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Cenozoic extension in the Kenya Rift from low-temperature thermochronology: Links to diachronous spatiotemporal evolution of rifting in East Africa

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Abstract The cooling history of rift shoulders and the subsidence history of rift basins are cornerstones for reconstructing the morphotectonic evolution of extensional geodynamic provinces, assessing their role in paleoenvironmental changes and evaluating the resource potential of their basin fills. Our apatite fission track and zircon (U-Th)/He data from the Samburu Hills and the Elgeyo Escarpment in the northern and central sectors of the Kenya Rift indicate a broadly consistent thermal evolution of both regions. Results of thermal modeling support a three-phased thermal history since the early Paleocene. The first phase (~65–50 Ma) was characterized by rapid cooling of the rift shoulders and may be coeval with faulting and sedimentation in the Anza Rift basin, now located in the subsurface of the Turkana depression and areas to the east in northern Kenya. In the second phase, very slow cooling or slight reheating occurred between ~45 and 15 Ma as a result of either stable surface conditions, very slow exhumation, or subsidence. The third phase comprised renewed rapid cooling starting at ~15 Ma. This final cooling represents the most recent stage of rifting, which followed widespread flood-phonolite emplacement and has shaped the present-day landscape through rift shoulder uplift, faulting, basin filling, protracted volcanism, and erosion. When compared with thermochronologic and geologic data from other sectors of the East African Rift System, extension appears to be diachronous, spatially disparate, and partly overlapping, likely driven by interactions between mantle-driven processes and crustal heterogeneities, rather than the previously suggested north–south migrating influence of a mantle plume.

1. Introduction

Continental rift systems are first-order tectonic features that record the early stages of continental breakup. In magmatically controlled rifts, long-wavelength crustal updoming and the development of dynamic topography prior to the onset of volcanism and normal faulting underscore the role of mantle-driven, thermally controlled processes in the evolution of these regions [e.g., Crough, 1983; Ebinger and Sleep, 1998; Moucha and Forte, 2011]. In the East African Rift System (EARS), such changes in topography and relief have had far-reaching consequences, including impacts on atmospheric circulation patterns that in turn affect rainfall patterns, drainage systems, and surface processes [e.g., Levin et al., 2009; Sepulcre et al., 2006; Ebinger and Scholz, 2012; Wichura et al., 2010, 2015]. In addition, the interplay between extensional tectonics and superposed changes in climate has given rise to gateways and migration corridors for hominids and other mammals, thus fostering speciation [Boebe and Behrensmeyer, 2004; Bailey and King, 2011; Bailey et al., 2011].

Uplift, volcanism, and normal faulting in the EARS are hallmarks of one of the largest magmatic extensional zones on Earth [Burke, 1996]. Composed of the largely amagmatic western and the magmatic eastern branch, the ~5000 km long EARS has generated a series of transiently linked and isolated rift basins [Tercelin and Lezzar, 2002; Ebinger and Scholz, 2012]. The areally extensive mantle anomaly that underlies the EARS [e.g., Simiu and Keller, 1997; Ebinger and Sleep, 1998; Achauer and Masson, 2002] helps support average elevations of ~1000 m [Moucha and Forte, 2011]. As such, understanding the mechanisms of rifting and its spatiotemporal evolution is critical for exploring how geodynamic and surface processes are potentially linked with topographic development, magmatic evolution, and long-term environmental and biotic impacts in rift systems. Changes in each of these aspects over time throughout the EARS have been difficult to quantify, but valuable information has been obtained from the sedimentary and volcanic rift-basin strata [e.g., Frostick and Reid, 1990; Morley et al., 1992;
Unfortunately, many of the rift basins do not allow for a direct inspection of such deposits, because old strata either lie below thick volcano-sedimentary sequences or areally extensive lakes [e.g., Flannery and Rosendahl, 1990; Cohen et al., 1993; Scholz and Finney, 1994; Hautot et al., 2000] or because differential faulting, uplift, and erosion along the rift flanks have been insufficient to expose such deposits. For these reasons, many studies of the spatiotemporal trends in rift-basin formation have relied on the combined analysis of geophysical data, field observations, and isolated drill-core data.

Low-temperature thermochronometry, such as apatite fission track and apatite and zircon (U-Th)/He dating, combined with thermal history modeling, constitute powerful alternative tools to investigate cooling as a proxy for tectono-thermal and climate-driven erosion processes in rift settings [Fitzgerald, 1992; Foster and Gleadow, 1992, 1996; van der Beek et al., 1998]. Thermochronology studies have been carried out throughout the EARS aimed at unraveling the history of basin formation, rift-shoulder exhumation, and different tectono-thermal episodes associated with mantle-plume activity. These studies have addressed the tectonic evolution of the western branch of the EARS in the Rwenzori Mountains [Bauer et al., 2010, 2012] and the Malawi and Rukwa rifts [van der Beek et al., 1998; Roberts et al., 2012], and the evolution of the eastern branch in Ethiopia [Pik et al., 2008], northern Tanzania [Noble et al., 1997; Mbede, 2001], and Kenya [Wagner et al., 1992; Foster and Gleadow, 1992, 1996; Spiegel et al., 2007]. In the greater Turkana region and the western rift-shoulder areas of northern Kenya (Figure 1), these investigations have...
provided helpful constraints on the regional onset of Mesozoic and early Cenozoic rifting [Foster and Gleadow, 1996; Spiegel et al., 2007]. However, the spatial extent of the related tectono-thermal events, the formation of older rift basins underlying the late Cenozoic rifts, and the role of their structures in influencing Miocene to Recent rifting are unclear.

The onset of rifting in East Africa has been associated with a southward-directed migration of volcanism, which in turn has been inferred to be linked with the northward motion of the African plate over a mantle plume [Ebinger and Sleep, 1998; Nyblade and Brazier, 2002]. Accordingly, the onset of tectonic activity in the eastern branch of the rift is also thought to have followed this temporal trend from Ethiopia to northern Tanzania [Nyblade and Brazier, 2002]. In contrast, Zeyen et al. [1997] proposed that the onset of extension was less systematic, and largely dictated by the spatially variable mechanical properties of the lithosphere and crust. The notion that rifting and volcanism did not follow a systematic, southward migration has been recently emphasized again by numerical modeling [Koptev et al., 2015] and by a synopsis of the age of volcanism in the different sectors of the EARS, from which a synchronous onset of volcanic activity in East Africa has been proposed [Michon, 2015]. Furthermore, studies of noble gases in East African lavas and xenoliths suggest the existence of separate plume heads in Ethiopia and Kenya [Halldórsson et al., 2014]. As the Kenya Rift is located between the Tanzania and Turkana-southern Ethiopia extensional provinces, it constitutes an important link between different extensional sectors in the volcanically active eastern branch of the EARS and is thus in a crucial location to further test these ideas.

Here we report 15 apatite (U-Th)/He (AHe), 13 apatite fission track (AFT), and 5 zircon (U-Th)/He (ZHe) ages from basement rocks that were collected along three elevation transects at the Elgeyo escarpment and the Samburu Hills (Nyiru Range) on the eastern and western rift shoulders and steep rift flanks of basins in northern Kenya (Figures 1 and 2). Following thermal modeling of our data using the HeFTy [Ketcham, 2005;
Ketcham et al., 2009] and QTQt software programs [Gallagher et al., 2009; Gallagher, 2012], we use the cooling histories to help elucidate the Cenozoic tectono-thermal evolution of the Kenya Rift shoulders and its implications for the overall structural evolution of the EARS.

2. Geological Setting and Tectonic History

In Kenya, the EARS is characterized by asymmetric rift basins, central volcanoes, and high rift shoulders (up to ~2000 m above sea level (asl) in western Kenya) bounded by escarpments where late Proterozoic crystalline basement rocks are exposed [Maboko et al., 1985; Smith, 1994; Smith and Mosley, 1993; Shackleton, 1993]. Our study area comprises the eastern shoulders of the northern Kenya Rift and the transition between the northern and central Kenya rifts along the western rift margin (Figure 1).

Geophysical data, stratigraphic information, and dated volcanic lava flows have furnished information on the early rifting history in northern Kenya, which began during the late Mesozoic [Foster and Gleadow, 1992, 1996; Morley et al., 1992; Wagner et al., 1992; Ebinger et al., 2000; Tiercelin et al., 2004, 2012]. These early episodes of extension affected areas between the Indian Ocean and the Lake Turkana region, leading to the accumulation of thick lacustrine, fluviatile, and eolian strata in the NW-SE oriented Cretaceous Anza Rift (Figure 1). In its central part, the buried Anza Rift was active during the late Cretaceous, whereas it continued to be active into the late Tertiary in the northern Lamu Embayment (Figure 1), in the vicinity of the present-day coast of the Indian Ocean [Bosworth and Morley, 1994]. Farther west, the Anza Rift is inferred to transition into the eastern sectors of the Cretaceous Central African Rift Zone [Schull, 1988].

Subsequent Cenozoic extension in northern Kenya is recorded in the Turkana-Lokichar rift zone (Figure 1b) but was preceded by volcanism at about 35 Ma, followed by Oligocene normal faulting and the formation of several half-graben basins [Morley et al., 1992]. These basins span the region between the Ugandan border and Lake Turkana (Figure 1), and they host sedimentary fills 5 to 8 km thick that include intercalated volcanic rocks [Morley et al., 1992; Tiercelin et al., 2012]. However, apatite (U-Th)/He and fission track data from the western rift flank (Cherangani Hills of northern Kenya) and from the present-day eastern rift flank southeast of Lake Turkana (Figures 1 and 2) only reveal clear episodes of cooling during the late Cretaceous to Paleocene and during the late Neogene [Foster and Gleadow, 1996; Spiegel et al., 2007].

Today, the northern Kenya Rift encompasses the wide, early Cenozoic Turkana-Lokichar rift zone, but the Miocene to Recent manifestations of rifting are primarily located in the eastern sector of this extensional province, between approximately 1°N and 4.5°N latitude. This region is characterized by the active, down-to-the-east half-graben basin of the NNE-oriented Suguta Valley [Bosworth and Maurin, 1993; Saneyoshi et al., 2006; Melnick et al., 2012] and the northern sector of the NNE-oriented Elgeyo Escarpment farther west (Figure 1b). The rift floor in the Suguta Valley is at ~300 m asl, whereas the antithetically faulted monocline on the eastern rift margin has an average elevation of ~1400 m asl [Bosworth and Maurin, 1993]. There, the rift shoulder exposes gneissic Precambrian basement in the Samburu Hills, our first study site, which locally reach up to 1900 m asl (Figures 1 and 2). To the north of the Suguta Valley is the active Kino Sogo rift zone; this extensional sector transitions northward into the Chew Bahir Rift of southern Ethiopia, which is an integral part of the Main Ethiopian Rift (Figure 1).

The Elgeyo Escarpment, our second study site, is located along the western margin of the northern Kenya Rift and the northernmost sector of the NNW oriented central Kenya Rift, where the escarpment transitions southward into the Mau Escarpment (Figure 1b). The Elgeyo Escarpment is one of the most prominent fault-line escarpments of the Kenya Rift. In its southern sector, the escarpment is ~900 m higher than the adjacent Kerio sedimentary basin (Figure 2). The Elgeyo Escarpment is related to a down-to-the-east normal fault, which exposes steeply eastward dipping Proterozoic gneisses of the pan-African orogeny [Chapman et al., 1978; Maboko et al., 1985; Hetzel and Strecker, 1994]. The Cenozoic faults along the Elgeyo Escarpment follow the gneissic foliation and change in strike, where the foliation is cut by NW striking Proterozoic dextral shear zones [Strecker et al., 1990; Shackleton, 1993]. These reactivated shear zones influence the geometry and areal distribution of strata of the Kerio sedimentary basin [Mugisha et al., 1997]. The basement shear zones are overlain by arkosic sandstones, conglomerates, and lacustrine shales of unknown age, and up to 150 m thick, 14 Myr old phonolites [Lippard, 1973; Ego, 1994; Supplementary Figure 1] that also cap the western rift shoulder. Faulting along the Elgeyo Escarpment in this sector postdates the emplacement of the middle Miocene phonolite flows and generated the westward tilted
Kamasia fault block to the east (Figure 2), which in turn is delimited by a major down-to-the-east normal fault to the east [e.g., Chapman et al., 1978]. The Kamasia Range thus forms the barrier between the Kerio Basin and the Baringo Basin farther east, which now hosts the active volcano-tectonic axis of this part of the Kenya Rift.

3. Methodology

We constrain the Cenozoic cooling of the Precambrian Mozambique Belt gneisses exposed on the western and eastern rift shoulders of the northern Kenya Rift and the northernmost sector of the central Kenya Rift using apatite fission track thermochronology (AFT; apatite partial annealing zone, PAZ, ~60 to ~120°C), apatite (U-Th-5m)/He thermochronology (AHe, partial retention zone, APRZ ~40°C–85°C), and zircon (U-Th)/He thermochronology (ZHe; ZPRZ ~120 to ~200°C) (assuming that zircons remained in this temperature range for about 90 Myr [e.g., Reiners and Brandon, 2006; Wolf et al., 1998; Ketcham et al., 1999; Reiners, 2005]). We collected seven samples (KN83 through KN89) along a W-E oriented profile between 1300 and 1900 m asl in the Samburu Hills (SH), on the eastern rift flank (Figure 3, Tables 1 and 2, and Table S1 in the supporting information). On the western flank, we collected one steep profile across the northern Elgeyo Escarpment (NEE) and one across the southern Elgeyo Escarpment (SEE) (Figure 2). The NEE profile comprises six samples (KN97 through KN102) collected between 1280 and 1850 m asl, and the SEE profile comprises five samples (KN53, KN55, KN91, KN92, and KN94) collected between 1350 and 1780 m asl. The Elgeyo escarpment profiles are separated by ~50 km and were collected perpendicular to the main fault-line escarpment (Figure 2). In addition, sample KN50 was collected between the two profiles and analyzed for AFT, although it was not used to help constrain the thermal modeling.

Mineral separation and sample preparation for AFT, AHe, and ZHe analyses followed standard procedures [e.g., Dobson, 2006] and are briefly described below. Eighteen samples yielded enough apatite for AFT and AHe analysis, and six samples were analyzed with the ZHe method.

3.1. Apatite Fission Track Analysis

AFT samples were analyzed with a Leica DMRM microscope at the University of Potsdam. Approximately 20 good-quality grains per sample were randomly selected and dated using the external detector method and the zeta calibration technique [Hurford and Green, 1983] (see Table 1). The pooled age is reported (±1σ) when samples pass the chi-square test ($P(\chi^2) \geq 5\%$) [Galbraith, 1981]; for KN50, the central age is reported because the sample failed this test. We also report confined track-length distributions of seven AFT samples (see Table 1). $^{252}$Cf irradiations were performed (see supporting information) to obtain a larger number of horizontally confined tracks for track-length measurements [e.g., Donelick et al., 2005]. The angle between the confined tracks and the crystallographic c axes was routinely measured. The size of the etch-pit diameter parallel to the c axis (Dpar) was also determined, as it is a kinetic parameter used in thermal history modeling [Donelick et al., 1999; Ketcham et al., 1999] At least four Dpar values were measured per crystal; the data were corrected following Sobel and Seward [2010].
Table 1. Apatite Fission Track (AFT) Data From the Kenya Rift

<table>
<thead>
<tr>
<th>Sample</th>
<th>Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>Elev (m asl)</th>
<th>Lithology</th>
<th>Xl</th>
<th>Rho-S ×10⁶</th>
<th>NS</th>
<th>Rho-I ×10⁶</th>
<th>NI</th>
<th>P(χ²) (%)</th>
<th>RhoD ×10⁶</th>
<th>ND</th>
<th>Ageb (Ma) ±σ (Ma)</th>
<th>Dpar mean</th>
<th>Dpar SD</th>
<th>n</th>
<th>Mean length</th>
<th>Mean lengths</th>
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aSample preparation and analysis similar to that used by Sobel and Strecker (2003). All apatites were etched in 5.5 mol nitric acid for 20 s at 21°C. Samples were analyzed with a Leica DMR microscope with drawing tube located above a digitalizing tablet and a kinetek computer-controlled stage driven by the Fstage program [Dumitru, 1993]. Analysis was performed with reflected and transmitted light at 1250X magnification. Samples were irradiated at Oregon State University. Following irradiation, the mica external detectors were etched with 21°C, 40% hydrofluoric acid for 45 min. The pooled age is reported for samples with P(χ²) greater than (less than) 5% as they pass (fail) the χ² test. Age errors are presented as one sigma, calculated using the zeta calibration method [Hurford and Green, 1983], AB:369.5 ± 7.9, unpublished, 2012.

bPooled age is reported for most of the samples, except for sample KN50, central age is reported.

c252Cf irradiations on additional grain mount of the sample.
3.2. Apatite (U-Th-Sm)/He Analysis

The AHe method is based on the accumulation of alpha particles produced by U, Th, and Sm series decay [Ehlers and Farley, 2003]. Ages are corrected for alpha ejection near the margins of the crystal assuming a homogenous distribution of U, Th, and Sm [Farley et al., 1996; Ehlers and Farley, 2003]. Samples were analyzed at the Scottish Universities Environmental Research Center (SUERC) following the procedure described by Foeken et al. [2006]. We analyzed between 1 and 3 apatite crystals of similar radius per sample (Table S1). Corrections for He recoil loss were made following the procedures of Farley and Stockli [2002].

The correction factor (Ft) was calculated based on the retention and stopping distance of the alpha particle in the crystal and the size of each grain analyzed. The reproducibility of the 3He spike was determined from three daily measurements against an accurately known 4He standard. Standard 4He abundance measurements have within-run precision of better than 0.1% (1σ) and between-run precision of 0.2% (1σ) [Persano et al., 2002].

U, Th, and Sm analyses were performed on a VG plasma Quad PQ2 + inductively coupled plasma–mass spectrometry. After 4He analysis, each crystal was dissolved and spiked with ~0.45 ng 230Th and 0.20 ng 235U with approximately 2 mL of HNO3. Total analytical uncertainty on all ages was approximately 1–3% (1σ), which is dominated by the uncertainty in the U and Th spike concentrations [e.g., Dobson, 2006; Foeken et al., 2006], He determinations, blank corrections, and uncertainties on grain-size measurements for α-correction.

3.3. Zircon (U-Th)/He Analysis

The (U-Th)/He analysis of zircons is also based on the accumulation and diffusion of alpha particles produced by the decay of U and Th. After mineral-separation at the University of Potsdam, we used the facilities at SUERC following analytical procedures described by Foeken et al. [2006] and Dobson [2006]. Seven samples with one to three single-crystal aliquots each were analyzed. With respect to the Ft correction, we followed procedures described in Farley and Stockli [2002], and we again assumed a uniform distribution of U and Th. Total analytical uncertainty is dominated by the uncertainty in the U and Th spike concentrations [Dobson, 2006; Foeken et al., 2006], He determinations, and blank corrections.

3.4. Age-Elevation Relationships and Thermal History Modeling

The relationship between thermochronologic ages and elevations of different samples along an elevation profile are commonly used to determine rates and amounts of exhumation [e.g., Dobson et al., 2009; Foeken et al., 2006]. However, the thermal structure of the uppermost crust could be influenced by topography, erosion, faulting, and advection [e.g., Ehlers and Farley, 2003; Gallagher et al., 2005], which complicate direct interpretations of the slope of age-elevation plots [e.g., Valla et al., 2010; van der Beek et al., 2010].

Table 2. (U-Th)/He Ages of Zircon (ZHe) in the Kenya Rift

<table>
<thead>
<tr>
<th>Sample</th>
<th>Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>Elev (m asl)</th>
<th>U (ng)</th>
<th>Th (ng)</th>
<th>4He (cc)</th>
<th>Ft</th>
<th>Corrected Age (Ma)</th>
<th>Error (1σ) (Ma)</th>
<th>Mineral</th>
</tr>
</thead>
<tbody>
<tr>
<td>KN91</td>
<td>35.599</td>
<td>0.305</td>
<td>1776</td>
<td>6.7</td>
<td>91.9</td>
<td>6.87E-07</td>
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</tr>
<tr>
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<td>0.305</td>
<td>1776</td>
<td>7.0</td>
<td>88.5</td>
<td>3.48E-07</td>
<td>0.85</td>
<td>37.3</td>
<td>3.8</td>
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</tr>
<tr>
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<td>0.302</td>
<td>1583</td>
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<td>34.2</td>
<td>1.07E-07</td>
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</tr>
<tr>
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<td>35.611</td>
<td>0.302</td>
<td>1583</td>
<td>5.1</td>
<td>18.7</td>
<td>9.63E-08</td>
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<td>54.0</td>
<td>5.5</td>
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</tr>
<tr>
<td>KN94</td>
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<td>0.312</td>
<td>1389</td>
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<td>41.7</td>
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<td>4.1</td>
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</tr>
<tr>
<td>KN94</td>
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<td>0.312</td>
<td>1389</td>
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<td>23.9</td>
<td>2.29E-07</td>
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<td>Zircon</td>
</tr>
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<td>1.74E-07</td>
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<tr>
<td>KN97</td>
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<td>3.6</td>
<td>7.5</td>
<td>5.03E-08</td>
<td>0.83</td>
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</tr>
<tr>
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<td>28.6</td>
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</tr>
<tr>
<td>KN102</td>
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<td>0.660</td>
<td>1852</td>
<td>0.3</td>
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<td>5.79E-09</td>
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<tr>
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<td>0.660</td>
<td>1852</td>
<td>8.6</td>
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<td>1.64E-07</td>
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</tr>
<tr>
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<td>0.660</td>
<td>1852</td>
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<td>32.7</td>
<td>1.94E-07</td>
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<td>1601</td>
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<tr>
<td>KN85</td>
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<td>2.033</td>
<td>1601</td>
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<td>1.0</td>
<td>6.54E-09</td>
<td>0.81</td>
<td>50.5</td>
<td>5.2</td>
<td>Zircon</td>
</tr>
</tbody>
</table>
To determine the Cenozoic cooling history recorded by samples within our age-elevation profiles, we used two different thermal modeling approaches. We performed forward and inverse Monte Carlo modeling of time-temperature paths combining AFT and ZHe data using the HeFTy 1.8.2 software [Ketcham, 2005] for individual samples, and we used the QTQt software [Gallagher et al., 2009; Gallagher, 2012] to derive integrated cooling constraints from multiple samples within an age-elevation transect. AHe data were not used for modeling due to the high dispersion of single-grain ages, as discussed in section 4.

The HeFTy program is effective for assessing the thermal history of a single sample. The output reflects the goodness of fit of the models, which allows for testing different cooling histories. However, the program has not yet been adapted for modeling multiple samples within an elevation profile. In contrast, the QTQt program is well suited for the latter task. Therefore, we use the HeFTy results to determine approximate constraints that can be applied during QTQt modeling of the complete profile.

To resolve the Cenozoic thermal history of the rocks along each profile, we tested two contrasting thermal history scenarios: monotonic cooling versus Neogene reheating and cooling. We used the HeFTy and QTQt programs to predict time-temperature paths for these two scenarios that are consistent with our AFT age and track-length distribution data and our ZHe ages. Additionally, we used geological constraints, such as the initiation of volcanism north of the study area [Morley et al., 1992; Ebinger and Sleep, 1998], to provide some brackets on possible cooling paths.

4. Results

4.1. Apatite Fission Track Dating Results

The apatite fission track cooling ages are shown in Table 1. Over the 400 m elevation range of the Samburu profile (SH), samples KN89, KN85, and KN83 yielded ages ranging from 39.4 ± 3.0 Ma to 50.3 ± 8.0 Ma. Although the elevation difference is small, this section provides the best relief and outcrop conditions in the area. Mean track lengths from samples KN83 and KN85 are 12.2 and 11.9 μm, respectively; sample KN89 did not yield a meaningful number of track lengths (n = 2; Table 1). Samples from the northern Elgeyo transect (NEE, four samples) yielded ages ranging from 26.2 ± 3.0 Ma to 38.8 ± 3.0 Ma, with mean track lengths of 11.2 to 11.8 μm that were measured from three out of the four samples (KN99, KN100, and KN102).

The southern Elgeyo transect comprises six samples with an age range between 4.6 ± 1.0 Ma and 9.2 ± 2.0 Ma. No confined track lengths were found in those samples. In comparison to the AFT ages of the NEE and SH profiles, the young AFT ages of this transect together with the lack of a clear increase in age with elevation may be related to reheating as a result of nearby fluorite mineralization associated with the emplacement of Miocene lava flows. The fluorite mineralization shows ENE to east trending growth fibers in the plane of the steeply east dipping foliation [Aljabri, 1992; Hetzel and Strecker, 1994; Ogola et al., 1994]. As fluorite mineralization involves temperatures between 120 and 260°C [e.g., Richardson and Holland, 1979], AFT ages were likely at least partially reset, and therefore, we do not use those data to model the tectono-thermal history in this part of the study area.

4.2. Apatite (U-Th-Sm)/He Results

We report 54 single-grain AHe ages from 15 samples (Table S1). In most samples, several single-grain AHe ages are older than the AFT age from the same sample, and only a few aliquots yield younger AHe ages (Table S1 and Figure 3). With respect to calibration and operation of the analytical machines, Ft corrections on potential nonideal geometries, and grain-size measurement uncertainties, none can explain independently the dispersion of the data. Below we consider other possibilities.

Because prominent zonation in U and Th can also affect AHe ages, we examined the distribution of induced tracks in the micas of AFT samples that were also analyzed with the AHe method to assess the variability of the U distribution within single grains. Many of the samples contained numerous grains with pronounced U zoning, often with U-rich rims, although U-rich cores relative to the rims are required to explain ages that appear too old. Two samples (KN53 and KN55) were also characterized by very high crystal-dislocation densities. During AFT analysis, such crystals were not analyzed. However, these characteristics can help explain the high age dispersion within the AHe data.
Particularly, in the case of slowly cooled apatites, differing amounts of radiation damage reflected by variable effective uranium ("eU") have been invoked to explain widely dispersed AHe ages [e.g., Flowers, 2009] or AHe ages that are older than AFT ages [e.g., Green et al., 2006; Hansen and Reiners, 2006; Gautheron et al., 2009]. Higher eU in apatite leads to more radiation damage, higher He retentivity, and hence a higher effective closure temperature and a greater accumulation of He for a given cooling history [Flowers et al., 2007]. Alternatively, variations in the effective diffusion domain size could influence the closure temperature of the crystal [Gautheron et al., 2009]. To explore these possibilities, we plotted AHe age versus eU and versus grain size. However, the ages do not show a clear correlation with either parameter (Figure S2 in the supporting information).

Radiation damage and variable eU problems are commonly discussed in cases with AHe ages that are much older (>100 Ma) [e.g., Flowers, 2009] than those in this study. However, prolonged residence time at a low temperature (i.e., ~80–100°C) prior to late burial heating has been shown to exert a strong influence on 4He retentivity and hence on the effective closure temperature [Fox and Shuster, 2014]. In our study, the basement rocks are ~500 Myr old and are related to the Pan-African orogeny [Maboko et al., 1985; Smith, 1994]. Between these early tectonic processes and the initiation of Cenozoic rifting [e.g., Foster and Gleadow, 1996; Spiegel et al., 2007], the samples remained at a relatively constant position below the surface (~3 km). This thermal history is quite analogous to the Grand Canyon [Fox and Shuster, 2014], suggesting that AHe results from there are applicable to our study. The extremely long residence time of samples at depths of a few kilometer led to a variable and large amount of accumulated radiation damage, which caused higher 4He retentivity and hence anomalously old AHe ages. Such a scenario provides a better explanation for the old ages in our study than variations in grain size, eU, zonation, or analytical problems.

4.3. Zircon (U-Th)/He Results

We analyzed a subset of samples from each of our three elevation profiles for ZHe. From the Samburu Hills profile, we analyzed samples KN58 and KN83. Two aliquots from sample KN85 provide a ZHe mean age of 50.6 ± 5.2 Ma. From sample KN83, the anomalously young age (1.5 Ma) is paired with very low He and high U and Th concentrations. Induced tracks in the external detector from occasional zircons included in the AFT grain mount of the same sample show that many of these zircons are strongly zoned and have U-rich rims. Therefore, data from this sample was not plotted in the age-elevation plot and not used for thermal modeling (Figure 3a). From the Northern Elgeyo Escarpment profile, samples KN97 and KN102 yielded ZHe ages of 59.7 ± 6.1 Ma (single aliquot) and mean age of 55.9 ± 5.6 Ma (three aliquots), respectively. A fourth aliquot of sample KN102 was discarded due to the low reproducibility of the age (Table 2). From the Southern Elgeyo Escarpment profile, samples KN91, KN92, and KN94 (Figure 3c) have mean ages of 53.9 ± 5.5 Ma (two aliquots), 42.2 ± 4.3 Ma (two aliquots), and 56.7 ± 5.8 Ma (three aliquots), respectively.

4.4. Thermal History Modeling of Elevation Transects

We performed thermal modeling on the NEE and SH transects. Thermal models that incorporated the AHe data were incompatible with the AFT data. Because of the large AHe age dispersion, low reproducibility of ages, and strongly likelihood of variable 4He retentivity (as discussed previously), we decided to base the thermal modeling and our interpretations on only the AFT and ZHe data. (see Table S1 and Figure S2).

Our general approach was to first model an individual sample from an elevation profile (sample KN85 from profile SH, elevation 1600 m, and sample KN102 from profile NEE, elevation 1852 m; Figure 3) using the HeFTy program. The HeFTy modeling results were used as the basis for defining a common thermal history of the samples [e.g., Prenzel et al., 2013] along the vertical profile. Next, we used QTQt to model the three samples from the SH profile simultaneously and the four samples from the NEE profile simultaneously. Monotonic cooling and reheating paths were tested with both programs. Constraint boxes were first defined to observe preferential cooling paths of the analyzed samples. In subsequent runs, the constraint boxes were shifted to examine the reliability of inflection points representing changes in modeled cooling rates and to allow for reheating.

Input parameters for modeling included AFT data (age, c axis nonprojected track lengths, and Dpar values), the ZHe data, and present-day surface temperatures. Acceptable-fit paths (goodness of fit >0.05) from the HeFTy models of individual samples were considered for interpretation.
Figure 4. Thermal modeling results for the Samburu Hills (SH) samples. (a and c) Time-temperature model for sample KN85 from the Samburu Hills (SH) using HeFTy software [Ketcham, 2005] with acceptable (yellow) and good (purple) time-temperature pathway envelopes and best fit shown with black line. The modeling scheme in Figure 4a only permits monotonic cooling while Figure 4c permits reheating between 40 and 20 Ma. (b and d) Time-temperature histories derived from QTQt modeling, with time-temperature constraints in B similar to those in A and with constraints in D similar to those in Figure 4c. (e) Time-temperature history paths derived from QTQt with no constraint boxes; red box defines the limits of the modeling, 0°C–220°C and 0–80 Ma. Blue and red lines correspond to the uppermost and lowermost sample; cyan and magenta lines correspond to 95% confidence intervals for uppermost and lowermost samples, respectively.
4.4.1. Thermal Modeling of Data From the Samburu Hills

For our thermal modeling of the Samburu Hills (SH) data, we first ran models with monotonic cooling paths using both the HeFTy and QTQt programs (Figures 4a and 4b). We defined a first constraint box with a temperature range from 135 to 15°C for HeFTy (Figure 4a) and for QTQt (Figure 4b) over a time range from 15 Ma to the present. Second constraint boxes were defined between ~130 and 30°C at ~30 to 20 Ma. Third constraint boxes were defined between 230 and 160°C at ~75 to 45 Ma to be compatible with the AFT and ZHe cooling ages. HeFTy models always end with a temperature constraint at the present day ranging from 25 to 15°C. The QTQt models end at present-day temperatures from 0° to 50°C.

A second set of models was run to test if the thermochronologic ages are compatible with either a reheating event during the Paleogene, the onset of volcanism in the late Eocene-early Oligocene, or early exhumation. The track-length distribution of sample KN85 (Figure 4a) suggests that reheating could be possible. Geologically, reheating could be associated with volcanism prior to or during the early stages of rifting. To allow for this possibility, we reduced the temperature limits of the second boxes to 100°C and 10°C, thus permitting paths to reach lower temperatures at an earlier phase in the cooling history and extending the duration of this interval to between 40 and 20 Ma (Figure 4c). Other constraints were similar to the first run.

Well-fitting HeFTy models for monotonic cooling of sample KN85 (Figure 4a) suggest that a first episode of relatively rapid cooling (12°C/Myr) occurred from ~60 to 50 Ma. Subsequently, very slow cooling (~1.5°C/Myr) took place from ~45 Ma to ~15 Ma, followed by cooling to the present. As the model results are not well constrained over this final timeframe, we refrain from interpreting this cooling trend further. Because the model results rely on the calculating paths through the lowest temperature part of the PAZ and below the PAZ, where not much control on the thermal history exists, it is not known if cooling may have accelerated or not.

Using QTQt, we obtained results for the entire profile (Figure 4b) that are broadly similar to the HeFTy results obtained for sample KN85 (Figure 4a). Rapid cooling from about 65 to 50 Ma (~10°C/Myr) is followed by slow reheating (~1°C/Myr) from ~40 to ~20 Ma. The model reveals a slight acceleration of cooling (~3°C/Myr) at ~15 Ma as part of a final continuous cooling episode, which has lasted to the present. Acceptable fits also permit an isothermal path without reheating from as early as ~50 Ma to as late as ~20 Ma. Despite some potential reheating of the uppermost samples within the transect between 40 and 20 Ma, the thermal histories for the different samples in the SH appear to have experienced similar t-T paths and are consistent with previous studies in the region (e.g., Foster and Gleadow, 1992, 1996). Acceptable fits for the reheating model using HeFTy (Figure 4c) follow a broader swath than for the monotonic cooling model (Figure 4a). The best fitting path shows that reheating is plausible. This scenario (Figure 4c) shows rapid cooling (~10°C/Myr) from 55 to 40 Ma. Cooling was followed by an ~20 Ma interval at a near-constant temperature. A short-lived early Miocene pulse of reheating to temperatures as high as 100°C could have occurred, although a broad range of other paths that do not include reheating lie within the acceptable-fit envelope. Final cooling occurred from ~15 Ma to the present.

The QTQt reheating model results for the entire profile in the reheating scenario (Figure 4d) show similar changes in the cooling rates. A first rapid cooling from ~65 to 50 Ma, reheating from ~50 to 35 Ma of ~2°C/Myr (particularly for the uppermost samples in the profile), then slow cooling from 35 to ~15 Ma, and final cooling from ~15 Ma until the present (Figure 4d). These results are quite similar to the previous QTQt modeling scenario (Figure 4b).

From these two sets of models (monotonic cooling and reheating) and the general geologic context of the region, we conclude that reheating is permissible, but not required by the thermochronologic data. The differences between the two models suggest that the magnitude and timing of such a reheating event can only be loosely constrained (Figures 4c and 4d).

4.4.2. Thermal Modeling of Data From the Northern Elgeyo Escarpment

Thermal models were run for the Northern Elgeyo Escarpment (NEE) profile in a similar manner as described above for the Samburu Hills profile. For the HeFTy (Figure 5a) and QTQt (Figure 5b) monotonic cooling models of sample KN102, the first constraint boxes were set to 160 to 10°C at 15 Ma to the present day. Second constraint boxes were set to ~150 to 30°C at 40 to 20 Ma for HeFTy and for QTQt, based on the AFT ages, sedimentary evidence in the hanging wall, and volcanic flows on top of the footwall. Third constraint boxes were set to 220 to 160°C at 75 to 40 Ma, based on an early to middle Tertiary regional extension with minor strain, which is also associated with the latest stages of extension of the Anza Rift in the north [Foster and Gleadow, 1996]. Again, HeFTy models always end with a constraint at the present-day surface temperature, ranging from 25 to 15°C, and QTQt models terminate at present-day temperatures ranging from 0° to 50°C.
Figure 5. Thermal modeling results for the northern Elgeyo Escarpment (NEE) samples. (a and c) Time temperature model for sample KN102 from the Northern Elgeyo Escarpment (NEE) using HeFTy software with acceptable (yellow) and good (purple) time-temperature pathway envelopes and best fit shown with black line. Figure 5a requires monotonic cooling while C permits reheating between 35 and 15 Ma. (b and d) Time-temperature histories derived from QTQt modeling, with time-temperature constraints in Figure 5b similar to those in Figure 5a and with constraints in Figure 5d similar to those in Figure 5c. (e) time-temperature history paths derived from QTQt with no constraint boxes; red box defines the limits of the modeling, 0°C–220°C and 0–80 Ma.
The second reheating model used similar constraints as the monotonic model, but with slight changes in the time-temperature limits of the boxes. The second constraint box shows the most pronounced change, extending the range of temperatures from 120 to 10°C over a time range from 49 to 20 Ma in the HeFTy and QTQt models (Figures 5b and 5d). This additional constraint permits reheating (by allowing for cooling to lower temperatures at an earlier period), which could be due to basaltic volcanism in the region starting in the northwest sector of Lake Turkana at ~37 Ma [i.e., Zanettin et al., 1983; McDougall and Brown, 2009] and farther south during the early Miocene (Samburu Basalts, ~20–11 Ma) [Hackman et al., 1990] (Figure S1).

HeFTy results for the monotonic cooling model for sample KN102 (Figure 5a) display similar cooling paths as modeled for sample KN85 in the Samburu Hills (Figure 4a). Rapid cooling (>20°C/Myr) from ~55 to 50 Ma is followed by very slow cooling (1.5°C/Myr) from 45 to 10 Ma, and finally moderate cooling of >2.0°C/Myr until the present (Figure 5a).

QTQt results from the entire profile at the Elgeyo Escarpment are broadly consistent with the HeFTy results, also showing rapid (>20°C/Myr) cooling from ~60 to 55 Ma. But rather than very slow cooling from ~50 to 15 Ma, the results show reheating (Figure 5b). This reheating is followed by final cooling (~3°C/Myr) from 15 Ma to the present.

The HeFTy thermal model involving reheating (Figure 5c) shows many similarities to the HeFTy model of monotonic cooling (Figure 5a). The best fit of the reheating model exhibits relatively fast cooling rates (~15°C/Myr) from ~60 to 50 Ma and moderately fast cooling from ~50 to 30 Ma. Slight reheating follows from ~30 to 10 Ma; this reheating is followed by rapid cooling to the surface until the present. As observed for sample KN85 from the Samburu Hills (Figures 4a and 4c), the broad envelope of the acceptable cooling paths shows that this model is not as tightly constrained as the monotonic cooling scenario (Figure 5a). The second QTQt model has constraint boxes that were set between 160 to 10°C at 15 to 0 Ma, 120 to 10°C at 40 to 20 Ma, and 220 to 160°C at 75 to 45 Ma. The results show reheating constrained to a shorter time window between ~35 and 10 Ma, even if the result looks forced by the constraint (Figure 5d). For the remaining time windows, the QTQt modeling results agree with those from HeFTy.

4.4.3. Unconstrained Thermal Modeling of Elevation Profiles
In addition, we ran QTQt models for both regions without using any constraint boxes (Figures 4e and 5e) to explore best fitting thermal histories when the program is set to freely search for acceptable solutions. The results do not differ substantially from the results of the constrained model runs described above. These results overall show that between periods of rapid cooling, the Samburu Hills profile may have experienced limited reheating, while the Northern Elgeyo Escarpment is likely to have experienced reheating. Ultimately, the modeling results are consistent, independent of the choice of the constraint boxes. Therefore, the input data, i.e., the ages and track lengths, guide the basic shape of the paths, rather than the user-defined constraint boxes. Other modeling results, including scenarios with short pulses of reheating, are provided in the supplementary material (Figure S3).

Overall, results of the modeling with single (HeFTy) and multiple (QTQt) samples suggest a similar thermal history for the eastern and the western flanks of the northern and central Kenya Rift sectors, with clear periods of rapid cooling between ~65 and 50 Ma and from ~15 Ma to today, with either stable temperatures or reheating in between (Figures 4 and 5).

5. Discussion
Our new thermochronological data and thermal modeling results from the Samburu Hills and the Elgeyo Escarpment of Kenya define three Cenozoic stages of thermal evolution: (1) rapid cooling between ~65 and ~50 Ma, (2) subsequent slow cooling or slight reheating during the Oligo-Miocene, and (3) renewed rapid cooling after 15 Ma. Below, we address these different cooling stages with respect to rift evolution in Kenya and within the broader context of regional extensional processes in East Africa.

5.1. Paleocene to Eocene (65–50 Ma) Rapid Cooling
Our HeFTy and QTQt thermal modeling of samples from the Samburu Hills elevation profile on the eastern side of the rift valley and from the Elgeyo Escarpment on the western side yielded similar results, documenting rapid, early Cenozoic regional cooling (>50°C) between ~65 and 50 Ma (Figures 4 and 5). These results are in good agreement with previous modeling of apatite fission track and length parameters from samples collected in the Cherangani Hills and the basement rocks SE of Lake Turkana, which revealed 60 to 70°C of...
cooling between approximately 60 and 50 Ma [Foster and Gleadow, 1996; Spiegel et al., 2007]. This cooling appears to have been coeval with renewed extension and tectonic subsidence in the Anza Rift [i.e., Morley et al., 1999; Bosworth and Morley, 1994; Morley, 2002], which is inferred to have been associated with flexural upwarping of rift flanks in that region [Foster and Gleadow, 1996].

5.2. Eocene Through Middle Miocene Monotonic Slow Cooling or Reheating

Our results show that very slow cooling or reheating occurred between ~45 and 15 Ma in the Samburu Hills and at the Northern Elgeyo Escarpment, with temperatures ranging between ~60° and 90°C from the Eocene through the middle Miocene. Similar results were reported by Foster and Gleadow [1996] and Spiegel et al. [2007], with very slow cooling along both the western (Cherangani Hills) and eastern (Ndoto Mountains) rift shoulders from the Eocene through the middle Miocene.

The models show that minor reheating is likely to have occurred from Eocene to middle Miocene time (~45 to 15 Ma) for the Northern Elgeyo Escarpment samples, and is permitted, but not required for the Samburu Hills samples over a similar time interval. The Miocene phonolites are too young to explain this reheating. However, for the Northern Elgeyo Escarpment, reheating is compatible with geological interpretations of the early Cenozoic rifting phase in the Kerio Basin (Figure 1a) based on seismic reflection data [Mugisha et al., 1997; Hautot et al., 2000]. The seismic reflection data document an early (as of yet undated) stage of tectonically controlled basin subsidence beneath the Neogene Kerio Basin followed by regional thermal basin subsidence and sedimentation within a sag basin [Morley et al., 1992; Mugisha et al., 1997; Morley and Ngenoh, 1999; Hautot et al., 2000]. Thermo-chronological and geophysical evidence for protracted subsidence in this sag basin is corroborated by regional pinch-outs of fluvial and organic-rich lacustrine sediments exposed along the Elgeyo Escarpment [Morley et al., 1992; Mugisha et al., 1997; Ego, 1994; Renaut et al., 1999] and the regional extent and thickness of the overlying <14.5 Ma phonolites that cover the present-day eastern and western rift shoulders [Lippard, 1973]. Alternatively, reheating could have been associated with the thermal impact of a mantle plume beneath the Tanzanian Craton since the Eocene-Oligocene [Ebinger and Sleep, 1998; George et al., 1998; Pik et al., 2008].

In this context, it is interesting that our thermochronologic data obtained from the Samburu Hills on the eastern rift shoulder exhibit a similar cooling history as samples from the Cherangani Hills on the western rift shoulder [Foster and Gleadow, 1996; Spiegel et al., 2007] (Figure 6), with reheating permitted, but not required. Due to their locations on the rift shoulders, neither area was affected by reheating related to subsidence and sedimentation within a sag basin. Thermal impacts of a mantle plume beneath the Tanzanian Craton since the Eocene-Oligocene [Ebinger and Sleep, 1998; George et al., 1998; Pik et al., 2008] remain a possibility, but clearly did not strongly influence the thermal evolution of the area as recorded in our samples.

5.3. Middle Miocene–Recent Renewed Cooling

Our thermal modeling shows a renewed, rapid cooling event starting at ~15 Ma in the Samburu Hills and the northern Elgeyo Escarpment, with rocks cooling from temperatures of ~60°C in the Samburu Hills and from ~90°C at the northern Elgeyo Escarpment. This accelerated cooling is compatible with the tectono-magmatic and sedimentary evolution of the northern and central Kenya rifts. Extensional faulting and the generation of transfer faults guided by foliation and inherited brittle shear zones in the Proterozoic basement gneisses affected the <14.5 Myr old phonolites as well as undated lacustrine shales, arkosic sandstones, and conglomerates [Hetzel and Strecker, 1994; Ogola et al., 1994; Ego, 1994], now exposed on the rift shoulders at the Elgeyo escarpment (Figure 6). These processes were responsible for the formation of a second rift basin superposed on the Paleogene basin (Figure 6). This younger basin has the geometry of a down-to-the-east halfgraben, which hosts the present-day Kerio Basin. The Neogene phase of extension generated an additional 2 km of sedimentary and volcanic fill in addition to the 4 km of basin deposits that had been deposited during Paleogene rifting, with the corresponding bounding fault being located farther west [Mugisha et al., 1997; Morley and Ngenoh, 1999; Hautot et al., 2000], (Figure 6).

5.4. Regional Implications for Rifting in East Africa

Our thermochronological results from the Elgeyo and Samburu sites in the northern and central Kenya Rift sectors agree with the well-documented, regionally widespread Paleogene episode of extension in the Turkana region [Morley et al., 1992; Foster and Gleadow, 1996; Morley et al., 1999; Morley, 2002; Tiercelin et al., 2012].
East of the present-day Elgeyo Escarpment, Paleogene normal faulting and coeval sedimentation along the proto-Kerio Basin had been previously inferred based on a pronounced negative Bouguer gravity anomaly, reflecting a thick sedimentary fill, thought to be incompatible with the amount of Neogene extension and tectonic basin subsidence (Morley et al., 1992; Mugisha et al., 1997). As neither currently available geophysical, geological, nor thermo-chronological data suggest similar coeval Paleogene extension processes in the central and southern sectors of the Kenya Rift, it appears that early Cenozoic rifting in Kenya was focused in the greater Turkana region (i.e., Morley et al., 1992; Foster and Gleadow, 1992; Foster and Gleadow, 1996; Morley and Ngenoh, 1999; Spiegel et al., 2007), the Anza Rift, where Mesozoic extensional faults were reactivated (i.e., Morley et al., 1992; Bosworth and Morley, 1994), and regions as far south as the transition between the central and northern Kenya rifts (this study).

Figure 6. Cartoon of the structural and thermal evolution of the Kenya Rift along the Elgeyo Escarpment of the Kerio Valley in the transition between the central and northern Kenya rifts. Episodes of cooling occurred from ~65 to 50 Ma and from 15 to 0 Ma. Tectonic quiescence or minor reheating occurred between ~45 and 15 Ma. From bottom to top, circles denote the approximate position of the analyzed samples from ~50 Ma to present-day (surface). Blue band represents the inferred position of the partial retention zone for ZHe (PRZ); red band represents the apatite fission track partial annealing zone (PAZ).
Thermo-chronologic data from other sectors of the EARS reveal spatially disparate and diachronous cooling histories (Figure 7). Approximately 300 km to the southeast of the Elgeyo region, the Pangani Rift of northern Tanzania is the closest manifestation of Paleocene-Eocene extensional processes in the southern continuation of the eastern branch of the EARS. Apatite fission track data indicate that a phase of rapid cooling began during the late Cretaceous and continued throughout the Paleogene [Noble et al., 1997; Mbede, 2001], although no thermal modeling was performed to better constrain the timing. In the western branch of the EARS, the Rwenzori Mountains and the Albertine Rift of Uganda have a distinctly different late Cretaceous to Cenozoic thermal history, with very slow cooling of <0.5 °C/Myr from the late Cretaceous through the middle Eocene (70 to 40 Ma), and faster cooling (~1 to 4 °C/Myr) from the middle Eocene at least through the Oligocene [Bauer et al., 2013]. The Malawi Rift in Tanzania (western branch of the EARS) records a broadly similar history as the Rwenzoris Mountains over that time interval, with slow cooling from the Cretaceous through the Paleocene, and more rapid cooling (~0.5 to 1°C/Myr) starting after ~40 Ma [van der Beek et al., 1998].

Sedimentologic, geomorphic, and geochronologic studies from different sectors of the EARS also suggest diachronous, yet partly overlapping rift-related exhumation in East Africa (Figure 7). For example, the Lake Rukwa region of the western branch and the rift basins west of Lake Turkana record an Oligocene onset of rifting [Morley et al., 1992, 1999; Roberts et al., 2012]. Our new data from the Elgeyo Escarpment combined...
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With field observations by Chapman et al. (1978) and geophysical data from the Kerio Basin [Mugisha et al., 1997] also suggest an earlier onset of extensional processes in this region, while the present-day morphologic characteristics of northern and central Kenya rifts are the result of renewed normal faulting after 12 Ma, coeval with extension recorded in the Central Tanganyika Basin [Chapman et al., 1978; Baker et al., 1988; Strecke et al., 1990], and the Albertine and southern Kenya rifts record the formation of extensional basin fills after approximately 9 and 7 Ma, respectively [Crossley, 1979; Pickford and Senut, 1994; Lezzar et al., 1996].

At first sight, the regional data regarding extension the EARS summarized in Figure 7 suggest an earlier onset of extension based on thermochronological data compared to stratigraphic information. This is probably an artifact, because sediments and volcanic rocks associated with Paleogene rifting have subsided, and coeval removal of rift-related deposits from exposed fault blocks and rift-shoulder areas precludes inspection of the early sedimentological vestiges of rifting. This problem is well emphasized by the Paleogene onset of rifting in the Kenyan Kerio Valley, where eroded sedimentary deposits from rift-shoulder areas have only been imaged by geophysical techniques [Mugisha et al., 1997].

In any case, the available thermochronologic and geologic data reveal a spatially disparate and diachronous evolution of Cenozoic rifting in East Africa, with clear differences in the onset of rifting in the western and eastern branches of the EARS. This spatiotemporal pattern of extension is inconsistent with tectonic models of rifting in East Africa that are based on a southward directed directed migration of volcanism and coeval extension [McConnell, 1972; Ebinger and Sleep, 1998; Ebinger et al., 2000; Nyblade and Brazier, 2002; Morley, 2010]. In light of the pronounced geophysical anomalies, evidence for mantle advection, and the evolution of dynamic topography associated with regional domal uplift [i.e., White and McKenzie, 1989; Simiu and Keller, 1997; Prodehl et al., 1997; Achauer and Masson, 2002; Mechie et al., 1997; Sepulcre et al., 2006; Moucha and Forte, 2011; Wichura et al., 2015], the timing of extension throughout East Africa likely reflects a large-scale, mantle-driven process that generated differential stresses [e.g., Crough, 1983; Zeyen et al., 1997] and the formation of rift basins in areas characterized by pronounced lithospheric and crustal-scale anisotropies and weaknesses [i.e., Ashwal and Burke, 1989; Ebinger and Sleep, 1998; Smith and Mosley, 1993; Smith, 1994]. As such, our new data from the Kenya Rift, combined with the synthesis of geological and thermo-chronological studies in East Africa, is compatible with recent numerical modeling results [Koptev et al., 2015] that predict a regionally overlapping initiation of amagmatic and magmatic rifting sectors in East Africa following the asymmetric impingement of a single mantle plume [i.e., Haldorsson et al., 2014] at the base of the lithosphere of the eastern sector of the Tanzania Craton.

6. Conclusions

Our new AFT and ZHe data from both escarpment and rift shoulders from the northern and central sectors of the Kenya Rift help to define two distinct stages of rapid cooling from ~65 to 45 Ma and from ~15 Ma to the present day, separated by a long phase of near-isothermal conditions or minor reheating. The initial stage of rapid cooling likely reflects the initiation of Cenozoic rifting, followed by a period of relative quiescence, and then renewed rift activity since the middle Miocene. While our thermal modeling results are consistent with those reported from the northern Kenya Rift and the Pangani Rift [e.g., the eastern branch of the EARS], they contrast markedly with those reported from the western branch (Rwenzori Mountains, and the Rukwa and Malawi rifts). As such, we suggest that the spatiotemporal evolution of rifting in the EARS is compatible with the impact of mantle upwelling, ensuing crustal uplift, and extensional fracture propagation guided by crustal heterogeneities, rather than being related to the southward progression of mantle-plume activity.

References
