system. Our study uses picrites and basalts associated with the Afar plume in NW Ethiopia and with prolonged extension in Turkana, N Kenya, as well as mafic lavas from Kivu and Rungwe in the Western Branch of the EARS. Basalts from NW Ethiopia and Turkana have low Os concentrations (9–22 ppt) and display a range of 187Os/188Os (0.1239–0.4366). The 30 Ma high-TiO2 picrites from NW Ethiopia and 20–23 Ma picrites from Turkana have higher Os concentrations (579–1120 ppt) and display a range of 187Os/188Os (0.1450–0.1483). None of the picrites display convincing evidence for crustal or subcontinental lithospheric mantle input. Instead, the data are consistent with geochemical and geophysical models that demonstrate early evolution of the EARS was supported dynamically by geochemically distinct regions of mantle upwelling. Specifically, NW Ethiopian lavas are chemically analogous to the “C”-like Afar plume while Miocene Turkana lavas display HIMU-like geochemical features. The HIMU component in Turkana lavas can be generated by mixing ~30% ancient (17–2 Ga) hydrothermally altered subducted oceanic crust with ~70% “C”-like mantle material (i.e. < 1 Ga recycled hydrothermally altered oceanic crust). In contrast, Kivu and Rungwe lavas have lower Os concentrations (3–87 ppt) and more radiogenic initial 187Os/188Os (0.1615–0.3610) that appear to be dominated by contributions from metasomatized lithospheric mantle. Seismological observations indicate that there are thermochemical heterogeneities within the deep-seated African superplume; these heterogeneities are a plausible source for the ancient recycled oceanic crust contributing to Miocene volcanism in Turkana. We propose that mafic magmatism in both the Afar region and northern Kenya are derived from different portions of this long-lived thermochemical feature.

© 2012 Elsevier B.V. All rights reserved.

1. Introduction

The East African Rift System (EARS) is the archetypal example of modern, active continental rifting. The EARS has experienced over 40 Ma of volcanism that generated inter alia >350,000 km3 of flood basalts that predate continental rifting (e.g. Hofmann et al., 1997). It is clear that locally-derived ambient mantle can upwell passively during late or mature stages of continental rifting and ocean basin formation (e.g., the Gulf of Aden and the Red Sea; Altherr et al., 1990; Barrat et al., 1990; Schilling et al., 1992; Rogers, 1993; Volker et al., 1993). However, in order to better understand the processes influencing the initiation of continental rifting, it is critical to determine the nature and origin of voluminous pre-rift mafic volcanism.

Although active debate rages on the existence of plumes in general (e.g. Konter et al., 2008; Anderson, 2011), the dynamic history and evolution of the EARS — especially pre-rift volcanism — has long been linked in previous studies to the influence of at least one upwelling mantle plume (e.g. Schilling, 1973; Richards et al., 1989; White and McKenzie, 1989; Vidal et al., 1991; Baker et al., 1996b; Stewart and Rogers, 1996; Hofmann et al., 1997; Pik et al., 1999; Kieffer et al., 2004; Furman et al., 2006a). Classic fluid dynamical experiments of thermal plumes predict the formation of a large, bulbous head trailed by a thin tail (Whitehead and Luther, 1975; Campbell and Griffiths, 1990). This geometry matches the pattern of volcanism observed in...
the northern EARS and a number of hotspot regions: as the plume head reaches the base of the lithosphere, large volume flood basalts are emplaced over a relatively short period of time followed by smaller volume plume tail volcanism persisting for tens of millions of years (Morgan, 1971). Such thermal plumes are thought to be caused by instabilities from thermal boundary layers within the mantle, perhaps at the base of the mantle (e.g. Morgan, 1971; Richards et al., 1989; van Keken, 1997).

There is a growing body of geophysical evidence illustrating that the deep mantle displays large scale chemical heterogeneity resulting from the accumulation of subducted slabs that penetrate the 660 km discontinuity (e.g. van der Hilst et al., 1997), leaving this portion of the mantle poorly mixed for long periods of time (e.g., Kennett et al., 1998; Lay et al., 1998). The evidence for chemical heterogeneity is particularly strong for the seismically anomalous region in the deep mantle beneath much of Africa and the southern Atlantic Ocean, known as the African superplume (Ni et al., 2002; Simmons et al., 2007). This large lower mantle region has similar characteristics to its complement below the Southern Pacific ocean and is recognized to be a thermochemical anomaly that is both warm and dense (Garnero et al., 2007). These regions have been suggested to be fairly stable over geological time (Burke et al., 2008) and the history of global hot spot activity suggests that localized upwelling may be derived from the edges of these structures (Torsvik et al., 2010). Recent modeling efforts (e.g. Davaille et al., 2003; Lin and van Keken, 2005, 2006; Kumagai et al., 2008) show that the incorporation of thermochemical heterogeneities (i.e. ancient recycled oceanic slabs) in the deep mantle can result in upwellings that display significant deviations from classic thermal plume behavior (e.g., episodic pulsing, an inconsequential plume head, and small secondary plumes branching from a larger plume) that is reflected in the nature and volume of volcanism on the Earth’s surface.

Geophysical data show that the EARS is supported dynamically by regions of mantle upwelling centered currently beneath the Afar region (Ethiopia) and Lake Victoria (Tanzania) (e.g. Ebinger et al., 1989a; Simiyu and Keller, 1997; Benoit et al., 2006). Early (i.e. pre rift) volcanism associated with these upwellings occurred in NW Ethiopia-SW Yemen (~30 Ma; Hofmann et al., 1997) and S Ethiopia–N Kenya (~45–40 Ma; George et al., 1998; Furman et al., 2006b), respectively. These two areas are readily distinguished on the basis of both eruptive behavior and geochemical characteristics of the mafic lavas. Specifically, 30 Ma flood basalts in NW Ethiopia are isotopically and geochemically similar to those of high 3He/4He OIB (up to 19.5 Ra; Marty et al., 1996). This region has been volcanically active through to the present though post-Miocene lavas are small in volume and confined to the modern rift axis. In contrast, Miocene basalts and picrites from N Kenya have isotopic ratios indicative of a HIMU mantle component, including low 3He/4He (5.9–7.8 Ra; Furman et al., 2006b; Pik et al., 2006). Volcanism in this region is more episodic in nature with 5–10 Ma gaps in volcanic activity. In order to more fully understand the processes that initiated rifting in the EARS, it is essential that we understand the origin of these early, distinct volcanic episodes.

In this paper we integrate new Os and Hf isotopic data from pre-rift EARS mafic lavas with existing Sr–Nd–Pb isotopic data. Os isotopic data are useful in investigating source contributions to mantle-derived volcanism, especially when used in conjunction with other radiogenic isotope systems. Our goals are to determine the geographic extent and petrogenetic origin of the distinct chemical signatures recorded in the two volcanic provinces, understand how these heterogeneities correlate with the distribution and timing of volcanism, and evaluate the relationship of these signatures to the mantle. We interpret these new data in conjunction with results of numerical and fluid dynamical models as well as observed spatial, temporal, and geophysical patterns to better constrain the early mantle dynamics in this region and their potential relationship to the African superplume.

2. Geologic setting

2.1. Timing of volcanism and rifting

The EARS (Fig. 1) is marked by a prolonged history of episodic volcanic activity. volcanism began ~40–45 Ma along the S Ethiopia–N Kenya border (George et al., 1998; George and Rogers, 2002; Furman et al., 2006b) and has continued over a broad area through to the present, although the location of activity has changed through time. Mafic volcanism in this region is volumetrically significant and, in most cases, involves mappable edifices and lava flows; no thick sequences of flood basalts have been emplaced. At 30 Ma, massive flood basalts (350,000 km³) were produced in NE Ethiopia, Yemen, and Eritrea (Baker et al., 1996a; Stewart and Rogers, 1996; Hofmann et al., 1997; Pik et al., 1999; Kieffer et al., 2004). Subsequent volcanism produced a few shield volcanoes in NW Ethiopia at 25–22 Ma and 11 Ma (Kieffer et al., 2004). Post-Miocene through modern activity is essentially confined to the valleys of the Main Ethiopian Rift, Kenya Rift and Western Rift (e.g. Ebinger et al., 1989b; Hart et al., 1989; Stewart and Rogers, 1996; George et al., 1998; Chernet and Hart, 1999; Furman et al., 2004, 2006b).

The eruption of Oligocene flood basalts in NW Ethiopia, Yemen and Eritrea at 30 Ma proved to be more tectonically influential than the earlier ~40–45 Ma volcanism to the south. Specifically, continental breakup of the Afro-Arabian plate began with rifting in the Red Sea (~28 Ma; e.g., Hofmann et al., 1997; Wolfenden et al., 2005) and the Gulf of Aden (~27 Ma; Fantozzi, 1996). The EARS represents the third limb of the triple junction, although rifting did not initiate until ~25 Ma far to the south in Turkana (N Kenya) by reactivating Cenozoic faults surrounding the Anza graben as well as generating new ruptures (Morley et al., 1992; Hendrie et al., 1994). This rifting episode occurred synchronously with another pulse of volcanism in Turkana (26–16 Ma). From there, volcanism and extension propagated both to the north and the south. Extension to the north created the southern Main Ethiopian Rift that eventually merged with faults.

![Fig. 1. Shaded relief map of the East African Rift system. Dashed lines enclose regions with elevation over 1200 m above sea level. The locations of initial volcanism in each area are represented by stars scaled for approximate total eruptive volumes. Sample locations are marked by black squares. Other localities referenced in the text are also highlighted.](image-url)
propagating southward from the Afar triple junction around 11 Ma (e.g., Wolfenden et al., 2004). Contemporaneous extension to the south developed around the Tanzania craton, producing the Kenya and Western Rifts (Ebbing, 1989; Nyblade and Brazier, 2002; Roberts et al., 2012) and their associated volcanic provinces including Kivu and Rungrwe (e.g. Ebbing, 1989; Ebbing et al., 1989b; Furman and Graham, 1999; Ebbing and Furman, 2002).

2.2. Geophysical observations

Geophysical data demonstrate low velocity anomalies attributed to elevated mantle temperatures extend to depths between 440 and 660 km beneath both the Afar region and the Tanzania craton (e.g., Simiu and Keller, 1997; Zeyen et al., 1997; Nyblade et al., 2000a, 2000b; Weeraratne et al., 2003; Benoit et al., 2006; Bastow et al., 2008; Chang and Van der Lee, 2011; Adams et al., 2012; Hansen et al., 2012). Seismic studies show that the low velocity zone beneath Afar dips to the west (Benoit et al., 2006) and may merge with the low velocity zone beneath Tanzania at depths greater than 660 km (Montelli et al., 2006). Global tomography models demonstrate the weak combined anomaly continues to dip westward and merges equivocally with a large, deep low-velocity feature known as the African superplume or superswell (Ritsema et al., 1999; Grand, 2002; Montelli et al., 2006). The African superplume originates at the core-mantle boundary beneath southern Africa and shallows towards the northeast, approaching the base of the lithosphere beneath the Lake Victoria region (e.g. Ritsema et al., 1999; Zhao, 2004).

The connection between the upper and lower mantle low velocity anomalies in this region is not clear because the spinel–perovskite phase change beneath the mantle transition zone precludes clear seismic imaging. Therefore, we cannot determine explicitly the ultimate depth of origin for the anomalies (Nyblade et al., 2000a; Benoit et al., 2006; Montelli et al., 2006). However, recent studies demonstrate there may be a connection between the shallow low velocity regions and the deep African superplume (Huerta et al., 2009; Adams et al., 2012; Hansen et al., 2012). Huerta et al. (2009) report a localized depression in the 440 km discontinuity beneath the rifts in Kenya and northern Tanzania as well as a large (~600,000 km²) regional depression across the 660 km discontinuity that they attribute to the perovskite–majorite phase transition at elevated mantle temperatures. Huerta et al. (2009) interpret these features as anomalously warm, rising mantle material — likely originating from the African superplume — that stagnated at the base of the transition zone and created a thermal perturbation that extends across the 660 km discontinuity. They suggest that the broad warming subsequently generated smaller, localized thermal instabilities that rose beneath the Kenya and Western rifts. In the deep mantle, Sun et al. (2010) used waveform tomography to image seismically narrow pipe-like structures emanating from the top of the African superplume, demonstrating that such structures can indeed originate from this larger feature. Though current limitations in seismic resolution make it impossible to connect definitively these shallow low velocity zones to the African superplume, the above observations support a direct relationship.

2.3. Geochemical overview

Previous work has shown that mafic lavas in central and northern Ethiopia are isotopically distinct from those found in S Ethiopia and Kenya (Pik et al., 1998, 1999; Rogers et al., 2000; George and Rogers, 2002; Kieffer et al., 2004; Furman et al., 2006b). The most primitive flood basalt lavas from NW Ethiopia are characterized by radiogenic 87Sr/86Sr~0.704, 143Nd/144Nd~0.51295, and 206Pb/204Pb~18.8–19.2 (Fig. 2; Pik et al., 1999). This material was originally thought to be HIMU-like (Schilling et al., 1992) but Rooney et al. (2012) recently attributed this signature to the “C” or “common” mantle component, which is interpreted to represent mantle containing young (~1 Ga) recycled hydrothermally altered oceanic lithosphere (Hanan and Graham, 1996; Geldmacher et al., 2011). On the whole, lavas of all ages erupted in the northern Ethiopia region have higher 3He/4He ratios (R/Ra=6±1; Graham, 2002). The elevated 3He/4He signature indicates these lavas are derived from mantle that contains higher concentrations of primitive 3He, though it is debated whether 3He resides in an incompletely degassed mantle (Zindler and Hart, 1986; Farley et al., 1992; Class and Goldstein, 2005; Day and Hilton, 2011) or a melt-depleted mantle (Parman et al., 2005; Albarède, 2008).

In contrast, early (40–32 Ma and 26–16 Ma) mafic volcanics in Turakana (N Kenya) consistently display less radiogenic 87Sr/86Sr (~0.7030), more radiogenic 206Pb/204Pb (up to 20.4) and lower 3He/4He (R/Ra=6–8) (Furman and Graham, 1994, 1998; Class et al., 2005; Furman et al., 2006b; Pik et al., 2006) than Ethiopian flood basalts (Fig. 2). These high 206Pb/204Pb, low 87Sr/86Sr, and low 3He/4He values are similar to the HIMU mantle component (Hoffman and White, 1982; Graham et al., 1992). Eocene and Oligocene mafic lavas from Turakana and neighboring Huri Hills were produced after continental rifting was well-established in the region. Compared to older volcanism in the region, these young lavas have markedly lower 206Pb/204Pb values (Class et al., 1994; Furman et al., 2004) that approach those observed in
northern Ethiopia. However, the $^3$He/$^4$He values of Kenya lavas never exceed 7 Ra (Furman et al., 2006b; Locke et al., 2008) and $^{87}$Sr/$^{86}$Sr values remain generally less radiogenic ($0.7030–0.7032$) than recorded in lavas to the north.

In the Western Rift, Kivu and Rungwe lavas are significantly younger ($<10$ Ma) than pre-rift NW Ethiopian flood basaltic and Turkanic lavas. Isotopically, mafic Kivu lavas have $^{87}$Sr/$^{86}$Sr = $0.7031–0.7052$ and $^{143}$Nd/$^{144}$Nd = $0.51303–0.51266$, which extend to more radiogenic values than either NW Ethiopia or Turkanic lavas (Furman and Graham, 1999). Rungwe lavas have $^{87}$Sr/$^{86}$Sr = $0.7045–0.7057$, $^{143}$Nd/$^{144}$Nd = $0.5124–0.5126$, and $^{206}$Pb/$^{204}$Pb = $17.65–18.90$ (Graham et al., 1995) as well as $^3$He/$^4$He that range from MORB-like ($7–9$ Ra; Graham et al., 1995) to plume-like ($9–15$ Ra; Hilton et al., 2011). Major and trace element data demonstrate the metasomatized subcontinental lithospheric mantle (SCLM) contributes to melt generation in both Kivu and Rungwe (Furman and Graham, 1994, 1999; Furman, 1995; Graham et al., 1995).

2.4. Samples

We analyzed Re and Os concentrations and isotope systematics of ten mildly alkaline mafic lavas from NW Ethiopia and Turkanic, N. Kenya. The NW Ethiopia samples comprise three ~30 Ma high-Ti picrites (11.8–15.2 wt.% MgO) and three ~23 Ma shield basaltic lavas with ~10 Ma) than pre-rift NW Ethiopian lavas. These lavas have $^{87}$Sr/$^{86}$Sr = $0.7031$–$0.7057$, $^{143}$Nd/$^{144}$Nd = $0.5130–0.5126$, and $^{207}$Pb/$^{204}$Pb = $17.65–18.90$ (Graham et al., 1995) as well as $^3$He/$^4$He that range from MORB-like ($7–9$ Ra; Graham et al., 1995) to plume-like ($9–15$ Ra; Hilton et al., 2011). Major and trace element data demonstrate the metasomatized subcontinental lithospheric mantle (SCLM) contributes to melt generation in both Kivu and Rungwe (Furman and Graham, 1994, 1999; Furman, 1995; Graham et al., 1995).

3. Methods

Sample preparation and analysis for Re–Os analyses took place at the Department of Terrestrial Magnetism (DTM) at the Carnegie Institution of Washington. For each sample, rock chips generated with an alumina jaw crusher were finely powdered using an alumina shatter box. The powders were spiked with dilute $^{185}$Os spike (0.4755 ng/g) and a dilute $^{185}$Re spike (3 ng/g in 6 M HCl) and then dissolved using the sealed Carius tube technique described in detail by Shirley and Walker (1998) with subsequent modifications for solvent extraction (Cohen and Waters, 1996) as described in Heinonen et al. (2010). For each sample 1.4–1.5 g of powder were digested in a Carius tube using an aqua regia solution. The tube was then sealed and heated for ~24 h at 240 °C to dissolve the Os-bearing sulfides, metals, and spines. Os was removed from aqua regia solution using CCl$_4$ extracted from the CCl$_4$ using 8N HBr. The Os aliquot was dissolved CrO$_3$, 12N H$_2$SO$_4$ and microdistilled using a technique described by Roy-Barman and Allègre (1994). The Re-bearing aqua regia solution was dried and dissolved in 1N HCl. Re was separated for mass spectrometry using anion exchange chromatography. Os separations were loaded on fused Pt ribbons and analyzed using a Thermo-Fisher Triton TIMS operated in negative ion mode. Re separations were diluted using 0.5 M HNO$_3$ and analyzed using a VG P-54 IC-MS. Os isotope ratios were corrected for mass fractionation using $^{192}$Os/$^{188}$Os = $3.082614$. In order to monitor analytical reproducibility, the DTM in-house Os standard was measured six times and yielded $^{187}$Os/$^{188}$Os = $0.17371 ± 0.00024$ (2σ). Os blanks ranged from 1 to 5 pg while Re blanks were 5 pg.

The same powders used for Re-Os analyses were also used to obtain $^{87}$Sr/$^{86}$Sr isotope values on the NW Ethiopia picrites. Nd–Pb–Hf isotopic determinations were performed at San Diego State University (SDSU) and repeated at DTM. SDSU techniques for Nd, Pb, and Hf separation using anion exchange chromatography are described in Hanan et al. (2004, 2008). At SDSU, Hf isotope sample data were normalized to $^{176}$Hf/$^{177}$Hf = $0.7235$ and the Nd data to $^{146}$Nd/$^{144}$Nd = $0.7219$ to correct for in-run fractionation. The Nd and Hf isotope ratios were corrected for instrumental machine bias and drift by applying a discrimination factor determined by sample-standard bracketing (every 2 samples) using the SDSU Ames Nd standard with $^{143}$Nd/$^{144}$Nd = $0.512130$ and the JMC 475 Hf standard with $^{176}$Hf/$^{177}$Hf = $0.282160$. The measured value of La Jolla Nd at SDSU = $0.511844$. The in-run mass discrimination for Pb was determined with NIST SRM 987 Ti where $^{207}$Pb/$^{205}$Pb = 2.3889 and NIST SRM 981 (Todt et al., 1996) using the method of White et al. (2000). The Pb isotope ratios were corrected for machine bias and drift by sample-standard bracketing, after correction for fractionation, with NIST SRM 981 (every two samples). The external reproducibility for $^{143}$Nd/$^{144}$Nd and $^{176}$Hf/$^{177}$Hf is 3 and 4 ppm, respectively, and for $^{206}$Pb/$^{204}$Pb, $^{207}$Pb/$^{204}$Pb, and $^{208}$Pb/$^{204}$Pb are 65, 58, and 130 ppm, respectively. Total procedural blanks were ~25 pg Hf, <90 pg Pb, <200 pg Nd. No blank corrections were applied to the data because they were insignificant.

At DTM, Pb, Sm, Nd, Lu, Hf, and Pb isotopic abundances were determined by isotope dilution (described by Carlson et al., 2006). Nd was measured by TIMS analysis using a Thermo-Fischer Triton, and the remaining elements were measured using a Nu Plasma HR ICP-MS. Nd was measured statically using double Re filaments and corrected for fractionation using $^{146}$Nd/$^{144}$Nd = $0.7219$. The data are reported relative to the standard JNd (143Nd/$^{144}$Nd = 0.512117) that gave an average $^{143}$Nd/$^{144}$Nd = $0.512181 ± 0.000007$ (2σ; n = 3) during the course of the run. For Pb and Lu, instrument fractionation was estimated by normalizing to bracketing standard runs. Sm measurements were monitored for Gd and Nd interferences, and mass fractionation was internally corrected to $^{147}$Sm/$^{152}$Sm = 0.56081. Hf isotopic analyses were monitored for W, Ta, Yb, and Lu interferences and corrected where necessary. The data are fractionation corrected to $^{176}$Hf/$^{177}$Hf = $0.7215$ and reported relative to JMC-475 ($^{176}$Hf/$^{177}$Hf = 0.282160) that gave an average $^{176}$Hf/$^{177}$Hf = $0.282142 ± 0.000007$ (2σ; n = 3). For Pb, mass fractionation was monitored by comparing bracketing runs of NBS 981 standard values to values reported by Todt et al. (1996). Standard values $^{206}$Pb/$^{204}$Pb = $17.15 ± 0.014$ (2σ), $^{207}$Pb/$^{204}$Pb = $15.819 ± 0.014$, and $^{208}$Pb/$^{204}$Pb = $37.736 ± 0.036$.

Strontium chemistry and analysis were performed at DTM using current separation techniques described by Heinonen et al. (2010). Samples were measured using TIMS, and the data were corrected for instrument mass fractionation using $^{88}$Sr/$^{86}$Sr = 0.1194. All data are reported relative to NBS 897 standard (accepted $^{87}$Sr/$^{86}$Sr = 0.71025). Two in-run measurements of NBS 987 yielded $^{87}$Sr/$^{86}$Sr = $0.710241 ± 0.000025$ (2σ).

4. Results

NW Ethiopia and Turkanic picrites have distinctive non-overlapping Hf isotopic signatures (Table 1; Fig. 3). The Hf–Nd isotopic compositions of NW Ethiopia picrites plot consistently just above the mantle array near the composition of the “C” mantle component (Fig. 3). Lavas from the Arabian Peninsula also plot near “C” but include less radiogenic compositions that likely reflect lithospheric contamination
Table 1
Summary of age-corrected initial isotopic data for lavas in this study. Sources for previously reported data: Furman (1995); Furman and Graham (1999); Kieffer et al. (2004); Furman et al. (2006b).

<table>
<thead>
<tr>
<th>NW Ethiopian Plateau</th>
<th>Rock type</th>
<th>Age (Ma)</th>
<th>MgO (wt.%)</th>
<th>Os (ppb)</th>
<th>Re (ppb)</th>
<th>$^{187}\text{Re} / ^{188}\text{Os}$ (m)</th>
<th>$^{187}\text{Os} / ^{188}\text{Os}$ (m)</th>
<th>$^{187}\text{Os} / ^{188}\text{Os}$ (i)</th>
<th>$^{87}\text{Sr} / ^{86}\text{Sr}$ (i)</th>
<th>$^{143}\text{Nd} / ^{144}\text{Nd}$ (i)</th>
<th>$^{176}\text{Hf} / ^{177}\text{Hf}$ (i)</th>
<th>$^{206}\text{Pb} / ^{204}\text{Pb}$ (i)</th>
<th>$^{207}\text{Pb} / ^{204}\text{Pb}$ (i)</th>
<th>$^{208}\text{Pb} / ^{204}\text{Pb}$ (i)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eth 232</td>
<td>Simien Shield Basalt</td>
<td>~18.7$^a$</td>
<td>10.9$^a$</td>
<td>0.009</td>
<td>0.133</td>
<td>69.89</td>
<td>0.2563</td>
<td>0.2214</td>
<td>0.70361</td>
<td>0.512871$^a$</td>
<td>~18.987$^a$</td>
<td>15.583$^a$</td>
<td>38.716$^a$</td>
<td></td>
</tr>
<tr>
<td>Eth 253</td>
<td>Choke Basalt</td>
<td>23$^a$</td>
<td>–</td>
<td>0.009</td>
<td>0.232</td>
<td>120.42</td>
<td>0.1700</td>
<td>0.1239</td>
<td>0.70338$^a$</td>
<td>0.512845$^a$</td>
<td>~19.221$^a$</td>
<td>15.692$^a$</td>
<td>39.343$^a$</td>
<td></td>
</tr>
<tr>
<td>Eth 267</td>
<td>Guguftu Basalt</td>
<td>23$^a$</td>
<td>7.74$^a$</td>
<td>0.003</td>
<td>0.504</td>
<td>1048.19</td>
<td>0.8284</td>
<td>0.4266</td>
<td>0.70375$^a$</td>
<td>0.512819$^a$</td>
<td>~18.968</td>
<td>15.577</td>
<td>38.672</td>
<td></td>
</tr>
<tr>
<td>Eth 643</td>
<td>Lalibela Picrite</td>
<td>~30$^a$</td>
<td>11.80$^a$</td>
<td>0.579</td>
<td>0.287</td>
<td>2.39</td>
<td>0.1323</td>
<td>0.1311</td>
<td>0.70426</td>
<td>0.512909</td>
<td>~18.968</td>
<td>15.577</td>
<td>38.672</td>
<td></td>
</tr>
<tr>
<td>Eth 643 (SDSU)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Eth 673</td>
<td>Sekota-Korma Rd Picrite</td>
<td>~30$^a$</td>
<td>12.80$^a$</td>
<td>0.864</td>
<td>0.325</td>
<td>1.81</td>
<td>0.1263</td>
<td>0.1254</td>
<td>0.70391</td>
<td>0.512892</td>
<td>~18.968</td>
<td>15.577</td>
<td>38.672</td>
<td></td>
</tr>
<tr>
<td>Eth 746</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Turkana (N Kenya)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>JCR 7</td>
<td>Jarigole Picrite</td>
<td>20–23$^b$</td>
<td>14.34$^b$</td>
<td>0.510</td>
<td>0.350</td>
<td>3.32</td>
<td>0.1493</td>
<td>0.1480</td>
<td>0.70324$^b$</td>
<td>0.512810$^b$</td>
<td>~20.33$^b$</td>
<td>15.72$^b$</td>
<td>39.92$^b$</td>
<td></td>
</tr>
<tr>
<td>JRG 9</td>
<td>Jarigole Basalt</td>
<td>20–23$^b$</td>
<td>6.33$^b$</td>
<td>0.022</td>
<td>0.682</td>
<td>148.33</td>
<td>0.2039</td>
<td>0.1470</td>
<td>0.70300</td>
<td>0.512794</td>
<td>~20.326</td>
<td>15.723</td>
<td>39.907$^b$</td>
<td></td>
</tr>
<tr>
<td>JRG 10</td>
<td>Jarigole Picrite</td>
<td>20–23$^b$</td>
<td>22.37$^b$</td>
<td>0.893</td>
<td>0.181</td>
<td>0.98</td>
<td>0.1478</td>
<td>0.1474</td>
<td>0.70300$^b$</td>
<td>0.512819$^b$</td>
<td>~20.319</td>
<td>15.70$^b$</td>
<td>39.76$^b$</td>
<td></td>
</tr>
<tr>
<td>JRG 12</td>
<td>Jarigole Picrite</td>
<td>20–23$^b$</td>
<td>28.65$^b$</td>
<td>1.190</td>
<td>0.181</td>
<td>0.74</td>
<td>0.1455</td>
<td>0.1452</td>
<td>0.70320$^b$</td>
<td>0.512754$^b$</td>
<td>~20.38$^b$</td>
<td>15.73$^b$</td>
<td>39.80$^b$</td>
<td></td>
</tr>
<tr>
<td>Kivu</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ZSC1</td>
<td>Tshibinda Basalt</td>
<td>0.2$^c$</td>
<td>9.10$^c$</td>
<td>0.084</td>
<td>0.319</td>
<td>18.59</td>
<td>0.2332</td>
<td>0.2328</td>
<td>0.70511$^c$</td>
<td>0.512669$^c$</td>
<td>~18.968</td>
<td>15.577</td>
<td>38.672</td>
<td></td>
</tr>
<tr>
<td>Z3D</td>
<td>Tshibinda Basalt</td>
<td>0.2c</td>
<td>8.74c</td>
<td>0.033</td>
<td>0.325</td>
<td>49.29</td>
<td>0.3175</td>
<td>0.3162</td>
<td>0.70511$^c$</td>
<td>0.512659$^c$</td>
<td>~20.326</td>
<td>15.723</td>
<td>39.907$^b$</td>
<td></td>
</tr>
<tr>
<td>R3L</td>
<td>Bugarama Basalt</td>
<td>8.8$^c$</td>
<td>8.93$^c$</td>
<td>0.026</td>
<td>0.288</td>
<td>57.73</td>
<td>0.2086</td>
<td>0.2009</td>
<td>0.70314$^c$</td>
<td>0.513026$^c$</td>
<td>~20.326</td>
<td>15.723</td>
<td>39.907$^b$</td>
<td></td>
</tr>
<tr>
<td>RSN</td>
<td>Gisakura Basalt</td>
<td>8.6$^c$</td>
<td>9.09$^c$</td>
<td>0.087</td>
<td>0.530</td>
<td>29.49</td>
<td>0.1639</td>
<td>0.1632</td>
<td>0.70314$^c$</td>
<td>0.512925$^c$</td>
<td>~20.326</td>
<td>15.723</td>
<td>39.907$^b$</td>
<td></td>
</tr>
<tr>
<td>Rungwe</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>RUN-27</td>
<td>Porotos Basalt</td>
<td>5–10$^d$</td>
<td>9.55$^d$</td>
<td>0.019</td>
<td>0.198</td>
<td>50.41</td>
<td>0.1644</td>
<td>0.1615</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td></td>
</tr>
<tr>
<td>RUN-23</td>
<td>Kiejo Picrite</td>
<td>&lt;1.0$^d$</td>
<td>18.87$^d$</td>
<td>0.020</td>
<td>0.252</td>
<td>62.91</td>
<td>0.2700</td>
<td>0.2663</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td></td>
</tr>
<tr>
<td>T10G</td>
<td>Kiejo Nephelinite</td>
<td>&lt;1.0$^d$</td>
<td>8.64$^d$</td>
<td>0.003</td>
<td>0.409</td>
<td>797.40</td>
<td>0.9216</td>
<td>0.8751</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td></td>
</tr>
<tr>
<td>T15B</td>
<td>Kiejo Nephelinite</td>
<td>&lt;1.0$^d$</td>
<td>9.03$^d$</td>
<td>0.007</td>
<td>0.452</td>
<td>304.99</td>
<td>0.3610</td>
<td>0.3610</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td></td>
</tr>
</tbody>
</table>

$^a$ Data reported by Kieffer et al. (2004).
$^b$ Data reported by Furman et al. (2006b).
$^c$ Data reported by Furman and Graham (1999).
$^d$ Data reported by Furman (1995).
(Bertrand et al., 2003). In contrast, the Turkana picrites (εHf = 3.18–4.14; from Locke et al., 2008) plot below mantle array values, similar to other HIMU OIB suites (e.g. St. Helena, Rapa, Tubuai, and Ravaa; Fig. 3).

NW Ethiopia and Turkana picrites have much higher Os contents (0.579–1.190 ppb) than the basalts from the same localities (0.003–0.022 ppb; Table 1). Rhenium contents show a much smaller range, with concentrations in the picrites (0.133 to 0.350 ppb) only slightly lower than those of the basalts (0.504 to 0.682 ppb). Other picritic and basaltic lavas from NW Ethiopia show a similar range of concentrations in both Os (0.05–3.92 ppb) and Re (0.03–0.53 ppb) (Rogers et al., 2010).

Re-Os isotope evolution lines were calculated for 30 Ma and 23 Ma lavas in order to approximate age-corrected initial isotopic compositions of their mantle sources (Fig. 4). Isotopically, picrites have much lower 187Re/188Os than the basalts (0.74–3.32 versus 69.9–1048, respectively). Os isotopes values for 30 Ma HT2 picrites (Rogers et al., 2010) plot amongst our new NW Ethiopia data and fall very near the 30 Ma isotope evolution line. The 23 Ma Choke volcano basalt from NW Ethiopia (Eth 253) falls on the less radiogenic 23 Ma isotope evolution line. The 30 Ma and 23 Ma isotope evolution lines for the NW Ethiopia lavas project to the same initial 187Os/188Os (i) for the NW Ethiopia picrites and the Choke volcano basalt. Eth 267 is not shown due to its anomalously high isotopic ratios (187Re/188Os = 1049; 187Os/188Os = 0.8284). Kivu and Rungwe magmatic lavas, Toro Ankole kamafugites (Rosenthal), and additional NW Ethiopia picrite data (Rogers et al., 2010) are plotted for comparison (B) Magnified look at the 187Re/188Os vs. 187Os/188Os relationship between various lavas. (C) 187Os/188Os vs. Re/Os plot illustrating Re enrichment (or lack thereof) compared to Os.
Continental crust often contributes to volcanism in the EARS (e.g. Pik et al., 1999; Baker et al., 2000; Kieffer et al., 2004) and its influence in pre-rift lavas from NW Ethiopia and Turkana must be evaluated. Continental crust has high Re and low Os abundances, leading to high Re/Os and a time-integrated radiogenic $^{187}\text{Os}/^{188}\text{Os}$ signature (0.70–1.20; Esser and Turekian, 1993; Horan et al., 1995; Saal et al., 1998b) relative to the mantle. NW Ethiopia basalts Eth 232 and Eth 267 as well as the Turkana picrites and basalts have a more radiogenic initial $^{187}\text{Os}/^{188}\text{Os}$ signature (0.1450–0.1483) than the Ethiopian lavas, consistent with global OIB, and specifically overlapping the range for HIMU basalts (Fig. 5).

5. Discussion

5.1. Assimilation of continental crust

Continental crust often contributes to volcanism in the EARS (e.g. Pik et al., 1999; Baker et al., 2000; Kieffer et al., 2004) and its influence in pre-rift lavas from NW Ethiopia and Turkana must be evaluated. Continental crust has high Re and low Os abundances, leading to high Re/Os and a time-integrated radiogenic $^{187}\text{Os}/^{188}\text{Os}$ signature (0.70–1.20; Esser and Turekian, 1993; Horan et al., 1995; Saal et al., 1998b) relative to the mantle. NW Ethiopia basalts Eth 232 and Eth 267 as well as the Turkana picrites and basalts have a more radiogenic initial $^{187}\text{Os}/^{188}\text{Os}$ signature (0.1450–0.1483) than the Ethiopian lavas, consistent with global OIB, and specifically overlapping the range for HIMU basalts (Fig. 5).

We evaluated the effect of crustal contamination explicitly by modeling mixing between “C”–like Afar plume material and continental crust. We used the isotopic composition of the Afar plume (Rooney et al., 2012) as our “C” mixing component. The isotopic composition of the crust beneath the EARS is poorly constrained. Mixing a non-unique continental crust component (see Supplementary Material for details on mixing parameters) with a “C”–like mantle component (mixing line 1; Fig. 6) demonstrates that adding ~15% melt of continental crust to a “C”–like Afar plume can generate radiogenic Os signatures similar to those measured in Eth 232 ($^{187}\text{Os}/^{188}\text{Os} = 0.2214$). A mafic crustal component would have a higher Os concentration and therefore less material would be required to generate the $^{187}\text{Os}/^{188}\text{Os}$ signature (Lassiter and Luhr, 2001). Likewise, bulk assimilation necessitates the incorporation of less (~5%) continental crust to generate radiogenic $^{187}\text{Os}/^{188}\text{Os}$ signatures in mafic lavas. Because Os concentrations of old (e.g. Proterozoic) crust (0.06–0.1 ppb; Saal et al., 1998b) and these basaltic melts (~0.01 ppb) differ by more than an order of magnitude, small Os contributions from radiogenic crustal material will have a large effect on the isotopic composition of the lavas without significantly affecting other major and trace elements. For example, crustal contamination has only a slight effect on La/Nb (0.53) in Eth 232 (e.g. Kieffer et al., 2004) and essentially no effect on concentrations, and hence isotopic compositions, of Sr and Nd (Fig. 6). As a result, the Os isotopic signatures of Ethiopian basalts Eth 232 and 267 and two Rungwe nephelinites T10G and T15B cannot be used to make interpretations about the volcanic source materials. The picrites from NW Ethiopia and Turkana have much higher Os contents (~0.05 ppb) that are not susceptible to minor crustal contamination and therefore can provide information regarding mantle source characteristics.

5.2. Geochemical characteristics of EARS mantle sources

The number of plumes contributing to volcanism in the EARS is debated, ranging from one runny plume (Ebinger and Sleep, 1998) to two separate plumes that may or may not be connected at depth (e.g. George et al., 1998; Furman et al., 2006b). In the two-plume model, Oligocene–Miocene volcanism is tied to low velocity features above the mantle transition zone that are currently located beneath the Afar region (N Ethiopia) and the Tanzanian craton (Nyblade et al., 2000a, 2000b; Weeraratne et al., 2003; Benoit et al., 2006); these features are referred to as the Afar and Kenya plumes, respectively. Pre-rift volcanism associated with these regions of mantle upwelling are isotopically distinct. Our new Os isotopic data are consistent with inferences based on Sr, Nd, Pb, Hf and He isotopic data (Figs. 2–4). Specifically, Oligocene–early Miocene Turkana lavas associated with the Kenya plume contain a HIMU component that is not present in Afar plume-derived flood basalts and shield lavas from NW Ethiopia; the NW Ethiopia lavas appear to have been derived from mantles sources dominated by the “C” component.

Several scenarios have been proposed for the generation of the HIMU mantle component. Traditional models invoke ancient (~2 Ga) recycled oceanic crust (e.g. Hofmann and White, 1982; Graham et al., 1992; Hauri and Hart, 1993) or metasomatized SCLM (Shaw et al., 2003). There is increasing evidence to suggest carbonate metasomatism of mantle materials plays a role in generating HIMU signatures (e.g., Saal et al., 1998a; Hoernle et al., 2002; Pilet et al., 2005, 2008; Dasgupta et al., 2006, 2007). In the following discussion, we evaluate the influence of the “C” mantle component, ancient recycled oceanic crust, the SCLM, and the role of metasomatism in the context of new Re–Os isotopic data in order to determine the origin of the HIMU signature associated with early Turkana lavas.

5.2.1. Contributions of “C” to mafic volcanism in Ethiopia

The “C” mantle component has been recognized recently as playing an important role in volcanism throughout Ethiopia (e.g. Meshesha and Shinjo, 2010; Rogers et al., 2010; Rooney et al., 2012). However, the $^{187}\text{Os}/^{188}\text{Os}$ ratio for “C” is not well constrained. NW Ethiopian picrites have $^{187}\text{Os}/^{188}\text{Os}$ values (0.126–0.131, this study; 0.125–0.133, Rogers et al., 2010) that overlap with and extend to values lower than the average composition of enriched plumes (0.130–0.135; Shirley and Walker, 1998). Rogers et al. (2010) proposed that the roughly chondritic $^{187}\text{Os}/^{188}\text{Os}$ signature recorded by the picrites could be inherent to the “C” component. However, if we assume the “C” component has an isotopic signature similar to “enriched plumes” defined by Shirley and Walker (1998), then the unradiogenic $^{187}\text{Os}/^{188}\text{Os}$ values suggest that the NW Ethiopian picrites may contain minor contributions from unradiogenic SCLM (0.104–0.122; Chesley et al., 1999; Burton et al., 2000; Reisberg et al., 2004). The addition of small amounts of SCLM has been shown to affect the isotopic composition of plume-derived lavas while leaving the trace element abundances of the lavas relatively untouched (e.g. Hanan et al., 2008). In order to estimate the amount of
possible SCLM added to Afar plume material, we mixed "C"-like Afar plume material (defined by Rooney et al., 2012) with SCLM material similar to the composition of mantle xenoliths found beneath the East African Rift System (mixing line 2; Fig. 6). The \( \text{\(^{187}\text{Os/^{188}Os}\)} \) signature of the Afar plume is unknown, we conservatively estimated it as \( \text{\(^{187}\text{Os/^{188}Os} = 0.130}\)\). When using the composition of the Afar plume narrowly defined by Rooney et al. (2012), isotope mixing models the NW Ethiopian picrites may be the result of up to 20% addition of SCLM material to the "C"-like Afar component. The error bars in Fig. 6 demonstrate that the "C" component was originally defined with a large isotopic range (Hanan and Graham, 1996). As a consequence, the mixing estimate is highly dependent on the \( \text{\(^{187}\text{Os/^{188}Os}\)} \) signature of the "C" component and the heterogeneity in the Afar plume itself and therefore must be considered a first order approximation. If the mildly unradiogenic \( \text{\(^{187}\text{Os/^{188}Os}\)} \) signature is inherent to the "C" component, then NW Ethiopian picrites are melts derived solely from the Afar plume and do not contain contributions from the SCLM.

The role of SCLM in the NW Ethiopian flood basalts and picrites is subject to debate. Kieffer et al. (2004) and Pik et al. (1998, 1999) attribute the presence of minor hydrous mineral phases, the overall trace element signature, and Sr – Nd – Pb isotopic signatures of the NW Ethiopian high-Ti picrites (a.k.a. HT2 lavas) as reflecting direct melting of the upwelling plume. High Nb/La values (1.2 – 1.6), high MgO and FeO contents, low \( \text{Al}_2\text{O}_3/\text{TiO}_2 \) and depleted HREE abundances further suggest that HT2 lavas were derived from a garnet-bearing sublithospheric source (Kieffer et al., 2004). However, high TiO\(_2\) and elevated Fe/Mn values indicate contributions from pyroxenite melts (Rogers et al., 2010). Similarly, trace element modeling suggests recently (i.e. synchronously with the arrival of the Afar plume) metasomatized peridotite contributed significantly to Ethiopian HT2 lavas (Beccaluva et al., 2009).

Basalt from the 23 Ma Choke shield volcano in NW Ethiopia (Eth 253) has \( \text{\(^{206}\text{Pb/^{204}Pb}\)} \) (19.22; Kieffer et al., 2004) and \( \text{\(^{187}\text{Os/^{188}Os}\)} \) (0.129) signatures similar to NW Ethiopia picrites (Fig. 5). Kieffer et al. (2004) suggest that Choke lavas are derived directly from the upwelling asthenosphere. This sample has a very low Os concentration (9 ppt), so interaction with the SCLM is unlikely. As such, the Choke lavas are derived from the upwelling asthenosphere. This suggests that the Choke sample to the 30 Ma NW Ethiopia HT2 lavas are derived from similar source(s).

5.2.2. HIMU signature in early Kenya plume lavas

Unlike Afar plume lavas, Miocene Turkana picrites have radiogenic \( \text{\(^{176}\text{Hf/^{177}Hf}\)} \), \( \text{\(^{187}\text{Os/^{188}Os}\)} \), and \( \text{\(^{206}\text{Pb/^{204}Pb}\)} \) values that overlap HIMU lavas from St. Helena, the South Pacific, and the Canary Islands (Figs. 3 and 5). As is the case for these OIB suites, the HIMU signature manifest in Kenya plume volcanism has been proposed to reside either in the deep mantle (e.g., Chesley et al., 1999; Furman et al., 2006b; Aulbach et al., 2008) or the SCLM (e.g. Stewart and Rogers, 1996; George and Rogers, 2002).

The origin of the HIMU component in OIB is attributed traditionally to the incorporation of ~2 Ga recycled oceanic crust (e.g. Hofmann and...
White, 1982; Graham et al., 1992; Hauri and Hart, 1993). Mixing 2 Ga recycled oceanic crust (Os = 0.1 ppb; \( ^{187}\text{Os}/^{188}\text{Os} = 1.78 \)) and Shirey and Walker, 1998) with “C′-like Afar plume material (Rooney et al., 2012) generates Os–Pb isotopic signatures that roughly approach those of the Turkana picrites (mixing line 3; Fig. 6). We can more closely generate isotopic compositions Turkana picrites by mixing ~20% 2 Ga hydrothermally altered oceanic crust (calculated using Stracke et al., 2003; details in Supplementary Material) with 80% Afar plume material (mixing line 4; Fig. 6). However, the best fit mixing model combines 20–30% 1.7 Ga oceanic crust with Afar plume material (mixing line 5; Fig. 6). This calculation demonstrates that combining ancient subducted oceanic crust with upwelling mantle plume material is a viable model for generating the observed HIMU signature in Turkana lavas. This interpretation is broadly consistent with numerical models by Brandenburg et al. (2008) that illustrate it is possible to generate MORB, EMI, and HIMU by melting subducted oceanic crust. Although our models are non-unique they suggest that ancient recycled oceanic crust is a necessary addition to the “C" mantle reservoir in generating the radiogenic \( ^{187}\text{Os}/^{188}\text{Os} \) signature observed in Turkana lavas.

The HIMU signature in Turkana lavas younger than 4 Ma is not pronounced (Furman et al., 2004), suggesting that the HIMU component could be located in the SCLM, and that the SCLM has progressively been removed by melt generation and/or extension during the Miocene. Afro-Arabian SCLM xenoliths have highly variable \( ^{87}\text{Sr}/^{86}\text{Sr} \) (0.7029–0.8306; Cohen et al., 1984; Aulbach et al., 2008, 2011) and \( ^{206}\text{Pb}/^{204}\text{Pb} \) (15.26–20.60; Cohen et al., 1984; Shaw et al., 2007) demonstrating that portions of the SCLM could potentially be the source for HIMU Turkana lavas. The Tanzanian lherzolite and harzburgite xenoliths with the most radiogenic \( ^{206}\text{Pb}/^{204}\text{Pb} \) (up to 20.60; Cohen et al., 1984) have unradiogenic to mildly radiogenic \( ^{187}\text{Os}/^{188}\text{Os} \) (0.104–0.122; Burton et al., 2000). Mixing this SCLM material (SCLM 2) with “C′-like Afar plume material can generate the radiogenic \( ^{206}\text{Pb}/^{204}\text{Pb} \) signature observed in Turkana lavas but cannot generate the radiogenic \( ^{187}\text{Os}/^{188}\text{Os} \) signature (mixing line 6; Fig. 6). Dunite xenoliths from the same localities have radiogenic \( ^{187}\text{Os}/^{188}\text{Os} \) (0.161–0.257; Burton et al., 2000) but low Os concentrations (\(<0.008 \) ppb) make the dunites an unlikely source for the Turkana picrites (Os \(<0.5–1.2 \) ppb). We conclude that the HIMU component in this region is likely generated by incorporating ancient subducted oceanic crust in upwelling “C′-like mantle material.

In order to test the viability of our Os–Pb models, we applied the same modeling parameters to Sr, Nd, and Hf isotopic ratios (Fig. 6) with focus placed on mixing between Afar plume material (Rooney et al., 2012) and hydrothermally altered oceanic crust (calculated using Stracke et al., 2003; see Supplementary Material). The Sr and Hf isotopic signatures of HIMU-like Turkana picrites can be approximated by mixing ~30% altered recycled oceanic crust with 70% “C′ mantle material. In order to reproduce the Nd isotopic compositions of the Turkana lavas (mixing line 5; Fig. 6), slight modification of the estimated mobility coefficients for both Nd and Hf is required (from 22% to 18% and 22% to 21%, respectively; Supplementary Materials). These changes are within reason, especially with respect to Nd because experimental studies demonstrate that it is difficult to estimate Sm and Nd mobility during subduction processes (Kogiso et al., 1997; Ayers, 1998; Stalder et al., 1998). This calculation demonstrates that a mixture of “C′-like Afar-plume material and altered ancient subducted oceanic crust can also reproduce the HIMU-like Sr, Nd, and Hf isotopic signatures of the Turkana picrites (Fig. 6).

5.2.3. The role of metasomatized SCLM in Western and Kenya rift volcanism

Metasomatized SCLM often melts to produce highly alkaline lavas (e.g. Wyoming lamproites – Mirnejad and Bell, 2006; South African melilitites – Janney et al., 2002; Brazilian kamafugites – Carlson et al., 2007). In the EARS, SCLM xenoliths from the Sidamo region (S Ethiopia; Bedini et al., 1997; Bedini and Bodinier, 1999; Lorand et al., 2003; Reisberg et al., 2004) and various localities adjacent to the eastern side of the Tanzanian craton (Cohen et al., 1984; Chesley et al., 1999; Burton et al., 2000; Becker et al., 2006; Aulbach et al., 2008, 2011), and Toro Ankole (Lloyd et al., 1985) demonstrate that the lithosphere has undergone variable degrees of silicate and/or carbonate metasomatism. Consequently, a significant proportion of Western Rift volcanism is derived in part from hydrous and/or carbonate metasomatized SCLM (Rogers et al., 1992, 1998; Furman, 1995; Furman and Graham, 1999; Rosenthal et al., 2009), resulting in moderate to highly alkaline lavas. Potassic kamafugites from Toro Ankole (Uganda) are partial melting products of plagiogiphite and clinopyroxene-bearing mantle xenoliths recovered in the same region (Lloyd et al., 1985). These highly alkaline lavas display a general enrichment in LILE with strong K depletions (Fig. 7) and record moderate Re/Os ratios that generated a range of radiogenic \( ^{187}\text{Os}/^{188}\text{Os} \) signatures (0.1476–0.4974; Rosenthal et al., 2009).

Some Kivu basalts (Rwanda-Burundi border; Furman and Graham, 1999) and most Rungwe basalts and nephelinites (S Tanzania; Furman, 1995) were derived from metasomatized SCLM and span a range of Os isotopic signatures (0.1632–0.3162 and 0.1615–0.3610, respectively) comparable to those of Toro Ankole lavas (Rosenthal et al., 2009) but more radiogenic than Turkana lavas (Fig. 4). Kivu and Rungwe lavas have low Os abundances (0.03–0.08 ppb and \(<0.02 \) ppb, respectively), which suggests the \( ^{187}\text{Os}/^{188}\text{Os} \) signature is easily influenced by high Re/Os associated with metasomatized SCLM.

The isotopic range for Toro Ankole lavas derived from metasomatized SCLM \( ^{187}\text{Os}/^{188}\text{Os} = 0.1476–0.4974; \) Rosenthal et al., 2009).
lavas (Rosenthal et al., 2009) with a fertile and depleted peridotite abundance (e.g. Hanan et al., 2008), mixing various Toro Ankole picrites, signature of a melt without noticeably changing the trace element budget of metasomatized SCLM to Turkana lavas. First, Turkana lavas do not display strong LILE enrichments or large K depletions (Furman et al., 2006b) associated with melting a hydrous source similar to that beneath Torok Ankole (Fig. 7). Second, though it has been demonstrated that small contributions of lithospheric material can influence the isotopic signature of a melt without noticeably changing the trace element abundances (e.g. Hanan et al., 2008), mixing various Toro Ankole lavas (Rosenthal et al., 2009) with a fertile and depleted peridotite melt (187Os/188Os = 0.127–0.132, Os = 1.0 ppb) require a 30–70% contribution from metasomatized material to generate the 187Os/188Os signature recorded by Turkana lavas. This scenario is unreasonable because the trace element budget for Turkana lavas does not demonstrate significant contributions from a phlogopite or amphibole-bearing source. Third, contributions from metasomatized SCLM will also influence other isotopic systems (Carlson et al., 2007; Hanan et al., 2008). Metasomatism tends to enrich the SCLM in Rb and Sr, which generates radiogenic 87Sr/86Sr over time. Turkana lavas have high Sr concentrations (3057–3034 ppm), radiogenic 87Sr/86Sr (0.7046–0.7053), and unradiogenic 143Nd/144Nd (εNd = −0.1 to −4.7) (Rosenthal et al., 2009), quite unlike mafic Turkana lavas (Sr = 179–490 ppm, 87Sr/ 86Sr = 0.7030–0.7032, εNd = +2.61±0.33; Furman et al., 2006b). Therefore, Western Rift volcanism was derived from a compositional and lithological source distinct from Turkana.

5.3. A new model for the dynamics of the EARS

Perhaps the strongest evidence to support the presence of one or more mantle plumes below the EARS is the lack of contribution from shallow, depleted MORB mantle in the source material for pre-rift volcanism (e.g. Schilling, 1973; Richards et al., 1989; Vidal et al., 1991; Baker et al., 1996b; Kieffer et al., 2004; Furman et al., 2006a). If this episode of volcanism were not plume-driven, we would expect passive riftting to have allowed such shallow mantle to dominate the geochemistry of the pre-rift lavas. This study and previous studies have demonstrated this is not the case. Using picrites from NW Ethiopia and Turkana as proxies, it is clear that the Afar and Kenya mantle plume volcanic products are characterized by distinct isotopic compositions that require sampling of different mantle source components. New 187Os/188Os data support the hypothesis that incorporation of ancient (~2 Ga) recycled oceanic crust in a dominantly "C"-like source region can account for the different isotopic signatures in the volcanic products of the two regions.

Geochemical data alone cannot resolve the depth at which the recycled material was incorporated into the upwelling plume (i.e. near the core-mantle boundary, near the 660 km discontinuity, or by melting of subducted material fused to the base of the continental lithosphere). Geophysical data have shown that density variations in the deep and thermally perturbed mantle beneath southern Africa are caused by both thermal and geochemical heterogeneities (Ni et al., 2002; Simmons et al., 2007). The African superplume is therefore a viable source for compositionally diverse mantle material represented by the geochemical range observed in plume-derived lavas. Fluid dynamic modeling suggests that the oceanic crust is likely to be incorporated into thermally buoyant material at the core-mantle boundary. More specifically, these models demonstrate that classical thermal plume behavior, which is characterized by an outpouring of flood basalts from a plume head followed by lower-volume plume tail eruptions, is strongly modified when compositional variations (i.e. subducted oceanic crust) are incorporated. In contrast to thermal plumes, compositionally variable (i.e. thermochemical) plumes rising from a thermochemical boundary layer generally form broad, discontinuous or episodic structures (Davaille et al., 2003; Farnetani and Samuel, 2005; Lin and van Keken, 2005; Tan and Gurnis, 2007) that may suppress the formation of a plume head. This style of upwelling will lead to diminished volcanic productivity and may prevent the formation of a flood basalt province (Farnetani and Samuel, 2005). Such thermochemical upwellings may also stagnate at mid-mantle depths, shedding smaller plumes to the surface (Farnetani and Samuel, 2005; Lin and van Keken, 2005; Samuel and Bercovici, 2006).

We propose that the differences in eruptive behavior and geochemical signature of Afar and Kenya plume lavas are best explained by secondary mantle plumes derived from and reflective of chemical heterogeneities within the African superplume. Our conceptual model is based on the thermochemical nature of the lower mantle region below Africa and the dynamical consequences of thermochemical convection on the nature of rising plumes. When applying this model to geochemical and geophysical data from the EARS, the distinct geochemical signatures of the Afar and Kenya plumes can best be reconciled if the Afar plume is viewed as a dominantly thermal feature composed of the widely dispersed "C" mantle component while the Kenya plume is a thermochemical anomaly containing ancient chemically distinct recycled material mixed with "C" mantle material. We assume that recycled oceanic crustal material introduces the chemical heterogeneities in the lower mantle (in Christensen and Hofmann, 1994; Brandenburg et al., 2008). As this denser material is entrained into the upwelling plume it can combine with peridotitic mantle plume material, thus generating the Sr–Nd–Pb–Hf–Os isotopic signatures observed in mafic lavas from Turkana. We assume that the higher pressure phases in recycled oceanic crust remain generally denser than the ambient mantle. The well-known density inversion between the recycled oceanic crust-derived component and ambient mantle that occurs in the uppermost lower mantle (e.g., Faust and Knittle, 1996) should not strongly affect this model. A significant number of geodynamical studies show that this density inversion is not sufficient to render subduction into the lower mantle impossible (Christensen, 1988) except under extreme rheological variations (van Keken et al., 1996). Diffuse upwellings are more significantly affected (e.g., Weinstein, 1992; van Summeren et al., 2009) but focused mantle plume-like upwellings are not significantly hindered as was demonstrated in Lin and van Keken (2005), which explicitly took the suggested density inversion into account. The thermochemical interplay between recycled oceanic crust and peridotite provides an explanation for the lack of flood basalt province associated with the initial arrival of the Kenya plume. Recycled oceanic crust is chemically and isotopically distinct from the deep mantle, making it a likely candidate for the seismically-imaged chemically dense material in the African superplume (Ni et al., 2002; Simmons et al., 2007). In contrast, the Afar plume does not demonstrate isotopic signatures or eruptive behavior supporting the involvement of recycled oceanic crust but rather displays features characteristic of a thermal plume. As a result, this plume produced voluminous flood basalts geochemically consistent with melting "C"-component mantle containing high 3He/4He values.

A deeper connection between the Afar and Kenya plumes is inferred from the geophysical data, but cannot be demonstrated unequivocally. We propose that the two shallow mantle upwellings tap thermochemically distinct regions within the heterogeneous African superplume (Fig. 8). The regional geochemical and eruptive differences are predicted by dynamical models that demonstrate small secondary plumes can originate from a large heterogeneous mantle upwelling (Davaille et al., 2003). In this scenario, the eruptive behavior and isotopic composition of the Afar plume suggest the plume rises from either (1) a region of the superplume lacking compositionally dense material, or (2) a region in the superplume with excess compositional density. In the second scenario the amount of dense material beneath the Afar plume would be too great for the thermal buoyancy of the upwelling mantle and therefore would not be incorporated into the rising plume. In this case, the Afar plume would sample primarily ambient mantle overlying the dense material, and would, therefore, behave as a thermal plume (Farnetani, 1997; Lin and van Keken, 2006). Thermochemical
models indicate dense material (i.e. subducted oceanic crust) can form piles or accumulate in the D’ of the core–mantle boundary (e.g. Brandenburg et al., 2008), which may generate regions of excess compositional density. However, it is also possible to create regions with low concentrations of compositionally dense material. In either situation, the Afar plume would not contain significant amounts of dense recycled material and therefore would have the behavioral and chemical characteristics of a dominantly thermal plume. In contrast, the upwelling that initially manifests itself beneath Kenya appears to rise from a region of the superplume with a high compositional density (i.e., abundant subducted oceanic crust) that is almost equal to (but less than) the plume’s thermal buoyancy. This scenario allows for incorporation of dense recycled material into the plume conduit that will disrupt typical mantle plume development and eruptive behavior.

5.4. Implications for other superplumes

The thermochemical plume model used here may also provide insight into the spatial, temporal, and chemical volcanic patterns associated with the South Pacific superplume. Islands generated from the South Pacific superplume are characterized by mafic lavas with HIMU, EM-I and EM-II compositions. HIMU lavas from the Cook-Austral islands show strong evidence for involvement of recycled oceanic crust (Hauri and Hart, 1993). They also lack clear flood basalts/oceanic plateau provinces (Clouard and Bonneville, 2001). Because these islands do not show a clear spatio-temporal progression, geochemical models ranging from multiple mantle plumes (Chauvel et al., 1997) to a total absence of mantle plumes (McNutt et al., 1997) have been proposed to explain the observed volcanic distribution. Small plumes stemming from a larger mantle plume is not a new model. For example, Janney and Castillo (1998) proposed this mechanism to explain the volcanic distribution of the Darwin Rise seamounts. Furman et al. (2006b) first proposed that both the Afar and Kenya plumes as offshoots of the African superplume volcanism. Our model builds on the steming plume model and suggests that secondary plumes can be thermochemically distinct, thereby accounting for the chemical heterogeneity and age distribution of the volcanism associated with both the African and the South Pacific superplumes. Because entrained dense material can profoundly affect plume behavior, the age and timing of each volcanic pulse is dependent on the amount of recycled oceanic material incorporated into the upwelling plume. Such a process can account for the lack of clear age progression in the volcanic record, and it can generate compositionally diverse lavas from a single thermochemical plume.

6. Summary

Pre-rift picrites and basalts from NW Ethiopia and Turkana (N Kenya) have distinct Os–Hf isotopic signatures that cannot be attributed to decompression melting of SCLM or depleted upper mantle. Instead, the lavas record signatures of sublithospheric sources associated with mantle plume volcanism. Volcanism in these two regions is geochemically distinct: NW Ethiopia flood basalts (−30 Ma) have $^{187}$Os/$^{188}$Os = 0.1239–0.1311 and $^{176}$Hf/$^{177}$Hf = 0.28311–0.28315, consistent with derivation from the “C” mantle component. In contrast, Turkana picrites (23–16 Ma) have radiogenic $^{187}$Os/$^{188}$Os (0.1450–0.1483) consistent with contributions from a HIMU component that is not present in NW Ethiopia lavas. Two-component mixing models demonstrate the HIMU component is likely generated by mixing 1.7 Ga hydrothermally altered oceanic crust with “C” mantle material.

The geochemically distinct early lavas from NW Ethiopia and Turkana are associated with thermal upwellings known as the Afar and Kenya mantle plumes. We propose that the “C”-like Afar plume and the HIMU-like Kenya plume sample different portions of the thermochemically heterogeneous African superplume. The heterogeneity is generated through mingling of deep, incompletely degassed mantle containing young (<1 Ga) hydrothermally altered subducted oceanic crust with smaller amounts of anciently (1.7–2 Ga) recycled oceanic crust residing in the lower mantle. The different lithologies are responsible for the chemical differences between the Kenya and Afar mantle plumes and for their distinct eruptive styles.

Acknowledgments

We would like to thank Nick Arndt for sharing samples from NW Ethiopia, Mary Horan, Tim Mock, and Megan Pickard for their generous assistance during the analytical work, and Matthew Jackson, Keith Putirka, Cindy Ebinger, Nick Rogers, and Nick Arndt for comments on earlier versions of this work. Thorough reviews by two anonymous reviewers helped to clarify our thinking and strengthen the manuscript. This research is supported by the National Science Foundation grants EAR-0551913 to Peter van Keken and EAR-0551954 to Tanya Furman as well as the Department of Geosciences Paul D. Krynine Memorial Fund (Pennsylvania State University) to Wendy Nelson.

Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.chemgeo.2012.05.020.

References


