

A low-temperature thermochronology investigation of the
Turkana Depression: Implications for the development of the
East African Rift System

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In loving memory of Delbert and Leonia Boone

Abstract

The Turkana Depression is one of the most important segments of the East African Rift System (EARS) for studying the onset of intracontinental rifting and the influence of pre-existing lithospheric heterogeneities on focussing later magmatism and strain. Defining the topographic lowlands separating the Ethiopian and East African Domes, the Turkana Depression hosts the earliest manifestations of EARS-related volcanism (Eocene) and basin formation (Paleogene), as suggested by seismic data. Here in northern Kenya and southern Ethiopia, the EARS is expressed by a diffuse region (~300 km) of deformation and highly attenuated crust (~20 km), in marked contrast to the narrow rift trends (~50 km) and thicker crust (~35-40 km) of the surrounding plateaux. The anomalous morphology and crustal architecture of the Turkana Depression may be attributed to earlier Cretaceous-early Paleogene Anza and South Sudan rifting. The similar age, geometry and gravity response of the ~NW-SE trending Anza and South Sudan rifts preserved to the east and west of Turkana has led to the hypothesis that these systems were once linked. However, due to the obscuring effect of subsequent volcanism and rifting and the scarcity of subsurface and geophysical data in the western Turkana Depression, the proposed connection is poorly constrained.

This study presents a low-temperature thermochronology (apatite fission track, apatite (U-Th-Sm)/He, and zircon (U-Th)/He) survey of the Turkana Depression, constraining the nature and extent of pre-EARS tectonism and the subsequent late Paleogene onset and propagation of EARS faulting in the region. Thermal history modelling shows that the Turkana basement records significant Cretaceous denudational cooling, coeval with significant Anza-South Sudan syn-rift sedimentation, suggesting that this area may have initially acted as a basement high and axial sediment source between the rift systems at that time. In the Late Cretaceous-early Paleogene however, parts of Turkana began to subside, in places accommodating up to 500 m of infill. This signified an important period of crustal thinning that may have facilitated the subsequent Eocene commencement of plume-related volcanism, marking the initiation of the EARS. Nonetheless, the discontinuous nature and relatively shallow depths of these depocentres argue against the Anza and South Sudan rifts having achieved a hard linkage within Turkana. The Eocene extrusion of voluminous lavas in the northern Turkana Depression coincided with the abandonment of these early depocentres,

possibly suggesting that dynamic topography associated with the arrival of a mantle plume was responsible for terminating early Paleogene subsidence in the region. Thermochronology data from the footwall of a major boundary fault in southern Turkana record a pronounced Eocene onset of cooling related to the development of the ~N-S Lokichar Basin. The Eocene nucleation of strain in the Turkana Depression, significantly predating faulting elsewhere in the EARS, highlights the importance of pre-existing rheological heterogeneities and mantle processes during the initiation of intracontinental rifting. By constraining the spatio-temporal commencement of ~E-W extension in East Africa, this study provides valuable insight into the causal mechanisms for EARS inception.

Declaration

This is to certify that

- i. The thesis comprises only my original work towards the PhD except where indicated in the Preface,
- ii. due acknowledgement has been made in the text to all other material used,
- iii. the thesis is less than 100,000 words in length, excluding tables, maps, bibliographies and appendices.

A handwritten signature in black ink that reads "Samuel Boone". The signature is written in a cursive style with a large, sweeping 'S' at the beginning and a distinct 'Boone' at the end.

Samuel C. Boone

Preface

This thesis contains material that is either published, in review or in preparation for future submission for publication. All data chapters of this thesis were written as free-standing papers suitable for submission to international scientific journals. Each chapter, therefore, consists of an abstract, introduction, geological setting, methodology, results, discussion and conclusion. Thus, a certain degree of unavoidable repetition (e.g. geological setting, methodology) exists. Reference lists and any supplementary material are presented at the end of the corresponding chapters. The data chapters are bookended by an Introduction, which includes a literature review that encompasses the historical and current literature, and Concluding Remarks, that synthesises the scientific findings resulting from this work.

At the time of submission, Chapter 2 represents an early draft prepared for submission to a short-format international scientific journal (e.g. *Geology*):

Boone, S. C., Kohn, B. P., Morley, C. K., Gleadow, A. J. W., Seiler, C., & Foster, D. A. (in prep.). Birth of the East African Rift System: Eocene nucleation of volcanism and faulting in the Turkana Depression.

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Co-authors of these manuscripts include academic supervisors of this PhD from the University of Melbourne, Prof. Andrew Gleadow, Prof. Barry Kohn and Dr. Christian Seiler, as well as Prof. Christopher Morley from Chiang Mai University in Thailand, Prof. David Foster at the University of Florida in the United States, Dr. Maria-Laura Balestrieri and Dr.

Giacomo Corti from the Istituto di Geoscienze e Georisorse, Consiglio Nazionale delle Ricerche in Italy, and Dr. Ling Chung from the University of Melbourne. The contributions of Prof. Andrew Gleadow and Prof. Barry Kohn were similar to those normally provided by advisors to a PhD student, including logistical and financial support, assistance during field work, assistance during data acquisition and interpretation, discussions on research strategy and comments on written work. Dr. Seiler aided in one of three field seasons in Kenya, commented on written material and gave general PhD and career advice. Prof. Morley aided in the project by supplying valuable insights into the geology of Turkana and comments on Chapters 2 and 3. Prof. Foster assisted in the first field season and provided comments on Chapters 2, 3 and 5. Dr. Balestrieri supplied the samples from Ethiopia for analyses presented in Chapter 4 and gave comments on the associated manuscript. Dr. Corti provided comments on Chapter 5. Dr. Chung assisted in data acquisition for Chapters 3 and 5. The project execution, data collection, interpretation of results, and writing of the manuscripts was the work of the present author.

During the PhD candidature, the present author also published two papers related to a separate project, which focussed on investigating the tectono-thermal evolution of southern Australia:

- Boone, S. C., Seiler, C., Reid, A. J., Kohn, B., & Gleadow, A. (2016). An Upper Cretaceous paleo-aquifer system in the Eromanga Basin of the central Gawler Craton, South Australia: evidence from apatite fission track thermochronology. *Australian Journal of Earth Sciences*, 63(3), 315–331. <https://doi.org/10.1080/08120099.2016.1199050>
- Hall, J. W., Glorie, S., Reid, A., Boone, S. C., Collins, A. S., & Gleadow, A. (2018). An apatite U-Pb thermal history map for the northern Gawler Craton, South Australia. *Geoscience Frontiers*. <https://doi.org/10.1016/j.gsf.2017.12.010>

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Introduction

Rifts play a fundamental role in the theory of plate tectonics, which along with convergent margins and transform faults define a continuous network of mobile belts that divide the Earth's surface into rigid plates [e.g. Wilson, 1965; Buck, 2004; Burke, 2011]. The breakup of continents along intracontinental rifts results in the creation of new oceans, which eventually diminish and close again, completing the so-called Wilson Cycle [Wilson, 1968]. Continental rift zones comprise regions of lithospheric stretching achieved through thinning of the brittle upper crust via normal faulting and attenuation of the lower ductile crust by viscous flow [e.g. Buck *et al.*, 1999; Lavier and Manatschal, 2006]. The distribution of strain varies between rift systems, depending on the thickness, composition and thermal state of the deforming lithosphere [e.g. Kuszniir and Park, 1987; Buck, 1991; Sokoutis *et al.*, 2007]. Extension is either distributed symmetrically throughout the ductile lower crust [‘pure-shear’ model, McKenzie, 1978] or asymmetrically, along lithospheric-scale shear zones [‘simple-shear’ model, Wernicke, 1985]. The distribution of extensional deformation and rift-magmatism may also be influenced by pre-existing lithospheric architecture, strength, thickness and compositional heterogeneities [e.g. Lister *et al.*, 1986, 1991; Sleep *et al.*, 2002; Ziegler and Cloetingh, 2004; Corti *et al.*, 2007, 2018a, 2018b]. Regardless of the distribution of strain, relatively dense mantle lithosphere rises upwards to replace the attenuating crust, enhancing subsidence in fault-bounded basins [McKenzie, 1978]. As the mantle lithosphere in turn undergoes necking and thinning, it is replaced by upwelling, hotter asthenosphere, transferring heat to the lithosphere, reducing rock density and causing regional dynamic uplift of the Earth's surface [e.g. Şengör and Burke, 1978]. The upward advection of mantle material beneath rift zones can generate basaltic melts by adiabatic decompression melting. These processes may be enhanced by upwelling mantle chemo-thermal anomalies, or plumes, where the anomalously high temperatures facilitate melting at greater depths, yielding voluminous flood volcanism and magma intrusion which may occur over time scales of up to

millions of years [e.g. *White and McKenzie*, 1989, 1995; *van Wijk et al.*, 2001]. Interaction between the upwelling asthenosphere and overlying lithosphere results in heat transfer, melt retention and metasomatism which may alter the overriding plate rheology, causing it to soften and thin [e.g. *Buck*, 2004; *Fischer et al.*, 2009; *Bialas et al.*, 2010; *Rooney et al.*, 2012a; *Schmeling and Wallner*, 2012; *Ebinger et al.*, 2017].

The Paleogene-Recent East African Rift Zone (EARS) is the pre-eminent modern example of a magma-assisted intracontinental rift zone on Earth, providing much of our understanding of early continental breakup processes [e.g. *Courtillot et al.*, 1999; *Ebinger and Casey*, 2001; *Ebinger and Scholz*, 2012]. Extending ~3,500 km southwards from the Red Sea to Mozambique, the EARS is considered to be an intra-cratonic prelude to oceanic opening (Figure 1.1) [*Morley et al.*, 1999a; *Chorowicz*, 2005]. The EARS is thought to have developed diachronously [e.g. *Torres Acosta*, 2015; *Vétel and Le Gall*, 2006] above one or multiple Paleogene mantle plumes [e.g. *Ebinger and Sleep*, 1998; *George et al.*, 1998; *George and Rogers*, 2002; *Rogers*, 2006; *Civiero et al.*, 2016]. Dynamic topography, magmatic underplating and volcanic build-up associated with the upwelling mantle thermal anomalies resulted in the formation of a contiguous series of broad uplifted regions that span much of eastern and southern Africa [e.g. *Nyblade and Robinson*, 1994; *Benoit et al.*, 2006; *Moucha and Forte*, 2011; *Pik*, 2011].

The large-scale forces driving present-day East African divergence, however, remain controversial. Some authors [e.g. *Forte et al.*, 2010; *Ghosh and Holt*, 2012] favour a model whereby rifting is driven primarily by basal shear tractions exerted by divergent mantle flow beneath East Africa. However, others suggest that buoyancy forces associated with the resulting lateral gradients in gravitational potential energy are largely responsible for driving modern extensional deformation in East Africa [e.g. *Coblentz and Sandiford*, 1994; *Coblentz et al.*, 1994; *Stamps et al.*, 2010, 2014, 2015]. Regardless, these forces are widely considered to be insufficient to rupture cold, thick lithosphere [e.g. *Kusznir and Park*, 1987; *Hopper and Buck*, 1993; *Buck*, 2004; *Stamps et al.*, 2015]. Thus, the forces responsible for initiating the EARS remain enigmatic [*Buck*, 2004; *Kendall et al.*, 2016]. Furthermore, debate continues as to the relative significance of pre-existing rheological heterogeneities and plume magmatism in the initial focusing of extensional strain [e.g. *Bellahsen et al.*, 2003; *Vétel and Le Gall*, 2006; *Ebinger and Sleep*, 1998; *Corti*, 2018a, 2018b].

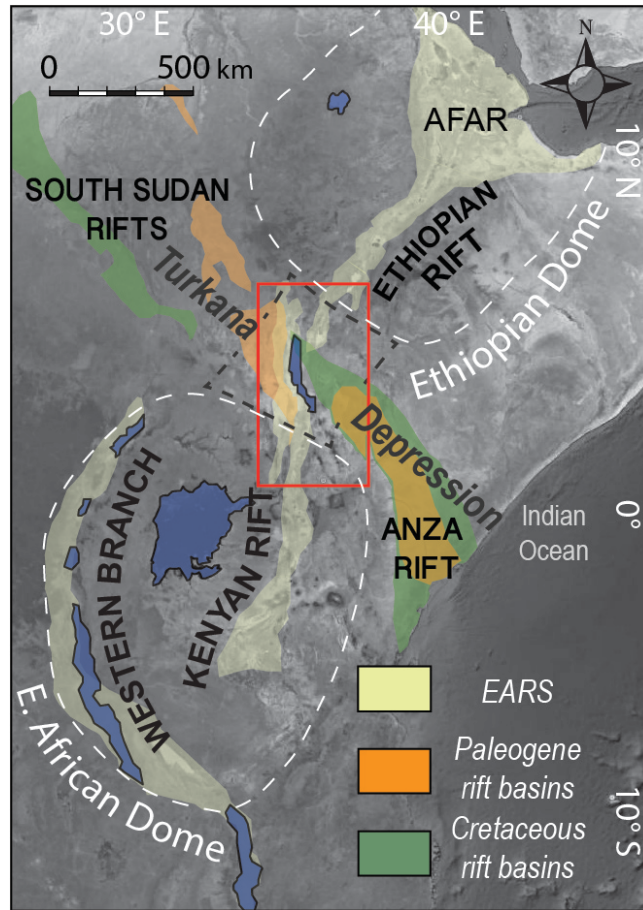
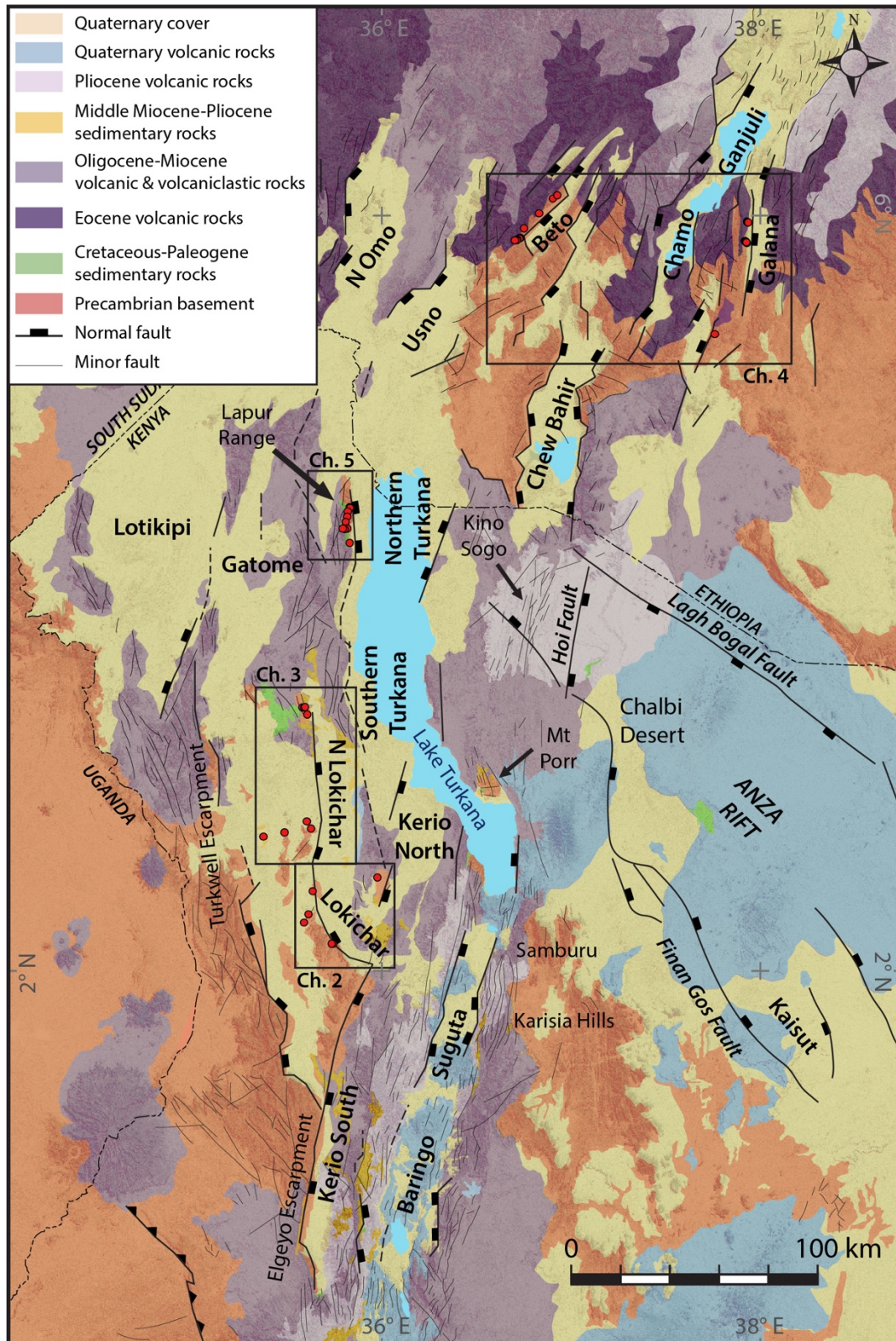


Figure 1.1. Topography of East Africa, distribution of the Turkana Depression (grey dashed line), the Ethiopian and East African Domes (white dashed line), the Cretaceous-Paleogene Anza and South Sudan rifts and the EARS, and the location of Figure 1.2 (red box).

An ideal location to investigate the early stages of EARS development is the Turkana Depression (Figure 1.1), a structurally complex and long-lived segment of the EARS that hosts both the earliest manifestation of rift-related magmatism [~ 45 Ma, Rooney, 2017] and earliest known $\sim$ N-S trending extensional basins in East Africa [Paleogene, Morley *et al.*, 1992, 1999b]. Furthermore, its coincident orientation with a $\sim$ NW-SE trending series of discontinuous Cretaceous-early Paleogene rift basins makes it an optimal region to study the influence of pre-existing lithospheric heterogeneities on the localisation of magmatism and strain. Resolving the driving forces and mechanisms responsible for rift inception and strain nucleation requires a thorough understanding of the chronology of faulting and lithospheric modification. However, the spatio-temporal evolution of extensional deformation in the Turkana Depression remains poorly constrained due to a lack of surface exposure of earliest syn-rift strata, the obscuring effects of extensive Neogene-Recent volcanic rocks and Quaternary sedimentary cover, limited subsurface data and a dearth of geological observations from eastern South Sudan.

This thesis aims to better our understanding of the tectono-thermal evolution of the Turkana Depression by presenting a low-temperature thermochronological survey of exposed Precambrian basement and Cretaceous-Neogene sedimentary rocks throughout the region (Figure 1.2). Combined apatite fission track (AFT), apatite (U-Th-Sm)/He (AHe) and zircon (U-Th)/He (ZHe) analyses provides an alternative approach to investigate landscape evolution and basin formation by dating periods of cooling and heating as proxies for rift-related erosion and burial processes [e.g. *Fitzgerald*, 1992; *van der Beek et al.*, 1994; *Ehlers et al.*, 2001; *Braun*, 2005; *Stockli*, 2005; *Mackintosh et al.*, 2017]. Here, I present combined AFT, AHe and ZHe analyses from 43 crystalline basement and five Cretaceous-Neogene sedimentary rock samples collected throughout northern Kenya and southern Ethiopia. Inverse thermal history modelling of these data reveals a complex Cretaceous-Recent thermal evolution, which provide new constraints on the nature and extent of Cretaceous-Paleogene ~NE-SW extension in the Turkana Depression, the possible topographic response to the Eocene onset of plume-related magmatism in the region, the initiation of extensional deformation associated with EARS inception and the propagation of strain during late Paleogene-Recent time. These findings provide important insights into the forces driving the initiation of the EARS, the role of inherited lithospheric architecture and crustal fabrics on rift nucleation and strain propagation, and the influence of earlier rifting phases on the lithospheric response to later crustal dilation.



1.1 Geological Background

Africa has been a centre of continental accretion since the late Proterozoic, establishing its modern configuration in the Late Neoproterozoic-Early Paleozoic Pan-African Orogeny and culminating in the amalgamation of the Gondwana supercontinent. The related ocean closure, arc and microcontinent accretion and the suturing of continental fragments resulted in a series of orogenic and mobile belts throughout Gondwana, such as the East African Orogen (Figure 1.3). The Mozambique Belt (~900-470 Ma), extending from southern Ethiopia to Mozambique, comprises the southern East African Orogen and consists of juvenile Proterozoic passive margin sediments intercalated with Archean gneisses [Begg *et al.*, 2009]. Neoproterozoic tectonism associated with the accretion of Gondwana has largely controlled the distribution of subsequent tectonic belts throughout Palaeozoic to Recent time [Boger and Miller, 2004; Collins and Pisarevsky, 2005].

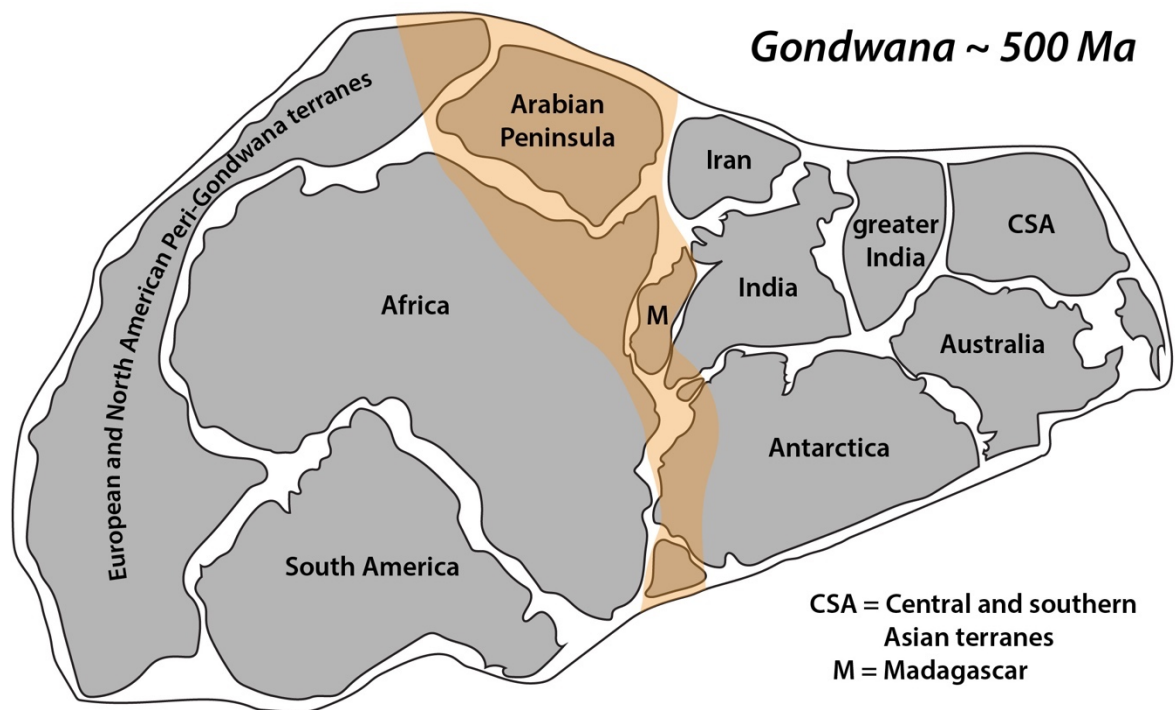


Figure 1.3. Reconstruction of the Gondwana supercontinent illustrating the central location of Africa and the East African Orogen (highlighted in orange). Adapted from Boger *et al.* [2015].

1.1.1 Breakup of Gondwana and the Cretaceous-Paleogene Anza and South Sudan rifts

The early Paleozoic was marked by pervasive postorogenic magmatism, shearing, major uplift and deep seated exhumation of mobile belts throughout Gondwana [Unrug, 1997]. Breakup of Gondwana began in earnest during Late Paleozoic-Jurassic Karoo rifting that affected much of southern and eastern Africa, as evidenced by widespread intracratonic rift basins [Kreuser, 1995].

Beginning in the Cretaceous, a series of rift basins developed throughout northern Africa in order to accommodate opening and spreading in both the Atlantic and Indian Oceans [Reeves *et al.*, 2016]. In East Africa, these included the ~NW-SE trending Anza Rift and South Sudan rift system (Figures 1.1 and 1.4). The Anza Rift extends ~ 600 km NW from the Kenyan coast to Lake Turkana (Figure 1.1), hosting highly faulted depocentres up to 7 km thick [Bosworth and Morley, 1994; Morley *et al.*, 1999c; Vétel and Le Gall, 2006]. The asymmetric Anza graben, bound by the Lagh Bogal and Finan Gos fault systems, is estimated to have accommodated 60 km of bulk extension [Dindi, 1994]. Further west, a series of similarly NW-SE trending Cretaceous-Paleogene basins, known collectively as the South Sudan Rifts, narrow as they approach the South Sudanese-Kenyan border [Browne and Fairhead, 1983]. The geometry of this termination however, remains unclear due to a general data gap in eastern South Sudan and the obscuring effect of subsequent Paleogene-Recent tectonism and volcanic activity in northern Kenya [Guiraud and Bosworth, 1997]. Despite this uncertainty, the compatible geometries, coeval evolution and similar gravity signals of the South Sudan and Anza rifts suggest that a possible link between the two systems may exist in Turkana [Schull, 1988; Ibrahim *et al.*, 1991; Bosworth, 1992; Ebinger and Ibrahim, 1994; Hendrie *et al.*, 1994]. Others, however, point to the general lack of Cretaceous-early Paleogene strata in Turkana and the high-gradient isobath pattern of pre-Cretaceous basement east of Lake Turkana to argue that the Anza and South Sudan Rifts were initially disconnected [Vétel and Le Gall, 2006]. Instead, Vétel and Le Gall [2006] suggest that the NW extent of the Anza graben terminated at a syn-rift faulted boundary along the Hoi Fault (Figure 1.2).

1.1.2 Paleogene-Recent East African Rift System

The ~N-S trending EARS is composed of two branches, the eastern branch forming the Ethiopian and Kenyan Rifts and the arcuate western branch comprised of a series of elongate, fault-bounded lakes running from Uganda to Mozambique. In general, these rift zones follow the trend of the Neoproterozoic Mozambique Belt [Morley *et al.*, 1999a; Kröner and Stern,

2005]. Brittle deformation in the EARS is generally localised within narrow rift zones (up to 80 km wide) that traverse the broad, uplifted Ethiopian and East African Domes [Figure 1.1; e.g. *Chorowicz, 2005; Ebinger, 2005; Corti, 2009; Ebinger and Scholz, 2012*]. These plateaux are separated by the topographically subdued lowlands of the Turkana Depression, a diffuse zone of deformation up to 300 km wide in northern Kenya and southern Ethiopia that is characterised by an extensive Bouguer gravity ridge anomaly, relative to broad lows centred on the neighbouring domes, and crustal thicknesses as low as 20 km (Figures 1.1 and 1.2) [*Ebinger and Ibrahim, 1994; Mariita and Keller, 2007*].

Rifting in the EARS is thought to have developed above one or multiple Paleogene mantle plumes [e.g. *Ebinger and Sleep, 1998; George et al., 1998; George and Rogers, 2002; Rogers, 2006; Civiero et al., 2016*]. Geochemical data from volcanics [e.g. *Pik et al., 2006; Furman et al., 2007; Rooney et al., 2012, 2017*] and the often significant hiatus between the onset of magmatism and strain [*Morley et al., 1992; Ebinger et al., 1993, 2000; Pik et al., 2008*] point to a mantle plume origin for the voluminous flood basalts of East Africa [*Hendrie et al., 1994; Morley, 1994; Ebinger and Sleep, 1998; Furman et al., 2006*]. The continuity, location and depth of the asthenospheric thermal upwelling(s) however, have been hotly debated [*Ebinger and Sleep, 1998; Weeraratne et al., 2003; Simmons et al., 2007; Li et al., 2008; Chang and Van der Lee, 2011; Fishwick and Bastow, 2011*]. Yet more recently, authors advocate for one extensive, elongate asthenospheric upwelling rising from the core-mantle boundary beneath southern Africa and impinging under the East African lithosphere [e.g. *Grand et al., 1997; Ritsema et al., 1999; Furman, 2007; Hansen et al., 2012*]. This so-called ‘African Superplume’ may split into several smaller hot spots in the shallow asthenosphere as it is channelled along the base of the lithosphere [e.g. *Ebinger and Sleep, 1998; Sleep et al., 2002; Koptev et al., 2017*].

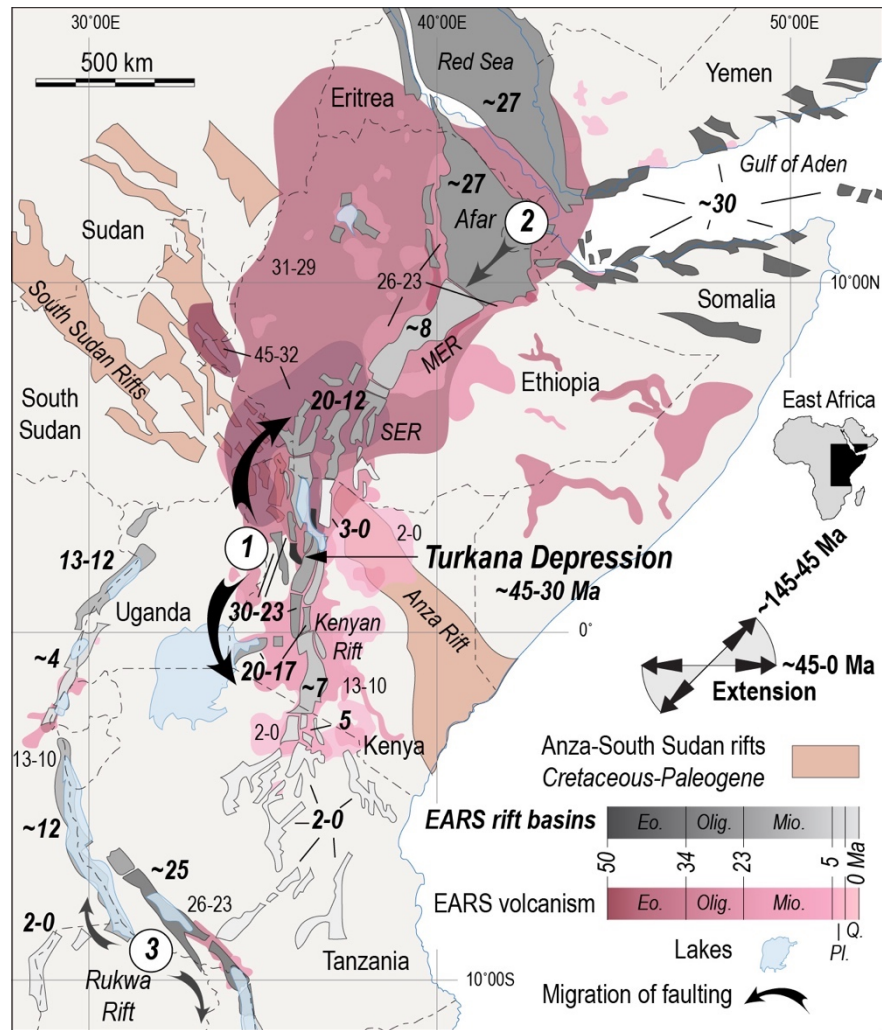


Figure 1.4. Chronology of Cretaceous-Recent rift basin formation and EARS-related volcanism in East Africa. Colours of EARS basins (grey) and volcanics (red) darken with age. EARS volcanism and faulting is suggested to have nucleated within inherently weak lithospheric zones (1: Turkana Depression, 2: Afar, 3: Rukwa Rift) before propagating north and south. Distribution of basins and surface volcanics, and corresponding ages from Macgregor [2015], Purcell et al. [2017], Rooney et al. [2017], and references therein. MER: Main Ethiopian Rift; SER: Southern Ethiopian Rift.

The EARS initiated during the Paleogene in discrete cells within first-order transverse inherited weakness zones (Figure 1.4) [Vétel and Le Gall, 2006; Torres Acosta et al., 2015]. Rifting is thought to have first nucleated in the northern Turkana Depression, as marked by the emplacement of widespread volcanism in southwest Ethiopia and northernmost Kenya ca. 45 Ma [Ebinger et al., 2000; Vétel and Le Gall, 2006]. Based on the predominant ~N-S orientation of major basin-bounding faults and the occurrence of N-S striking Oligocene dikes in the Lapur Range area [Vétel and Le Gall, 2006], extension during this early phase of rifting is thought to have been ~E-W. Thus, these deformational features represent an important rotation of the paleostress field, at least locally, from the general ~NE-SW

direction which dominated during Cretaceous-early Paleogene Anza and South Sudan rifting [Bosworth and Morley, 1994; Morley *et al.*, 1999c; Vétel and Le Gall, 2006]. By the Oligocene, disparate rift nucleation cells had also formed in the Gulf of Aden-Red Sea-Afar region [31-27 Ma, e.g. Hughes *et al.*, 1991; Leroy *et al.*, 2004; Rooney, 2017] and the Rukwa Rift sector in the southern EARS [~25 Ma, Roberts *et al.*, 2010, 2012]. Subsequent propagation of rifting both north and south eventually led to the formation of an almost continuous rift system that extended over nearly the length of the African continent by the Pliocene [Macgregor, 2015; Purcell, 2017].

1.1.3 Turkana Depression

In this study, the Turkana Depression is taken to include the entire zone of diffuse deformation occupying the area between the East African and Ethiopian Domes, extending northwards from the Turkwell Escarpment and Lokichar Basin of Kenya and including the Broadly Rifted Zone (BRZ) of the southern Ethiopia Rift (Figure 1.2). Here, extensional deformation is manifested in a series of basins and ranges up to 300 km in width, which is in stark contrast to the generally narrow rift zones (up to 80 km) transecting the surrounding uplifted plateaux. The anomalous rift morphology and crustal architecture of the Turkana Depression are often attributed to an earlier failed Cretaceous-early Paleogene rifting phase recorded in the neighbouring ~NW-SE trending Anza-South Sudan rifts systems [Pérez-Gussinyé *et al.*, 2009; Brune *et al.*, 2017], which are thought to have once extended into and connected within Turkana. This allows for the possibility that the contiguous high plateaux of East Africa represent one elongate uplifted region extending from southern Africa to the Red Sea [Nyblade and Robinson, 1994; Ritsema *et al.*, 1998]. Indeed, Benoit *et al.* [2006] showed that the anomalously thin crust of the modern Turkana region (~20 km) is unable to isostatically support long-wavelength topography.

Much of our understanding of the structural architecture of the Turkana Depression upper crust derives from a series of seismic reflection surveys carried out by the Shell and Amoco Oil companies in the 80's and 90's [Morley *et al.*, 1992; Morley *et al.*, 1999b, Wescott *et al.*, 1999]. However, due to a limited number of exploration wells and poor surface exposure of pre-Neogene syn-rift strata, the localisation and subsequent migration of extensional basin formation in the Turkana Depression, particularly prior to the middle Miocene, are poorly constrained [Morley *et al.*, 1999b].

The earliest known manifestation of strain in the EARS is the late Eocene?-Oligocene to Miocene Lokichar Basin of southern Turkana [Morley, 1999; Morley *et al.*, 1999b; Vétel and Le Gall, 2006]. This ~N-S striking half-graben hosts a ~7 km thick series of late Paleogene-middle Miocene fluvial and lacustrine sediments, however the exact nature and age of the oldest strata of the Lokichar Basin is unknown due to a lack of surface exposure and exploration well data from the deeper parts of the basin [Morley *et al.*, 1992, 1999b]. Low-temperature thermochronology analyses of the Lokichar Basin margin (Chapter 2) provide new constraints on the timing of strain localisation in Turkana. Evidence of more widespread Eocene?-Oligocene extensional deformation is found on the Kino Sogo plateau of eastern Turkana, where significant expansion (up to 1100 m) of Eocene-Oligocene basalts is documented across a N-S fault [Morley, 1999; Morley *et al.*, 1999b; Vétel and Le Gall, 2006]. Seismic reflection data also reveals two ~N-S trending basins in the subsurface of the Lotikipi and Gatome plains of western Turkana underlying an 1800 m thick succession of olivine basalts as old as 33 Ma [Zanettin *et al.*, 1983], thought to be Cretaceous to Paleogene in age [Wescott *et al.*, 1999]. Zones of Eocene faulting without volcanism are also found in easternmost South Sudan [Coffin and Rabinowitz, 1988]. However, the nature and age of these basins are unknown due to the lack of well data and surface outcrop in both the Lotikipi-Gatome and eastern South Sudan area [Tiercelin *et al.*, 2004]. Seismic data also reveal a ~N-S sub-basin beneath the late Miocene-Pliocene North Lokichar Basin of central Turkana, which is thought to be late Paleogene-middle Miocene in age [Morley, 1999; Morley *et al.*, 1999b]. Thermal history modelling of exposed basement rocks in the North Lokichar region (Chapter 3) provides new constraints on the Paleogene-Recent tectono-thermal evolution of this region.

After nucleating in northern Kenya, west of Lake Turkana, extensional deformation propagated both north into the Broadly Rifted Zone (BRZ) of southern Ethiopia by ~20-18 Ma (Chapter 4) [Ebinger *et al.*, 2000; Bonini *et al.*, 2005; Pik *et al.*, 2008] and south into the Kenyan Rift around 13 Ma [e.g. Mechie *et al.*, 1997; Morley *et al.*, 1999c; Wichura *et al.*, 2011; Macgregor, 2015] or possibly earlier in the late Oligocene-early Miocene [e.g. Mugisha *et al.*, 1997]. Within the Turkana Depression, strain simultaneously migrated eastwards during Neogene-Recent time [Morley *et al.*, 1999a; Vétel and Le Gall, 2006], developing a series of extensional basin beneath the modern Lake Turkana in the middle Miocene (Chapter 5), and eventually forming the arcuate Kino Sogo Fault Belt to the east of Lake Turkana [< 3 Ma, Vétel *et al.*, 2005]. Surface volcanism within Turkana exhibits a

similar eastward younging pattern [e.g. *McDougall and Brown*, 2009], perhaps suggesting an eastward migration of the underlying mantle thermal anomaly over time [*Ebinger et al.*, 2000]. Alternatively, the eastward propagation of rifting may have been caused by strain hardening [*Morley et al.*, 1992].

1.2 Low-temperature thermochronology

Low-temperature thermochronology involves the application of thermally sensitive radioisotopic dating techniques to constrain the time-temperature history for a given mineral at relatively low temperatures ($< 300\text{ }^{\circ}\text{C}$). For each individual radioisotopic decay scheme, thermally sensitive daughter products are produced via nuclear decay of radioactive parent isotopes, at a rate proportional to the concentration and decay rate of the parent [*Reiners et al.*, 2005]. The radiogenic daughter products are then retained below the closure temperature specific to that decay scheme [e.g. *Dodson*, 1973]. As temperature can be used as a proxy for crustal depth, low-temperature thermochronology data may provide estimates of the timing and rates of geological processes which disturb the upper crustal thermal regime over geological timescales, such as exhumation, burial, and magmatically- or gravity-driven changes in basal heat flow [*Ehlers*, 2005; *Reiners et al.*, 2005].

This study employs the widely used apatite fission track (AFT), zircon (U-Th)/He (ZHe) and apatite (U-Th-Sm)/He (AHe) thermochronology systems. The ubiquitous nature of apatite [$\text{Ca}_5(\text{PO}_4)_3(\text{F}, \text{Cl}, \text{OH})$] and zircon [ZrSiO_4] in many rock types and relative enrichment in the U, Th and in the case of AHe analysis, Sm parent nuclides make these ideal systems for constraining the low-temperature thermal evolution of continental crustal rocks. The combined temperature sensitivities of the ZHe [$< 50\text{--}220\text{ }^{\circ}\text{C}$, *Guenther et al.*, 2013; *Johnson et al.*, 2017], AFT [$\sim 60\text{--}120\text{ }^{\circ}\text{C}$, *Gleadow and Duddy*, 1981; *Green et al.*, 1989] and AHe [$\sim 30\text{--}90\text{ }^{\circ}\text{C}$, *Flowers et al.*, 2009] thermochronometers, when performed on the same rock sample, provide a more complete record of cooling over a temperature range of $\sim 30\text{--}220\text{ }^{\circ}\text{C}$, corresponding to typical crustal depths of $\sim 1\text{--}10\text{ km}$ [e.g. *Farley and Stockli*, 2002]. However, the constrained crustal depth range may vary depending on the geothermal gradient of the study area.

Low-temperature thermochronology data can be forward or inversely modelled using thermal history modelling software and suitable annealing and diffusion algorithms to reconstruct the

most probable time-temperature histories of a sample (Figure 1.5) [Ketcham *et al.*, 2007; Gallagher, 2012]. In this work, AFT data, along with ZHe and AHe results when available, were jointly inverted using the *QTQt* thermal history modelling software [Gallagher, 2012]. *QTQt* employs a Bayesian transdimensional Markov Chain Monte Carlo (MCMC) approach during low-temperature thermochronology data inversion, producing a range of accepted time-temperature reconstructions [Gallagher *et al.*, 2009; Gallagher, 2012]. A user-defined time-temperature space, annealing and diffusion information, and thermal history constraints (e.g. present-day surface temperature) are treated as prior probability information in the Bayesian approach [Gallagher, 2012]. The MCMC algorithm produces a series of random time-temperature histories that are either accepted or rejected based on the observed data [Vermeesch and Tian, 2014; Wildman *et al.*, 2016]. The *QTQt* software is well-suited for this study due to its ability to robustly model large, complex datasets [Vermeesch and Tian, 2014] and invert AHe and ZHe data from fragmented crystals using the fragmentation model of Brown *et al.* [2013]. Details of the modelling protocol and input parameters, such as thermochronology data, AFT compositional information, diffusion and annealing kinetic models, and time-temperature constraints, are presented in Chapter 2-5.

A more detailed discussion of the AFT, ZHe and AHe methodologies are presented in each of the four data chapters, along with in-depth synopses of the sample preparation and analytical procedures. Furthermore, descriptions of how these methods are used to constrain tectonic and surface processes are presented in the context of the specific research questions addressed in each of the corresponding chapters.

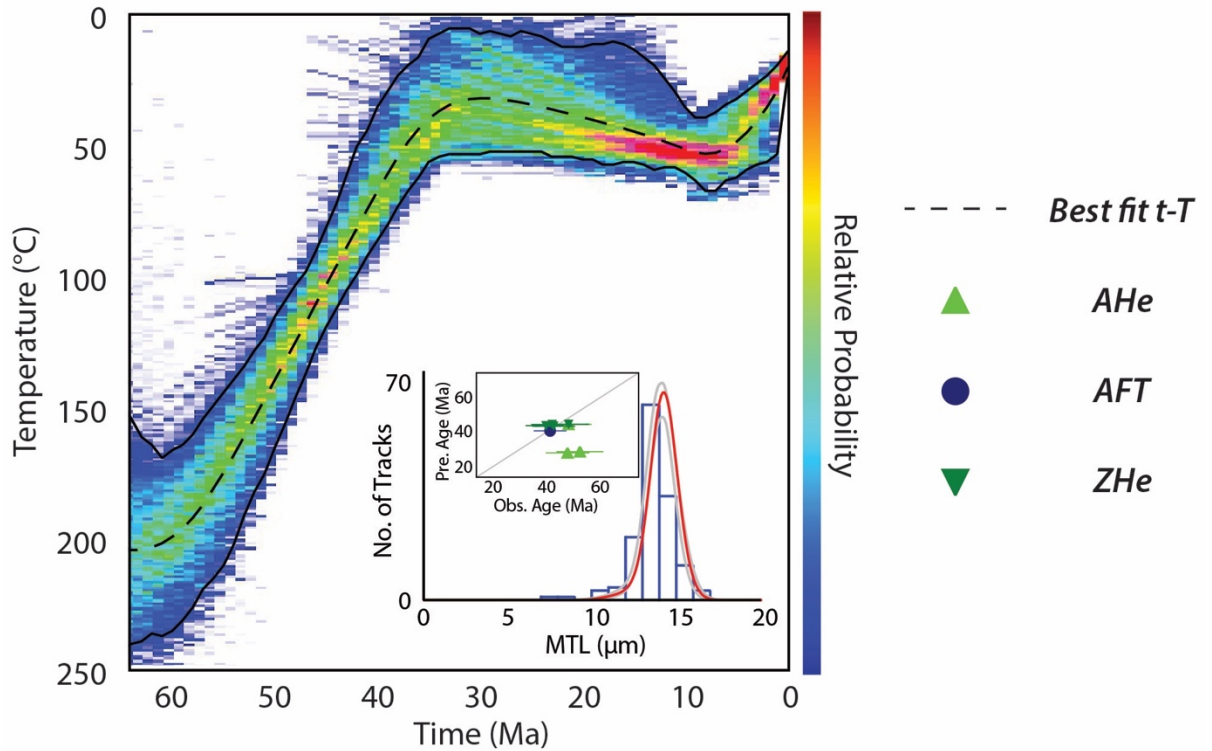


Figure 1.5. An example of a thermal history model of ZHe, AFT and AHe data generated using the software *QTQt* [Gallagher, 2012]. The colour map represents the posterior-probability distribution of accepted time-temperature paths, with warmer colours indicating higher probabilities. Black solid lines represent the 95% confidence interval limits. The data fit for each thermal history model is illustrated by panel inserts showing observed versus predicted thermochronology ages and observed confined track distributions overlain by the predicted distributions (red curve) and the 95% confidence interval (grey curves) of the predicted distributions.

1.2.1 Thermochronological insights into upper crustal thermal processes

Thermochronology analyses of exposed basement rocks provide an alternative method to decipher a record of the geological history that may have since been exhumed and eroded away, and thus undetectable by conventional field observation. In highly active tectonic setting, such as intracontinental rifts, the creation or destruction of high amplitude-short wavelength topography, rapid periods of denudation, and significant changes in surface relief often result in isotherm perturbations [e.g. *Stüwe et al.*, 1994; *Mancktelow and Grasemann*, 1997; *Braun*, 2005; *Braun et al.*, 2016].

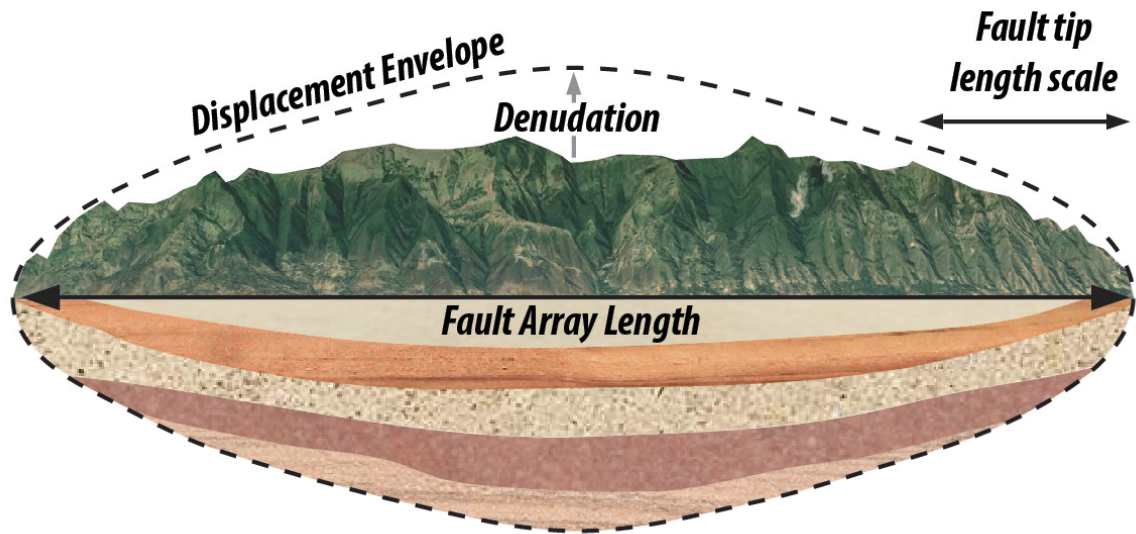


Figure 1.6. Schematic cross-section of a crustal-scale normal fault array. The fault array displacement envelope is illustrated by a dashed line, which defines the amount of footwall uplift and hanging wall subsidence. These in turn determine the development of bedrock topography in the form of a mountain range and a sedimentary basin, respectively. Denudation of the mountain ridge is defined by the difference between the displacement envelope and the footwall topography. Footwall relief and basin subsidence is low in the fault tip region. The total amount of relief is erosionally limited above some threshold, thought to be determined by the regional spacing of major faults and the elevation of base level on both flanks of the range [Densmore et al., 2005]. Thus, beyond a characteristic fault tip length scale, footwall relief is nearly uniform. Adapted from Densmore et al. [2004].

Low-temperature thermochronology can therefore be utilised in extensional settings to constrain periods of cooling and heating associated with uplift and exhumation of normal fault footwall blocks and burial and subsidence of the hanging wall [Fitzgerald, 1992; Noble et al., 1997; Ehlers and Chapman, 1999; Ehlers et al., 2001; Ehlers and Farley, 2003; Seiler et al., 2009, 2011]. The associated isotherm fluctuations which result from extensional basin formation produce low-temperature thermochronology spatial trends characterised by relatively younger ages in the footwall and older ages in the hanging wall [Ehlers et al., 2001; Ehlers, 2005]. Within the up-thrown footwall block, thermochronometric ages decrease with increased proximity to the normal fault trace, where exhumation rates are highest due to footwall tilt [Ehlers et al., 2001]. Along strike, fault surface displacement will vary, increasing from zero at the fault tips to some maximum near the fault centre [Barnett et al., 1987], thus defining a displacement envelope (Figure 1.6). The difference between the displacement envelope and normal footwall topography is equivalent to the amount of denudation [Densmore, 2004]. Along-strike spatial and temporal variations in footwall cooling histories of exposed boundary fault scarps can also be utilised to decipher the

structural evolution of basin bounding fault arrays (Figure 1.7) [Krugh, 2008; Curry *et al.*, 2016; Mortimer *et al.*, 2016; Brown *et al.*, 2017]. Such an approach is employed in Chapter 4 to examine the structural evolution of the southern Ethiopian Rift.

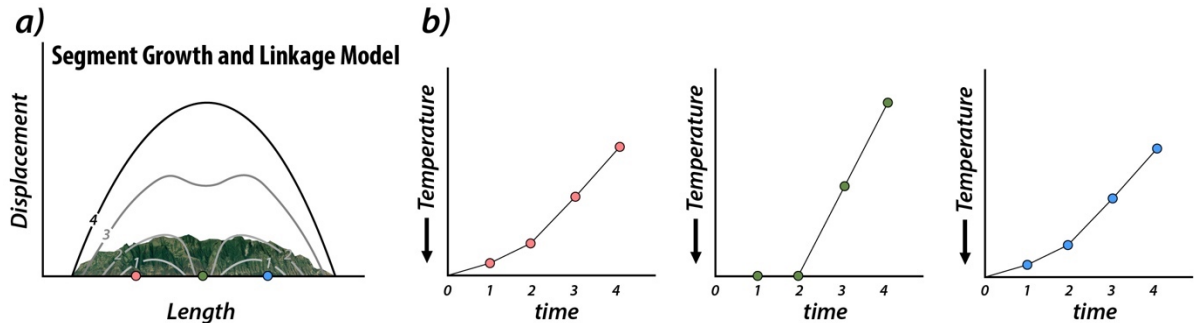


Figure 1.7. Illustration of how the structural evolution of a normal fault array can be reflected as along-strike spatial and temporal variations in footwall cooling histories. **a)** Idealised fault displacement versus length plot for the segment growth and linkage fault array evolution model, where multiple (in this case two), initially independent fault segments grow via tip propagation before interacting and linking to form a through-going fault array [Anders and Schlische, 1994; Cartwright *et al.*, 1995; Gawthorpe and Leeder, 2000; Gupta *et al.*, 1998; Peacock and Sanderson, 1994; Trudgill and Cartwright, 1994]. Displacement versus length plot adapted from Kim and Sanderson [2005]. **b)** Theoretical cooling histories for three basement rock samples (red, green and blue circles) collected along strike from the base of the fault scarp. Note, temperature increases down the y-axis while time increases along the x-axis, following the convention of thermal history models presented in low-temperature thermochronology studies (see Figure 1.5). Relative temperatures for each rock sample are shown for time periods 0-4, corresponding to the four stages of fault growth illustrated in the D/L plot, whereby temperature is assumed to be inversely proportional to displacement. In nature, temporal and spatial variations in along-strike footwall cooling histories may be further complicated by factors such as fault strength heterogeneities [Cowie and Shipton, 1998], the efficacy of surface processes [Densmore *et al.*, 2004; Densmore and Guidon, 2005] and antecedent topography [Densmore *et al.*, 2009].

The timing, amount and rate of rift-related footwall denudation resulting from normal faulting can be estimated from the initiation, magnitude and rate of observed cooling in best-fit thermal history models [e.g.; Gallagher, 1995]. However, the determination of denudation magnitude and rate require an estimate of paleo-geothermal gradients [Kohn *et al.*, 2005]. In the case of the Turkana Depression, even the present-day upper crustal thermal regime is poorly constrained, let alone in the geological past. The sole published heat flow measurement from Turkana [42 ± 5 °C/km, Morley *et al.*, 1999b] suggests that the present-day geothermal gradient in Turkana is high, consistent with 3D thermal modelling of the Kenyan lithosphere [Sippel *et al.*, 2017]. This is likely due to the significantly thinned crust of Turkana [~ 20 km, Benoit *et al.*, 2006] and the presence of a mantle thermal anomaly beneath the region [e.g. Nyblade, 2011; Hansen *et al.*, 2012]. However, the chronology of

crustal attenuation and mantle plume activity in the Turkana Depression are very poorly understood. Thus, particular care is taken when interpreting periods of observed reheating (see Chapter 3).

1.2.2 Previous Low-Temperature Thermochronology Studies in East Africa

Previous low-temperature thermochronological studies in East Africa have shown that the crystalline basement has experienced periodic episodes of rapid denudation and cooling from Mesozoic-Recent times [Abebe *et al.*, 2010; Balestrieri *et al.*, 2016; Bauer *et al.*, 2012, 2013; van der Beek *et al.*, 1998; Foster and Gleadow, 1992, 1993, 1996; Noble *et al.*, 1997; Philippon *et al.*, 2014; Pik *et al.*, 2003, 2008; Spiegel *et al.*, 2007; Torres Acosta *et al.*, 2015; Wagner *et al.*, 1992]. These studies are largely based on AFT analyses, with the majority focussed on the tectono-thermal evolution of Kenya (Figure 1.8). The oldest AFT ages (>250 Ma) are found in Archean cratonic rocks, such as those of the Tanzanian Craton exposed in southwest Kenya [Wagner *et al.*, 1992]. In Kenya, up to 2 km of early Cretaceous (>120 Ma) denudational cooling was documented in the Karisia Hills (Figure 1.2), possibly associated with the development of the neighbouring Anza Rift [Foster and Gleadow, 1992]. A more widespread and pervasive late Cretaceous-Paleogene (~70-50 Ma) cooling event is recorded throughout southern Ethiopia and Kenya [Balestrieri *et al.*, 2016; Foster and Gleadow, 1992, 1993, 1996; Noble *et al.*, 1997; Philippon *et al.*, 2014; Spiegel *et al.*, 2007; Torres Acosta *et al.*, 2015], interpreted as reflecting uplift and denudation of the Anza Rift shoulders and adjacent hinterlands.

Surprisingly, few thermochronology studies from East Africa record little, if any, Miocene-recent cooling associated with the development of the EARS, suggesting rift-shoulder uplift has been insufficient to exhume rocks which have cooled through the temperature sensitivity ranges of the AFT and AHe systems. Only studies from the southern Ethiopian Rift report Neogene thermochronometer ages [Abebe *et al.*, 2010; Balestrieri *et al.*, 2016; Pik *et al.*, 2008; Philippon *et al.*, 2014], perhaps suggesting a difference in rift development and/or thermal regime between the Ethiopian and Kenya Rifts (see Chapter 5). In Kenya, only studies that have employed the lower temperature AHe system report a degree of rift-related Neogene cooling as recorded by thermal history modelling [Spiegel *et al.*, 2007; Torres Acosta *et al.*, 2015]. Prior to the study presented here, no low-temperature existed from the Turkana region of northern Kenya.

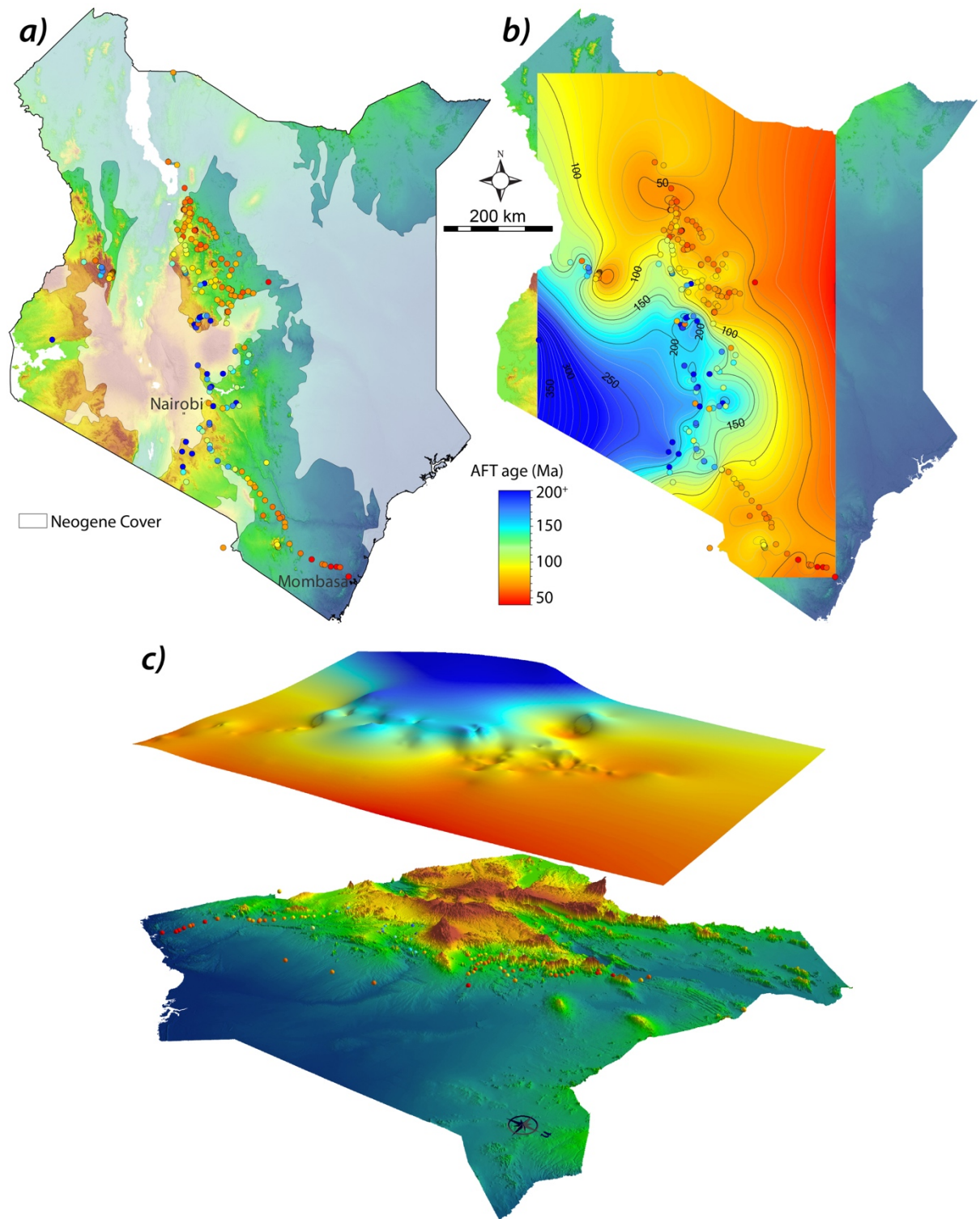


Figure 1.8. Previously published Kenyan AFT data. AFT ages are represented by colour, with colours becoming cooler with increasing ages. **a)** AFT ages superimposed on a digital elevation model (DEM) [NASA, 2000]. The distribution of Neogene cover in Kenya is showed in light grey. **b)** Interpolation of previously published Kenyan AFT ages. **c)** 3D ‘topographic’ perspective of AFT data interpolation above Kenyan DEM. AFT ages (Ma) were compiled from Gleadow [1980], Wagner et al. [1992], Foster & Gleadow [1992, 1993 and 1996], Noble et al. [1997], Spiegel et al. [2007], and Torres Acosta et al., [2015]. DEM sourced from NASA [2000].

1.3 Research Questions

This research aims to constrain where, when and how strain localised and propagated within the Turkana Depression since Cretaceous time by the application of AFT, ZHe and AHe thermochronology and thermal history modelling to exposed Precambrian crystalline basement and Cretaceous-Paleogene sedimentary rocks. These data, combined with inversely modelled time-temperature reconstructions, provide new insights into the tectono-thermal evolution of the upper Turkana crust associated with multiple phases of extensional deformation and plume-magmatism.

In particular, these data will be used to address the following key research questions:

- How does the pattern of cooling and reheating preserved in basement rocks from Turkana Depression relate to the topographic and tectonic evolution of the Turkana Depression?
- Did the Anza and South Sudan rifts systems once extend into and connect within Turkana?
- When was the Turkana crust attenuated? And what does that mean for the ability of the Turkana crust to have isostatically supported large-scale topography in the Paleogene?
- If significant crustal attenuation post-dates the Eocene onset of plume magmatism in the region, did a period of detectable thermal uplift accompany the arrival of a mantle thermal anomaly beneath the Turkana Depression?
- When and where did extensional strain associated with the EARS nucleate in the Turkana Depression? How was this affected by the earlier rifting history of the region? And how does the initiation of rifting in the Turkana Depression relate to the inception of the greater EARS?
- What does the late Paleogene-Recent rift evolution of the Turkana Depression tell us about the forces driving rifting during the early stages of EARS development?
- What is the influence of pre-existing crustal fabrics and lithospheric architecture on the localisation and propagation of strain in the Turkana Depression?

1.4 Thesis outline

In this thesis, four data chapters are presented which focus on constraining the tectono-thermal evolution of particular regions of the Turkana Depression through the application of AFT, AHe and, in the case of Chapters 2, 3 and 5, ZHe thermochronology.

Chapter 2 focusses on constraining the formational age of the Lokichar Basin in southern Turkana, the oldest known basin in the Turkana Depression [Paleogene-middle Miocene; *Morley et al.*, 1999b]. Thermochronology data from the basin margins are combined with previously published seismic and sedimentological data to determine the onset of rift-related cooling. The mechanisms which may have facilitated strain nucleation in the Turkana Depression and the implications for EARS inception are also explored.

Chapter 3 is devoted to constraining the tectono-thermal evolution of the central Turkana upper crust. These data provide new insights into the poorly constrained periods of Cretaceous and Paleogene rifting previously suggested to have modified the Turkana lithosphere.

In Chapter 4, newly acquired AHe analyses, augmented by AFT data are presented from the BRZ to constrain the spatial and temporal propagation of extensional strain in the southern Ethiopian Rift. This chapter also explores the utilisation of spatiotemporal variations in observed basin margin denudation patterns to provide insights into master fault system growth and displacement trends.

Chapter 5 explores the thermal evolution of the Lapur Range, an exposed basement block in the footwall of the northern Turkana Basin bounding master fault. The influence of crustal modification from earlier Cretaceous-early Paleogene rifting on the lithospheric response to later EARS extension is examined.

These chapters are followed by a Concluding Remarks chapter, in which the cumulative findings are succinctly synthesised to develop a regional understanding of the tectono-thermal evolution of the Turkana Depression. The resulting insights into the propagation of

strain and regional denudation/subsidence patterns are then discussed in the framework of the research questions outlined above.

References

- Abebe, T., M. L. Balestrieri, and G. Bigazzi (2010), The Central Main Ethiopian Rift is younger than 8 Ma: confirmation through apatite fission-track thermochronology, *Terra Nov.*, 22(6), 470–476, doi:10.1111/j.1365-3121.2010.00968.x.
- Anders, M. H., and R. W. Schlische (1994), Overlapping Faults, Intrabasin Highs, and the Growth of Normal Faults, *J. Geol.*, 102, 165–180.
- Balestrieri, M. L., M. Bonini, G. Corti, F. Sani, and M. Philippon (2016), A refinement of the chronology of rift-related faulting in the Broadly Rifted Zone, southern Ethiopia, through apatite fission-track analysis, *Tectonophysics*, 671, 42–55, doi:10.1016/j.tecto.2016.01.012.
- Barnett, J. A., J. Mortimer, J. H. Rippon, J. J. Walsh, and J. Watterson (1987), Displacement geometry in the volume containing a single normal fault, *Am. Assoc. Pet. Geol. Bull.*, 71(8), 925–937.
- Bauer, F. U., M. Karl, U. A. Glasmacher, B. Nagudi, A. Schumann, and L. Mroszewski (2012), Journal of African Earth Sciences The Rwenzori Mountains of western Uganda – Aspects on the evolution of their remarkable morphology within the Albertine Rift, *J. African Earth Sci.*, 73–74, 44–56, doi:10.1016/j.jafrearsci.2012.07.001.
- Bauer, F. U., U. A. Glasmacher, U. Ring, M. Karl, A. Schumann, and B. Nagudi (2013), Tracing the exhumation history of the Rwenzori Mountains, Albertine Rift, Uganda, using low-temperature thermochronology, *Tectonophysics*, 599, 8–28, doi:10.1016/j.tecto.2013.03.032.
- van der Beek, P., S. Cloetingh, and P. Andriessen (1994), Mechanisms of extensional basin formation and vertical motions at rift flanks: Constraints from tectonic modelling and fission-track thermochronology, *Earth Planet. Sci. Lett.*, 121, 417–433.
- Begg, G. C. et al. (2009), The lithospheric architecture of Africa: Seismic tomography, mantle petrology, and tectonic evolution, *Geosphere*, 5(1), 23–50, doi:10.1130/GES00179.1.
- BEICIP (1987), Geological Map of Kenya,
- Bellahsen, N., C. Faccenna, F. Funiciello, J. M. Daniel, and L. Jolivet (2003), Why did Arabia separate from Africa? Insights from 3-D laboratory experiments, *Earth Planet. Sci. Lett.*, 216, 365–381, doi:10.1016/S0012-821X(03)00516-8.
- Benoit, M. H., A. A. Nyblade, and M. E. Pasyanos (2006), Crustal thinning between the Ethiopian and East African plateaus from modeling Rayleigh wave dispersion, *Geophys. Res. Lett.*, doi:10.1029/2006GL025687.
- Bialas, R. W., W. R. Buck, and R. Qin (2010), How much magma is required to rift a continent?, *Earth Planet. Sci. Lett.*, 292, 68–78, doi:10.1016/j.epsl.2010.01.021.
- Boger, S. D., and J. M. Miller (2004), Terminal suturing of Gondwana and the onset of the Ross-Delamerian Orogeny: the cause and effect of an Early Cambrian reconfiguration of plate motions, *Earth Planet. Sci. Lett.*, 219, 35–48, doi:10.1016/S0012-821X(03)00692-7.

- Boger, S. D., W. Hirdes, C. A. M. Ferreira, T. Jenett, R. Dallwig, and C. M. Fanning (2015), The 580–520 Ma Gondwana suture of Madagascar and its continuation into Antarctica and Africa, *Gondwana Res.*, **28**, 1048–1060, doi:10.1016/j.gr.2014.08.017.
- Bonini, M., G. Corti, F. Innocenti, P. Manetti, F. Mazzarini, T. Abebe, and Z. Pecskey (2005), Evolution of the Main Ethiopian Rift in the frame of Afar and Kenya rifts propagation, *Tectonics*, **24**(1), n/a-n/a, doi:10.1029/2004TC001680.
- Bosworth, W. (1992), Mesozoic and early Tertiary rift tectonics in East Africa, *Tectonophysics*, **209**(1–4), 115–137, doi:10.1016/0040-1951(92)90014-W.
- Bosworth, W., and C. K. Morley (1994), Structural and stratigraphic evolution of the Anza rift, Kenya, *Tectonophysics*, **236**(1–4), 93–115, doi:10.1016/0040-1951(94)90171-6.
- Braun, J. (2005), Quantitative Constraints on the Rate of Landform Evolution Derived from Low-Temperature Thermochronology, *Rev. Mineral. Geochemistry*, **58**, 351–374, doi:10.2138/rmg.2005.58.13.
- Braun, J., C. Stippich, and U. A. Glasmacher (2016), The effect of variability in rock thermal conductivity on exhumation rate estimates from thermochronological data, *Tectonophysics*, **690**, 288–297, doi:10.1016/j.tecto.2016.09.027.
- Brown, R. W., R. Beucher, S. Roper, C. Persano, F. Stuart, and P. Fitzgerald (2013), Natural age dispersion arising from the analysis of broken crystals. Part I: Theoretical basis and implications for the apatite (U-Th)/He thermochronometer, *Geochim. Cosmochim. Acta*, **122**(120), 478–497, doi:10.1016/j.gca.2013.05.041.
- Brown, S. J., J. R. Thigpen, J. A. Spotila, W. C. Krugh, L. M. Tranel, and D. A. Orme (2017), Onset Timing and Slip History of the Teton Fault, Wyoming: A Multidisciplinary Reevaluation, *Tectonics*, **36**, doi:10.1002/2016TC004462.
- Browne, S. E., and J. D. Fairhead (1983), Gravity study of the Central African Rift system: A model of continental disruption 1. The Ngaoundere and Abu Gabra Rifts, *Tectonophysics*, **94**(1–4), 187–203, doi:10.1016/0040-1951(83)90016-1.
- Brune, S., G. Corti, and G. Ranalli (2017), Controls of inherited lithospheric heterogeneity on rift linkage: Numerical and analog models of interaction between the Kenyan and Ethiopian rifts across the Turkana depression, *Tectonics*, **1**–30, doi:10.1002/2017tc004739.
- Buck, W. R. (1991), Modes of Continental Lithospheric Extension, *J. Geophys. Res.*, **96**, 20,161–20,178.
- Buck, W. R. (2004), Consequences of asthenospheric variability on continental rifting, in *Rheology and deformation of the lithosphere at continental margins*, edited by G. B. Karner, N. D. Taylor, and B. Kohlstedt, pp. 92–137, Columbia University Press.
- Buck, W. R., L. L. Lavie, and A. N. B. Poliakov (1999), How to make a rift wide, *Philos. Trans. R. Soc. London A*, **357**, 671–693.
- Burke, K. (2011), Plate Tectonics, the Wilson Cycle, and Mantle Plumes: Geodynamics from the Top,

- Annu. Rev. Earth Planet. Sci.*, 39, 1–29, doi:10.1146/annurev-earth-040809-152521.
- Cartwright, J. A., B. D. Trudgill, and C. S. Mansfield (1995), Fault growth by segment linkage: an explanation for scatter in maximum displacement and trace length data from the Canyonlands Grabens of SE Utah, *J. Struct. Geol.*, 17(9), 1319–1326.
- Chang, S.-J., and S. Van der Lee (2011), Mantle plumes and associated flow beneath Arabia and East Africa, *Earth Planet. Sci. Lett.*, 302, 448–454, doi:10.1016/j.epsl.2010.12.050.
- Chorowicz, J. (2005), The East African rift system, *J. African Earth Sci.*, 43(1), 379–410, doi:10.1016/j.jafrearsci.2005.07.019.
- Civiero, C. et al. (2016), Small-scale thermal upwellings under the northern East African Rift from S travel time tomography, *J. Geophys. Res. Solid Earth*, 121, 7395–7408, doi:10.1002/2016JB013070. Received.
- Coblentz, D., and M. Sandiford (1994), Tectonic stresses in the African plate: Constraints on the ambient lithospheric stress state, *Geology*, 22(September), 831–834, doi:10.1130/0091-7613(1994)022<0831.
- Coblentz, D. D., R. M. Richardson, and M. Sandiford (1994), On the gravitational potential of the Earth's lithosphere, *Tectonics*, 13(4), 929–945.
- Coffin, M. F., and P. D. Rabinowitz (1988), Evolution of the conjugate East African-Madagascar margins and the western Somali Basin, *Geol. Soc. Am. Spec. Pap.*, 226, 1–79.
- Collins, A. S., and S. A. Pisarevsky (2005), Amalgamating eastern Gondwana: The evolution of the Circum-Indian Orogens, *Earth-Science Rev.*, 71, 229–270, doi:10.1016/j.earscirev.2005.02.004.
- Corti, G. (2009), Continental rift evolution: From rift initiation to incipient break-up in the Main Ethiopian Rift, East Africa, *Earth-Science Rev.*, 96(1–2), 1–53, doi:10.1016/j.earscirev.2009.06.005.
- Corti, G., J. van Wijk, S. Cloetingh, and C. K. Morley (2007), Tectonic inheritance and continental rift architecture: Numerical and analogue models of the East African Rift system, *Tectonics*, 26, 1–13, doi:10.1029/2006TC002086.
- Corti, G., P. Molin, A. Sembroni, I. D. Bastow, and D. Keir (2018a), Control of Pre-rift Lithospheric Structure on the Architecture and Evolution of Continental Rifts: Insights From the Main Ethiopian Rift, East Africa, *Tectonics*, 37, doi:10.1002/2017TC004799.
- Corti, G., F. Sani, S. Agostini, M. Philippon, D. Sokoutis, and E. Willingshofer (2018b), Off-axis volcano-tectonic activity during continental rifting: Insights from the transversal Goba-Bonga lineament, Main Ethiopian Rift (East Africa), *Tectonophysics*, 728–729, 75–91, doi:10.1016/j.tecto.2018.02.011.
- Courtillot, V., C. Jaupart, I. Manighetti, P. Tapponnier, and J. Besse (1999), On causal links between flood basalts and continental breakup, *Earth Planet. Sci. Lett.*, 166, 177–195, doi:10.1016/S0012-821X(98)00282-9.
- Cowie, P. A., and Z. O. E. K. Shipton (1998), Fault tip displacement gradients and process zone

- dimensions, *J. Struct. Geol.*, 20(8), 983–997.
- Curry, M. A. E., J. B. Barnes, and J. P. Colgan (2016), Testing fault growth models with low-temperature thermochronology in the northwest Basin and Range, USA, *Tectonics*, 35, 2467–2492, doi:10.1002/2016TC004211. Received.
- Davidson, A. (1983), *The Omo River Project: reconnaissance geology and geochemistry of parts of Ilubabor, Kefa, Gemu Gofa and Sidamo, Ethiopia*.
- Densmore, A. L., and R. Guidon (2005), What sets topographic relief in extensional footwalls?, *Geology*, 33(6), 453–456, doi:10.1130/G21440.1.
- Densmore, A. L., N. H. Dawers, S. Gupta, R. Guidon, and T. Goldin (2004), Footwall topographic development during continental extension, *J. Geophys. Res.*, 109, 1–16, doi:10.1029/2003JF000115.
- Densmore, A. L., R. Hetzel, S. Ivy-ochs, W. C. Krugh, N. Dawers, P. Kubik, and R. Range (2009), Spatial variations in catchment-averaged denudation rates from normal fault footwalls, *Geology*, 37(12), 1139–1142, doi:10.1130/G30164A.1.
- Dindi, E. W. (1994), Crustal structure of the Anza graben from gravity and magnetic investigations, *Tectonophysics*, 236(1–4), 359–371, doi:10.1016/0040-1951(94)90184-8.
- Dodson, M. H. (1973), Closure Temperature in Cooling Geochronological and Petrological Systems, *Contrib. to Mineral. Petrol.*, 40, 259–274.
- Ebinger, C. (2005), Continental break-up: The East African perspective, *Astron. Geophys.*, 46(2), 2–16.
- Ebinger, C., and C. A. Scholz (2012), Continental rift basins: the East African perspective, in *Tectonics and Sedimentary Basins: Recent Advances*, edited by C. Busby and A. Azor, pp. 185–208, Blackwell Publishing Ltd.
- Ebinger, C. J., and M. Casey (2001), Continental breakup in magmatic provinces: An Ethiopian example, *Geology*, 29(6), 527–530.
- Ebinger, C. J., and A. Ibrahim (1994), Multiple episodes of rifting in Central and East Africa: A re-evaluation of gravity data, *Geol. Rundschau*, 83(4), 689–702, doi:10.1007/BF00251068.
- Ebinger, C. J., and N. H. Sleep (1998), Cenozoic magnetism throughout east Africa resulting from impact of a single plume, *Nature*, 395, 788–791.
- Ebinger, C. J., T. Yemane, G. Woldegabriel, J. L. Aronson, and R. C. Walter (1993), Late Eocene–Recent volcanism and faulting in the southern main Ethiopian rift, *J. Geol. Soc. London.*, 150(1), 99–108, doi:10.1144/gsjgs.150.1.0099.
- Ebinger, C. J., T. Yemane, D. Harding, S. Tesfaye, S. Kelley, and D. C. Rex (2000), Rift deflection, migration, and propagation: Linkage of the Ethiopian and Eastern rifts, Africa, *GSA Bull.*, 112(2), 163–176.
- Ebinger, C. J., D. Keir, I. D. Bastow, K. Whaler, J. O. S. Hammond, A. Ayele, M. S. Miller, C. Tiberi, and S. Hautot (2017), Crustal Structure of Active Deformation Zones in Africa: Implications for

- Global Crustal Processes, *Tectonics*, 36, 1–35, doi:10.1002/2017TC004526.
- Ehlers, T. a. (2005), Crustal Thermal Processes and the Interpretation of Thermochronometer Data, *Rev. Mineral. Geochemistry*, 58(August), 315–350, doi:10.2138/rmg.2005.58.12.
- Ehlers, T. A., and D. S. Chapman (1999), Normal fault thermal regimes: conductive and hydrothermal heat transfer surrounding the Wasatch fault, Utah, *Tectonophysics*, 312, 217–234.
- Ehlers, T. A., and K. A. Farley (2003), Apatite (U-Th)/He thermochronometry: methods and applications to problems in tectonic and surface processes, *Earth Planet. Sci. Lett.*, 206, 1–14.
- Ehlers, T. A., P. A. Armstrong, and D. S. Chapman (2001), Normal fault thermal regimes and the interpretation of low-temperature thermochronometers, *Phys. Earth Planet. Inter.*, 126, 179–194.
- Farley, K. A., and D. F. Stockli (2002), (U-Th)/ He Dating of Phosphates: Apatite, Monazite, and Xenotime, *Rev. Mineral. Geochemistry*, 48(1), 559–577, doi:10.2138/rmg.2002.48.15.
- Fischer, T. P., P. Burnard, B. Marty, D. R. Hilton, E. Füre, F. Palhol, Z. D. Sharp, and F. Mangasini (2009), Upper-mantle volatile chemistry at Oldoinyo Lengai volcano and the origin of carbonatites, *Nat. Lett.*, 459, 77–80, doi:10.1038/nature07977.
- Fishwick, S., and I. D. Bastow (2011), Towards a better understanding of African topography: A review of passive-source seismic studies of the African crust and upper mantle, *Geol. Soc. London, Spec. Publ.*, 357, 343–371, doi:10.1144/SP357.19.
- Fitzgerald, P. G. (1992), The Transantartic Mountains of Southern Victoria Land: The Application of Apatite Fission Track Analysis to a Rift Shoulder Uplift, *Tectonics*, 11(3), 634–662.
- Flowers, R. M., R. A. Ketcham, D. L. Shuster, and K. A. Farley (2009), Apatite (U-Th)/He thermochronometry using a radiation damage accumulation and annealing model, *Geochim. Cosmochim. Acta*, 73(8), 2347–2365, doi:10.1016/j.gca.2009.01.015.
- Forte, A. M., S. Quéré, R. Moucha, N. A. Simmons, S. P. Grand, J. X. Mitrovica, and D. B. Rowley (2010), Joint seismic–geodynamic–mineral physical modelling of African geodynamics: A reconciliation of deep-mantle convection with surface geophysical constraints, *Earth Planet. Sci. Lett.*, 295, 329–341, doi:10.1016/j.epsl.2010.03.017.
- Foster, D. A., and A. J. W. Gleadow (1992), The morphotectonic evolution of rift-margin mountains in central Kenya: Constraints from apatite fission-track thermochronology, *Earth Planet. Sci. Lett.*, 113(1–2), 157–171, doi:10.1016/0012-821X(92)90217-J.
- Foster, D. A., and A. J. W. Gleadow (1993), Episodic Denudation in East Africa: A Legacy of Intracontinental Tectonism, *Geophys. Res. Lett.*, 20(21), 2395–2398.
- Foster, D. A., and J. W. Gleadow (1996), Structural framework and denudation history of the flanks of the Kenya and Anza Rifts , East Africa, *Tectonics*, 15(2), 258–271, doi:10.1029/95TC02744.
- Furman, T. (2007), Geochemistry of East African Rift basalts: An overview, *J. African Earth Sci.*, 48, 147–160, doi:10.1016/j.jafrearsci.2006.06.009.
- Furman, T., J. Bryce, T. Rooney, B. Hanan, G. Yirgu, and D. Ayalew (2006), Heads and tails: 30

- million years of the Afar Plume, *Geol. Soc. London, Spec. Publ.*, 259, 95–119.
- Gallagher, K. (1995), Evolving temperature histories from apatite fission-track data, *Earth Planet. Sci. Lett.*, 136(3–4), 421–435.
- Gallagher, K. (2012), Transdimensional inverse thermal history modeling for quantitative thermochronology, *J. Geophys. Res. Solid Earth*, 117, 1–16, doi:10.1029/2011JB008825.
- Gallagher, K., K. Charvin, S. Nielsen, M. Sambridge, and J. Stephenson (2009), Markov chain Monte Carlo (MCMC) sampling methods to determine optimal models, model resolution and model choice for Earth Science problems, *Mar. Pet. Geol.*, 26, 525–535, doi:10.1016/j.marpetgeo.2009.01.003.
- Gawthorpe, R. L., and M. R. Leeder (2000), Tectono-sedimentary evolution of active extensional basins, *Basin Res.*, 12, 195–218.
- George, R., N. Rogers, and S. Kelley (1998), Earliest magnetism in Ethiopia: Evidence for two mantle plumes in one flood basalt province, *Geology*, 26(10), 923–926.
- George, R. M., and N. W. Rogers (2002), Plume dynamics beneath the African plate inferred from the geochemistry of the Tertiary basalts of southern Ethiopia, *Contrib. to Mineral. Petrol.*, 144, 286–304, doi:10.1007/s00410-002-0396-z.
- Ghosh, A., and W. E. Holt (2012), Plate Motions and Stresses from Global Dynamic Models, *Science* (80-.), 335, 838–843, doi:10.1126/science.1214209.
- Gleadow, A. J. W. (1980), Fission track age of the KBS Tuff and associated hominid remains in northern Kenya, *Nature*, 284, 225–230, doi:10.1038/284225a0.
- Gleadow, A. J. W., and I. R. Duddy (1981), A Natural Long-Term Track Annealing Experiment for Apatite, *Nucl. Tracks*, 5(1–2), 169–174.
- GMRD-Sudan (1981), Geological Map of the Sudan,
- Grand, S. P., R. D. van der Hilst, and S. Widiyantoro (1997), Global Seismic Tomography: A Snapshot of Convection in the Earth, *GSA Today*, 7(4), 1–7.
- Green, P. F., I. R. Duddy, A. J. W. Gleadow, and J. F. Lovering (1989), Apatite Fission-Track Analysis as a Paleotemperature Indicator for Hydrocarbon Exploration, in *Thermal History of Sedimentary Basins, Methods and Case Histories*, edited by N. D. Naeser and T. H. McCulloh, pp. 181–196, Springer-Verlag, New York.
- Guenther, W. R., P. W. Reiners, R. A. Ketcham, L. Nasdala, and G. Giester (2013), Helium diffusion in natural zircon: radiation damage, anisotropy, and the interpretation of zircon (U-TH)/He thermochronology, *Am. J. Sci.*, 313(3), 145–198, doi:10.2475/03.2013.01.
- Guiraud, R., and W. Bosworth (1997), Senonian basin inversion and rejuvenation of rifting in Africa and Arabia : synthesis and implications to plate-scale tectonics, *Tectonophysics*, 282, 39–82.
- Gupta, S., P. A. Cowie, N. H. Dawers, and J. R. Underhill (1998), A mechanism to explain rift-basin subsidence and stratigraphic patterns through fault-array evolution, *Geology*, 26(7), 595–598, doi:10.1130/0091-7613(1998)026<0595.

- Hansen, S. E., A. A. Nyblade, and M. H. Benoit (2012), Mantle structure beneath Africa and Arabia from adaptively parameterized P-wave tomography: Implications for the origin of Cenozoic Afro-Arabian tectonism, *Earth Planet. Sci. Lett.*, 319–320, 23–34, doi:10.1016/j.epsl.2011.12.023.
- Hendrie, D. B., N. J. Kusznir, C. K. Morley, and C. J. Ebinger (1994), Cenozoic extension in northern Kenya: a quantitative model of rift basin development in the Turkana region, *Tectonophysics*, 236(1–4), 409–438, doi:10.1016/0040-1951(94)90187-2.
- Hopper, J. R., and W. R. Buck (1993), The Initiation of Rifting at Constant Tectonic Force: Role of Diffusion Creep, *J. Geophys. Res.*, 98, 16,213–16,221.
- Hughes, G. W., O. Varol, and Z. R. Beydoun (1991), Evidence for Middle Oligocene rifting of the Gulf of Aden and for Late Oligocene rifting of the southern Red Sea, *Mar. Pet. Geol.*, 8, 354–358.
- Ibrahim, A. E., C. J. Ebinger, and J. D. Fairhead (1991), Interpretation of the Central African Rift system in Sudan based on new gravity and aeromagnetic data, *Eos (Washington. DC)*, 462.
- Johnson, J. E., R. M. Flowers, G. B. Baird, and K. H. Mahan (2017), “Inverted” zircon and apatite (U–Th)/He dates from the Front Range, Colorado: High-damage zircon as a low-temperature (<50 °C) thermochronometer, *Earth Planet. Sci. Lett.*, 466, 80–90, doi:10.1016/j.epsl.2017.03.002.
- Kazmin, V. (1972), Geological Map of Ethiopia,
- Kendall, J.-M., and C. Lithgow-Bertelloni (2016), Why is Africa rifting?, *Geol. Soc. London, Spec. Publ.*, 420, 11–30.
- Ketcham, R. A., A. Carter, R. A. Donelick, J. Barbarand, and A. J. Hurford (2007), Improved modeling of fission-track annealing in apatite, *Am. Mineral.*, 92, 799–810, doi:10.2138/am.2007.2281.
- Kim, Y., and D. J. Sanderson (2005), The relationship between displacement and length of faults: a review, *Earth-Science Rev.*, 68, 317–334, doi:10.1016/j.earscirev.2004.06.003.
- Kohn, B. P., A. J. W. Gleadow, R. W. Brown, K. Gallagher, M. Lorencak, and W. P. Noble (2005), Visualizing Thermotectonic and Denudational Histories Using Apatite Fission Track Thermochronology, *Rev. Mineral. Geochemistry*, 58, 527–565, doi:10.2138/rmg.2005.58.20.
- Koptev, A., S. Cloetingh, T. Gerya, E. Calais, and S. Leroy (2017), Non-uniform splitting of a single mantle plume by double cratonic roots: Insight into the origin of the central and southern East African Rift System, *Terra Nov.*, 30, 125–134, doi:10.1111/ter.12317.
- Kreuser, T. (1995), Rift to drift evolution in Permian-Jurassic basins of East Africa, *Geol. Soc. London, Spec. Publ.*, 80(1), 297–315.
- Kröner, a, and R. J. Stern (2005), Pan-African Orogeny, *Encycl. Of Geol.*, 1(2004), 1–12, doi:10.1016/S1342-937X(05)70162-3.
- Krugh, W. (2008), Low-temperature thermochronologic constraints on fault array evolution and

- patterns of range-scale denudation, ETH Zurich.
- Kusznir, N. J., and R. G. Park (1987), The extensional strength of the continental lithosphere: its dependence on geothermal gradient, and crustal composition and thickness, *Geol. Soc. Spec. Publ.*, (28), 35–52.
- Lavier, L. L., and G. Manatschal (2006), LETTERS magma-poor margins, , 440(March), doi:10.1038/nature04608.
- Lehto, T. et al. (2014), Geological Map of Uganda, Scale 1:1 000 000,
- Leroy, S. et al. (2004), From rifting to spreading in the eastern Gulf of Aden: a geophysical survey of a young oceanic basin from margin to margin, *Terra Nov.*, 16, 185–192, doi:10.1111/j.1365-3121.2004.00550.x.
- Li, C., R. D. van der Hilst, E. R. Engdahl, and S. Burdick (2008), A new global model for P wave speed variations in Earth's mantle, *Geochemistry Geophys. Geosystems*, 9(5), 1–21, doi:10.1029/2007GC001806.
- Lister, G. S., M. A. Etheridge, and P. A. Symonds (1986), Detachment faulting and the evolution of passive continental margins, *Geology*, 14, 246–250, doi:10.1130/0091-7613(1986)14<891.
- Lister, G. S., M. A. Etheridge, and P. A. Symonds (1991), Detachment Models for the Formation of Passive Continental Margins, *Tectonics*, 10(5), 1038–1064.
- Macgregor, D. (2015), History of the development of the East African Rift System: A series of interpreted maps through time, *J. African Earth Sci.*, 101, 232–252, doi:10.1016/j.jafrearsci.2014.09.016.
- Mackintosh, V., B. Kohn, A. Gleadow, and Y. Tian (2017), Phanerozoic Morphotectonic Evolution of the Zimbabwe Craton: Unexpected Outcomes From a Multiple Low-Temperature Thermochronology Study, *Tectonics*, 36, 1–24, doi:10.1002/2017TC004703.
- Mancktelow, N. S., and B. Grasemann (1997), Time-dependent effects of heat advection and topography on cooling histories during erosion, *Tectonophysics*, 270, 167–195.
- Mariita, N. O., and G. R. Keller (2007), An integrated geophysical study of the northern Kenya rift, *J. African Earth Sci.*, 48(2–3), 80–94, doi:10.1016/j.jafrearsci.2006.05.008.
- McDougall, I., and F. H. Brown (2009), Timing of volcanism and evolution of the northern Kenya Rift, *Geol. Mag.*, 146(1), 34–37, doi:10.1017/S0016756808005347.
- Mckenzie, D. A. N. (1978), Some remarks on the development of sedimentary basins, , 40, 25–32.
- Mechie, J., G. R. Keller, C. Prodehl, M. a. Khan, and S. J. Gaciri (1997), A model for the structure, composition and evolution of the Kenya rift, *Tectonophysics*, 278, 95–119, doi:10.1016/S0040-1951(97)00097-8.
- Morley, C. K. (1994), Interaction of deep and shallow processes in the evolution of the Kenya rift, *Tectonophysics*, 236, 81–91.
- Morley, C. K. (1999), Basin Evolution Trends in East Africa, in *Geoscience of Rift Systems-Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 131–150.

- Morley, C. K., W. a. Wescott, D. M. Stone, R. M. Harper, S. T. Wigger, and F. M. Karanja (1992), Tectonic evolution of the northern Kenyan Rift, *J. Geol. Soc. London.*, 149(3), 333–348, doi:10.1144/gsjgs.149.3.0333.
- Morley, C. K., W. Bosworth, R. A. Day, R. Lauck, R. Boshier, D. M. Stone, S. T. Wigger, W. A. Wescott, D. Haun, and N. Bassett (1999a), Geology and Geophysics of the Anza Graben, in *Geoscience of Rift Systems-Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 67–90.
- Morley, C. K., D. M. Stone, R. M. Harper, and S. T. Wigger (1999b), Geology and Geophysics of the Western Turkana Basins, Kenya, in *Geoscience of Rift Systems—Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 19–54.
- Morley, C. K., D. K. Ngenoh, and J. K. Ego (1999c), Introduction to the East African Rift System, in *Geoscience of Rift Systems-Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 1–18.
- Mortimer, E., L. A. Kirstein, F. M. Stuart, and M. R. Strecker (2016), Spatio-temporal trends in normal-fault segmentation recorded by low-temperature thermochronology: Livingstone fault scarp, Malawi Rift, East African Rift System, *Earth Planet. Sci. Lett.*, 455, 62–72, doi:10.1016/j.epsl.2016.08.040.
- Moucha, R., and A. M. Forte (2011), Changes in African topography driven by mantle convection, *Nat. Geosci.*, 4(10), 707–712, doi:10.1038/ngeo1235.
- Mugisha, F., C. J. Ebinger, M. Strecker, and D. Pope (1997), Two-stage rifting in the Kenya rift: implications for half-graben models, *Tectonophysics*, 278, 63–81.
- NASA (2000), Shuttle Radar Topography Mission (SRTM) 1 Arc-Second Global,
- Noble, W. P., D. A. Foster, and A. J. W. Gleadow (1997), The post-Pan-African thermal and extensional history of crystalline basement rocks in eastern Tanzania, *Tectonophysics*, 275(4), 331–350, doi:10.1016/S0040-1951(97)00026-7.
- Nyblade, A. A. (2011), The upper-mantle low-velocity anomaly beneath Ethiopia, Kenya, and Tanzania: Constraints on the origin of the African superswell in eastern Africa and plate versus plume models of mantle dynamics, *Geol. Soc. Am. Spec. Pap.*, 478, 1–14, doi:10.1130/2011.2478(03).
- Nyblade, A. A., and S. W. Robinson (1994), The African Superswell, *Geophys. Res. Lett.*, 21(9), 765–768.
- Peacock, D., and D. J. Sanderson (1994), Geometry and Development of Relay Ramps in Normal Fault Systems, *Am. Assoc. Pet. Geol. Bull.*, 78(2), 147–165, doi:10.1306/BDF9046-1718-11D7-8645000102C1865D.
- Pérez-Gussinyé, M., M. Metois, M. Fernández, J. Vergés, J. Fulla, and A. R. Lowry (2009), Effective elastic thickness of Africa and its relationship to other proxies for lithospheric structure and surface tectonics, *Earth Planet. Sci. Lett.*, 287, 152–167,

doi:10.1016/j.epsl.2009.08.004.

- Philippon, M., G. Corti, F. Sani, M. Bonini, M. Balestrieri, P. Molin, E. Willingshofer, D. Sokoutis, and S. Cloetingh (2014), Evolution, distribution, and characteristics of rifting in southern Ethiopia, *Tectonics*, *33*, 1–24, doi:10.1002/2013TC003430.
- Pik, R. (2011), Geodynamics: East Africa on the rise, *Nat. Geosci.*, *4*(10), 660–661, doi:10.1038/ngeo1274.
- Pik, R., B. Marty, J. Carignan, and J. Lavé (2003), Stability of the Upper Nile drainage network (Ethiopia) deduced from (U-Th)/He thermochronometry: implications for uplift and erosion of the Afar plume dome, *Earth P*, *215*, 73–88, doi:10.1016/S0012-821X(03)00457-6.
- Pik, R., B. Marty, and D. R. Hilton (2006), How many mantle plumes in Africa? The geochemical point of view, *Chem. Geol.*, *226*, 100–114, doi:10.1016/j.chemgeo.2005.09.016.
- Pik, R., B. Marty, J. Carignan, G. Yirgu, and T. Ayalew (2008), Timing of East African Rift development in southern Ethiopia : Implication for mantle plume activity and evolution of topography, *Geology*, *36*(2), 167–170, doi:10.1130/G24233A.1.
- Purcell, P. G. (2017), Re-imagining and re-imaging the development of the East African Rift, *Pet. Geosci.*, *36*, doi:https://doi.org/10.1144/petgeo2-17-036.
- Reeves, C. V., J. P. Teasdale, and E. S. Mahanjane (2016), Insight into the Eastern Margin of Africa from a new tectonic model of the Indian Ocean, *Geol. Soc. London Spec. Publ.*, *431*, doi:10.1144/SP431.12.
- Reiners, P. W., T. A. Ehlers, and P. K. Zeitler (2005), Past, present, and future of thermochronology, *Revi*, *58*, 1–18, doi:10.2138/rmg.2005.58.1.
- Ritsema, J., A. A. Nyblade, T. J. Owens, C. A. Langston, and J. C. VanDecar (1998), Upper mantle seismic velocity structure beneath Tanzania, east Africa: Implications for the stability of cratonic lithosphere, *J. Geophys. Res.*, *103*(B9), 21,201–21,213.
- Ritsema, J., H.-J. van Heijst, and J. H. Wodhouse (1999), Complex Shear Wave Velocity Structure Imaged Beneath Africa and Iceland, *Science (80-.)*, *286*(5446), 1925–1928.
- Roberts, E. M., P. M. O. Connor, N. J. Stevens, M. D. Gottfried, Z. A. Jinnah, S. Ngasala, A. M. Choh, and R. A. Armstrong (2010), Sedimentology and depositional environments of the Red Sandstone Group, Rukwa Rift Basin, southwestern Tanzania: New insight into Cretaceous and Paleogene terrestrial ecosystems and tectonics in sub-equatorial Africa, *J. African Earth Sci.*, *57*, 179–212, doi:10.1016/j.jafrearsci.2009.09.002.
- Roberts, E. M., N. J. Stevens, P. M. O'Connor, P. H. G. M. Dirks, M. D. Gottfried, W. C. Clyde, R. A. Armstrong, a. I. S. Kemp, and S. Hemming (2012), Initiation of the western branch of the East African Rift coeval with the eastern branch, *Nat. Geosci.*, *5*(4), 289–294, doi:10.1038/ngeo1432.
- Rogers, N. W. (2006), Basaltic magmatism and the geodynamics of the East African Rift System, *Geol. Soc. London, Spec. Publ.*, *259*, 77–93, doi:10.1144/GSL.SP.2006.259.01.08.

- Rooney, T. O. (2017), The Cenozoic magmatism of East-Africa: Part I — Flood basalts and pulsed magmatism, *Lithos*, 286–287, 264–301, doi:10.1016/j.lithos.2017.05.014.
- Rooney, T. O., C. Herzberg, and I. D. Bastow (2012), Elevated mantle temperature beneath East Africa, *Geology*, 40(1), 27–30, doi:10.1130/G32382.1.
- Schmeling, H., and H. Wallner (2012), Magmatic lithospheric heating and weakening during continental rifting: A simple scaling law, a 2-D thermomechanical rifting model and the East African Rift System, *Geochemistry, Geophys. Geosystems*, 13(8), doi:10.1029/2012GC004178.
- Schull, T. J. (1988), Rift Basins of Interior Sudan: Petroleum Exploration and Discovery, , 72(10), 1128–1142.
- Seiler, C., A. J. W. Gleadow, J. M. Fletcher, and B. P. Kohn (2009), Thermal evolution of a sheared continental margin: Insights from the Ballenas transform in Baja California, Mexico, *Earth Planet. Sci. Lett.*, 285(1–2), 61–74, doi:10.1016/j.epsl.2009.05.043.
- Seiler, C., J. M. Fletcher, B. P. Kohn, A. J. W. Gleadow, and A. Raza (2011), Low - temperature thermochronology of northern Baja California , Mexico : Decoupled slip - exhumation gradients and delayed onset of oblique rifting across the Gulf of California, *Tectonics*, 30(1), doi:10.1029/2009TC002649.
- Sengör, A. M. C., and K. Burke (1978), Relative Timing of Rifting and Volcanism on Earth and its Tectonic Implications, *Geophys. Res. Lett.*, 5(6), 419–421.
- Simmons, N. A., A. M. Forte, and S. P. Grand (2007), Thermochemical structure and dynamics of the African superplume, *Geophys. Res. Lett.*, 34, 1–5, doi:10.1029/2006GL028009.
- Sippel, J., C. Meeßen, M. Cacace, J. Mechie, S. Fishwick, C. Heine, M. Scheck-Wenderoth, and M. R. Strecker (2017), The Kenya rift revisited: insights into lithospheric strength through data-driven 3-D gravity and thermal modelling, *Solid Earth*, 8, 45–81, doi:10.5194/se-8-45-2017.
- Sleep, N. H., C. J. Ebinger, and J.-M. Kendall (2002), Deflection of mantle plume material by cratonic keels, *Geol. Soc. London, Spec. Publ.*, 1999, 135–150.
- Sokoutis, D., G. Corti, M. Bonini, J. Pierre, S. Cloetingh, T. Mauduit, and P. Manetti (2007), Modelling the extension of heterogeneous hot lithosphere, *Tectonophysics*, 444, 63–79, doi:10.1016/j.tecto.2007.08.012.
- Spiegel, C., B. P. Kohn, D. X. Belton, and A. J. W. Gleadow (2007), Morphotectonic evolution of the central Kenya rift flanks: Implications for late Cenozoic environmental change in East Africa, *Geology*, 35(5), 427–430, doi:10.1130/G23108A.1.
- Stamps, D. S., L. M. Flesch, and E. Calais (2010), Lithospheric buoyancy forces in Africa from a thin sheet approach, *Int. J. Earth Sci.*, 99, 1525–1533, doi:10.1007/s00531-010-0533-2.
- Stamps, D. S., L. M. Flesch, E. Calais, and A. Ghosh (2014), Current kinematics and dynamics of Africa and the East African Rift System, *J. Geophys. Res. Solid Earth*, 119, 5161–5186, doi:10.1002/2013JB010717.Received.
- Stamps, D. S., G. Iaffaldano, and E. Calais (2015), Role of mantle flow in Nubia-Somalia plate

- divergence, *Geophys. Res. Lett.*, *42*, 290–296, doi:10.1002/2014GL062515. Received.
- Stockli, D. F. (2005), Application of Low-Temperature Thermochronometry to Extensional Tectonic Settings, *Rev. Mineral. Geochemistry*, *58*, 411–448, doi:10.2138/rmg.2005.58.16.
- Stüwe, K., L. White, and R. Brown (1994), The influence of eroding topography on steady-state isotherms. Application to fission track analysis, *Earth Planet. Sci. Lett.*, *124*, 63–74.
- Tiercelin, J.-J., J.-L. Potdevin, C. K. Morley, M. R. Talbot, H. Bellon, A. Rio, B. Le Gall, and W. Vétel (2004), Hydrocarbon potential of the Meso-Cenozoic Turkana Depression, northern Kenya. I. Reservoirs: depositional environments, diagenetic characteristics, and source rock–reservoir relationships, *Mar. Pet. Geol.*, *21*, 41–62, doi:10.1016/j.marpetgeo.2003.11.007.
- Torres Acosta, V., A. Bande, E. R. Sobel, M. Parra, T. F. Schildgen, F. Stuart, and M. R. Strecker (2015), Cenozoic extension in the Kenya Rift from low-temperature thermochronology: Links to diachronous spatiotemporal evolution of rifting in East Africa, *Tectonics*, *34*(12), 2367–2386, doi:10.1002/2015TC003949.
- Trudgill, B., and J. Cartwright (1994), Relay-Ramp forms and normal-fault linkages, Canyonlands National Park, Utah, *Geol. Soc. Am. Bull.*, *106*, 1143–1157.
- Unrug, R. (1997), Rodinia to Gondwana: The geodynamic map of Gondwana supercontinent assembly, *GSA Today*, *7*(1), doi:10.2460/javma.242.12.1666.
- Vermeesch, P., and Y. Tian (2014), Thermal history modelling: HeFTy vs. QTQt, *Earth-Science Rev.*, *139*, 279–290, doi:10.1016/j.earscirev.2014.09.010.
- Vétel, W., and B. Le Gall (2006), Dynamics of prolonged continental extension in magmatic rifts: the Turkana Rift case study (North Kenya), *Geol. Soc. London Spec. Publ.*, *259*, 209–233.
- Vétel, W., B. Le Gall, and J. J. Walsh (2005), Geometry and growth of an inner rift fault pattern: the Kino Sogo Fault Belt, Turkana Rift (North Kenya), *J. Struct. Geol.*, *27*, 2204–2222, doi:10.1016/j.jsg.2005.07.003.
- Wagner, M., R. Altherr, and P. Van Den Haute (1992), Apatite fission-track analysis of Kenyan basement rocks: constraints on the thermotectonic evolution of the Kenya dome. A reconnaissance study, *Tectonophysics*, *204*(1–2), 93–110, doi:10.1016/0040-1951(92)90272-8.
- Weeraratne, D. S., D. W. Forsyth, K. M. Fischer, and A. A. Nyblade (2003), Evidence for an upper mantle plume beneath the Tanzanian craton from Rayleigh wave tomography, *J. Geophys. Res.*, *108*(B9), 1–14, doi:10.1029/2002JB002273.
- Wernicke, B. (1985), Uniform-sense normal simple shear of the continental lithosphere, *Can. J. Earth Sci.*, *22*, 108–125.
- Wescott, W. A., S. T. Wigger, D. M. Stone, and C. K. Morley (1999), Geology and Geophysics of the Lotikipi Plain, *Geosci. Rift Syst. East Africa*, (44), 55–66.
- White, R., and D. McKenzie (1989), Magmatism at Rift Zones: The Generation of Volcanic Continental Margins and Flood Basalts, *J. Geophys. Res.*, *94*(B6), 7685–7729.
- White, R. S., and D. McKenzie (1995), Mantle plumes and flood basalts, *J. Geophys. Res.*, *100*(B9),

17,543-17,585.

- Wichura, H., R. Bousquet, R. Oberhänsli, M. R. Strecker, and M. H. Trauth (2011), The Mid-Miocene East African Plateau: a pre-rift topographic model inferred from the emplacement of the phonolitic Yatta lava flow, Kenya, *Geol. Soc. London, Spec. Publ.*, 357(1), 285–300, doi:10.1144/sp357.15.
- Wijk, J. W. van, R. S. Huismans, M. ter Voorde, and S. A. P. L. Cloetingh (2001), Melt Generation at Volcanic Continental Margins: no Need for Mantle Plume?, *Geophys. Res. Lett.*, 28(20), 3995–3998.
- Wildman, M., R. Brown, R. Beucher, C. Persano, F. Stuart, K. Gallagher, J. Schwanethal, and A. Carter (2016), The chronology and tectonic style of landscape evolution along the elevated Atlantic continental margin of South Africa resolved by joint apatite fission track and (U-Th-Sm)/He thermochronology, *Tectonics*, 35(3), 511–545, doi:10.1002/2015TC004042.
- Wilson, J. T. (1965), A New Class of Faults and Their Bearing on Continental Drift, *Nature*, 207(4995), 343–347.
- Wilson, J. T. (1968), Static or Mobile Earth: The Current Scientific Revolution, *Proc. Am. Philos. Soc.*, 112(5), 309–320.
- Zanettin, B., E. Justin Visentin, G. Bellieni, E. M. Piccirillo, and R. Francesca (1983), The volcanism of the North Turkana Basin, Kenya: age, succession and structural evolution, *Bull. des Centres Rech. Explor. Elf-Aquitaine*, 7, 249–255.
- Ziegler, P. A., and S. Cloetingh (2004), Dynamic processes controlling evolution of rifted basins, *Earth-Science Rev.*, 64, 1–50, doi:10.1016/S0012-8252(03)00041-2.

Birth of the East African Rift System: Eocene nucleation of volcanism and faulting in the Turkana Depression

Abstract

The Turkana Depression of northern Kenya and southern Ethiopia contains voluminous plume-related basalt flows that mark the onset of the Paleogene-Recent East African Rift System (EARS) approximately 45 Ma. Thus, the Turkana Depression is crucial to our understanding of the inception of intracontinental rifting. However, the precise chronology of early rift basin formation in Turkana is poorly constrained. We present low-temperature thermochronology data from basement rocks from the margins of the N-S trending Lokichar Basin. Apatite fission track, apatite (U-Th-Sm)/He and zircon (U-Th)/He analyses constrain the onset of rift-related cooling in Turkana. Thermal history modelling of these data document a period of pronounced Eocene to Miocene denudational cooling of the basin-bounding Lokichar Fault footwall, while data from the Lokichar Basin flexural margin record a concurrent period of subsidence and reheating. These results, along with ~7 km of Paleogene to middle Miocene syn-rift strata preserved in the Lokichar Fault hanging wall, are consistent with an onset of Lokichar Basin formation between ~45-40 Ma. Pre-existing rheological heterogeneities inherited from earlier Mesozoic rifting and plume magmatism likely facilitated the Eocene localisation of strain in the Turkana Depression, significantly earlier than EARS initiation elsewhere. The broadly concurrent nucleation of plume-volcanism and faulting provides valuable insights into the forces driving magma-assisted continental rift initiation.

2.1 Introduction

Intracontinental rift systems are key features in continental breakup, yet the mechanisms behind their initiation and propagation lack consensus. The Paleogene-Recent East African Rift System (EARS) is the pre-eminent modern example of an intracontinental rift zone and remains the focus of considerable debate regarding early-stage continental breakup processes [e.g. *Buck, 2004; Brune et al., 2017*]. The EARS is considered to have developed diachronously [e.g. *Vétel and Le Gall, 2006*] above an elongate, deep-seated asthenospheric thermal anomaly impinged beneath East Africa, known as the African Superplume [e.g. *Ebinger and Sleep, 1998; Moucha and Forte, 2011*]. Dynamic topography, magmatic underplating and volcanic buildup associated with the African Superplume resulted in the formation of contiguous high (~1 km) plateaux that span much of eastern and southern Africa [*Benoit et al., 2006*]. Buoyancy forces associated with the resulting lateral gradients in gravitational potential energy are largely responsible for driving modern extensional deformation in East Africa [e.g. *Stamps et al., 2015*]. However, these forces are widely considered to be insufficient to rupture cold, thick lithosphere [*Buck, 2004*]. Thus, the forces driving rifting during initiation of the EARS remain enigmatic. Moreover, controversy exists as to the relative significance of pre-existing rheological heterogeneities and plume magmatism in the initial focusing of extensional strain [e.g. *Vétel and Le Gall, 2006; Ebinger and Sleep, 1998*]. Resolving the timing and distribution of extensional strain during the initial phase of the EARS is critical to addressing these questions.

The Turkana Depression is a structurally complex and long-lived segment of the EARS (Figure 2.1), making it an ideal region to study the early stages of rifting. Flood basalts emplaced there ca. 45 Ma are considered to represent the earliest manifestation of EARS-related volcanism [*Morley, 1994; Ebinger and Sleep, 1998*]. Seismic reflection data reveal the presence of probable Paleogene-age rift basins in the Turkana Depression [~66-23 Ma; *Morley et al., 1999*], suggesting it also hosts the earliest manifestation of extensional faulting in the EARS. However, the lack of surface exposure of the earliest syn-rift strata, cover by Cenozoic volcanism and the dearth of geological data from South Sudan have precluded the determination of the onset of extensional strain in the Turkana Depression.

I present apatite fission track (AFT), apatite (U-Th-Sm)/He (AHe) and zircon (U-Th)/He (ZHe) data from basement rocks from the Lokichar Basin margins to constrain the onset of

rift development in southern Turkana. Thermal history modelling of these data provides new insights into the timing of rift basin formation and subsequent strain migration during early EARS development.

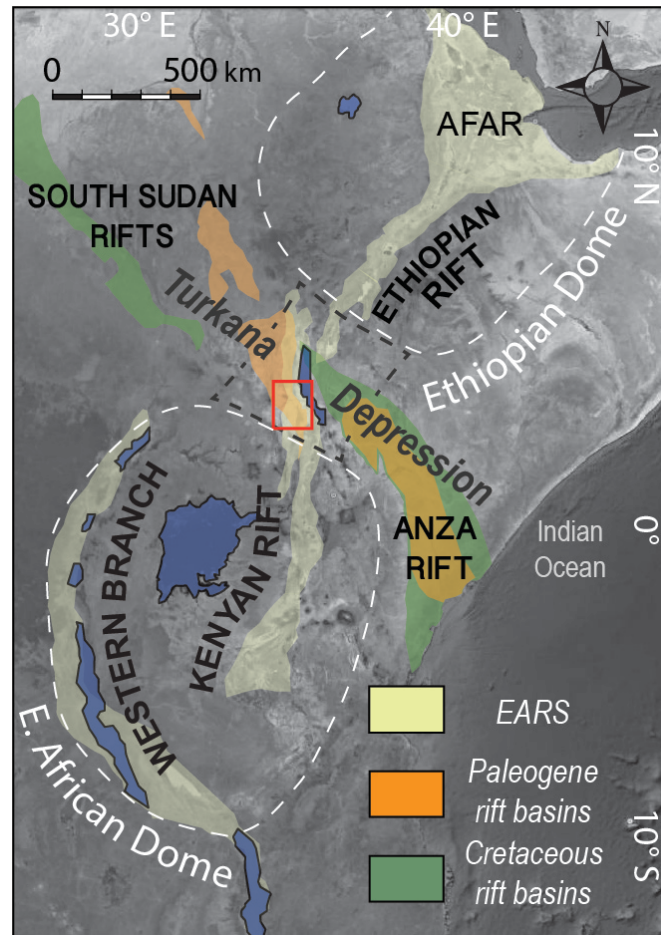


Figure 2.1. Topography of East Africa, showing the Turkana Depression (grey dashed), the Ethiopian and East African Domes (white dashed), Cretaceous-Paleogene Anza and South Sudan rifts and the EARS, and location of Figure 2.2A (red box).

2.2 Geological Setting

The EARS comprises a series of fault-bounded extensional basins and uplifted shoulders that stretch from the Red Sea to southern Africa. In general, the EARS is defined by a narrow (~80 km) zone of extensional deformation that traverses two wide regions of elevated topography, the Ethiopian (~1500 m) and East African Domes (~1200 m) [Figure 2.1; *Purcell, 2017*]. These plateaux are separated by the Turkana Depression, a ~300 km wide zone of diffuse deformation and subdued topography [*Morley et al., 1999*]. Geophysical data reveal that the Turkana crust has been significantly attenuated (~20 km thick), resulting in the region being unable to support long-wavelength topography [*Benoit et al., 2006*].

Some authors attribute the anomalous crustal architecture and morphology of the Turkana Rift to pre-existing rheological heterogeneities inherited from earlier Cretaceous-early Paleogene extension in the NW-SE trending Anza and South Sudan rifts (Figure 2.1) [e.g. *Brune et al., 2017*]. However, evidence for this earlier phase of rifting in Turkana is generally lacking [*Vétel and Le Gall, 2006*], suggesting later Paleogene tectonism/plume activity played a significant role in modifying the Turkana crust (Chapter 3).

The oldest known EARS basin within the Turkana Depression is the Paleogene-middle Miocene Lokichar Basin (Figures 2.2A and B). The ~N-S trending half-graben thickens westwards towards the basin-bounding, east-dipping Lokichar Fault, which generally lacks a present-day topographic expression. The eastern Lokichar Basin margin is defined by a flexural margin where syn-rift sediments onlap an exposed block of Precambrian basement called the Lokone Horst [*Morley et al., 1999*]. This horst defines the western margin of the west-facing, middle Mio-Pliocene North Kerio Basin half-graben [*Vétel and Le Gall, 2006*].

The Lokichar Basin has undergone ~12 km of extension, focused on the Lokichar Fault [*Morley et al., 1999*]. Over most of its length (>150 km), the Lokichar Fault strikes ~N-S, rotating to a N140°E trend towards its southern end. Alignment of the N140°E segment with a prominent ~NW-SE trending morphological scarp >100 km further east led *Vétel and Le Gall* [2006] to suggest that the southern Lokichar Fault reactivated an older Cretaceous transverse structure called the Karisia Fault [Figure 2.2A]. The Lokichar Fault continues northwards without any apparent break or en-echelon offset, to bound the westward thickening middle Miocene-Pliocene North Lokichar Basin.

The Lokichar Basin hosts a ~ 7km thick succession of lacustrine and fluviolacustrine deposits [Morley *et al.*, 1999]. The shallower syn-rift sequences have been studied in extensive surface exposures along the Lokichar flexural margin, in the Loperot-1 well (2950 m total depth) and fourteen shallow (90 m) wells [Morley *et al.*, 1999]. The sedimentary package comprises arkosic fluvio-deltaic sandstones and lacustrine shales, with a palynomorph assemblage dated to the Oligocene-lower Miocene in age [Morley *et al.*, 1999]. However, due to onlap onto basement near the flexural margin, the deeper reflection packages are not exposed, leading Morley *et al.* [1999] to suggest a possible Eocene age for the basin.

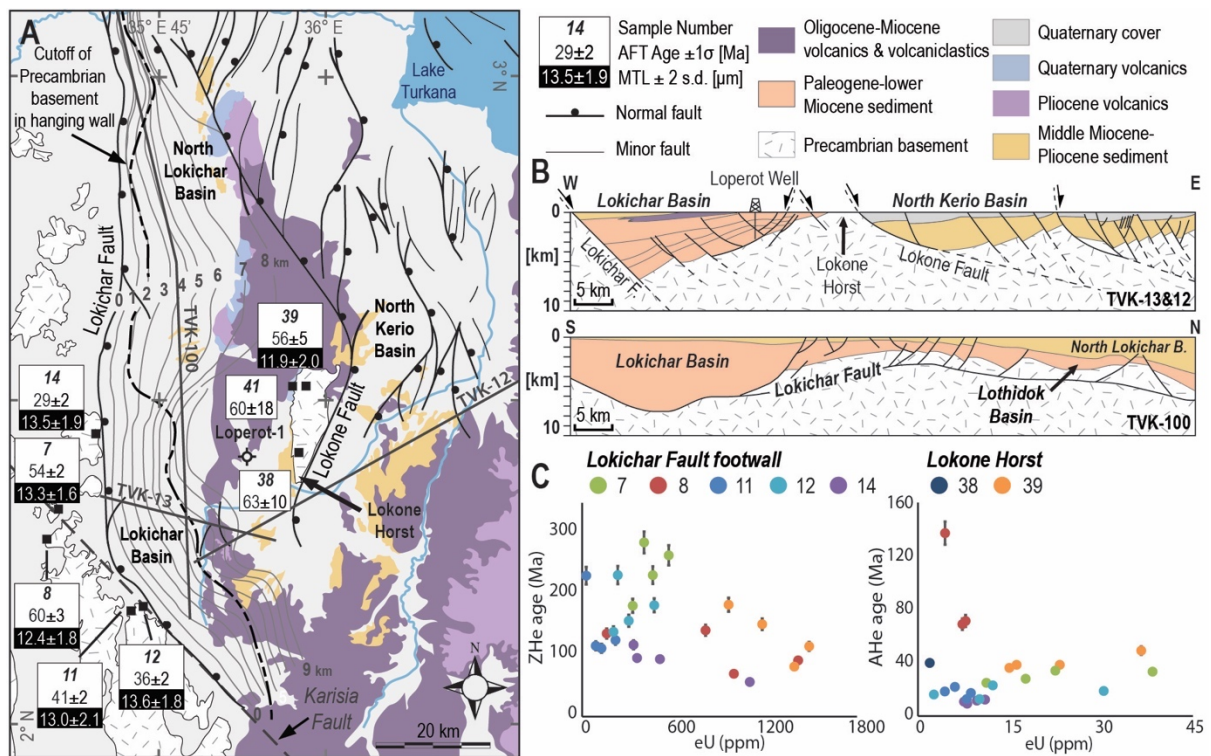


Figure 2.2. A: Geological map of the Lokichar Basin region, after [BEICIP, 1987], illustrating sample localities (black boxes) and AFT data. Lokichar Fault plane is illustrated as a series of depth contours (white lines) as determined from seismic data [Morley *et al.*, 1999]. Trace of inferred Karisia Fault [Vétel and Le Gall, 2006] marked by dashed grey line. B: Interpretations of the TVK-13 & 12 and TVK-100 seismic lines modified from Morley *et al.* (1999). C: ZHe (left) and AHe (right) ages versus effective uranium content (eU).

2.3 Methods and Results

Low-temperature thermochronology is a powerful tool for constraining the timing of rift basin formation and rates of fault slip [e.g., *Stockli*, 2005]. Over geological timescales, the ZHe, AFT and AHe systems are broadly sensitive to temperatures between ~210-130 °C, ~120-60 °C and ~90-30 °C, respectively [e.g. *Farley et al.*, 2002; *Gleadow and Duddy*, 1981; *Reiners*, 2005; see Supplementary Data for expanded methodology]. Combined application of these systems quantifies cooling and reheating associated with footwall uplift and hanging wall subsidence in response to extensional basin formation. Of eight Precambrian crystalline basement rock samples collected from the western and eastern margins of the Lokichar Basin (sample details are listed in Table SD2.1), five were collected from the Lokichar bounding fault footwall, and three from the Lokone Horst area of the half-graben flexural margin (Figure 2.2A).

Single grain ZHe ages ($n=20$), ranging from 285 ± 18 to 49 ± 3 Ma, exhibit significant intra- and inter-sample variation [Figure 2.2C and Table SD2.2]. The observed negative correlation between ZHe age and effective uranium content ($eU = U + 0.235 \cdot Th$; Figure 2.2C) and positive age-grain size relationships (Figure SD1) are consistent with the effects of radiation damage [*Guenther et al.*, 2013] and/or crystal size [*Reiners et al.*, 2005] on He diffusion.

AFT ages from the Lokichar Fault footwall range from 29 ± 2 to 60 ± 3 Ma with mean track lengths (MTL) of 12.4-13.6 μm and standard deviations of 1.56-2.10 μm (Figures 2.2A and Table SD2.1). AFT ages decrease and MTL increase with proximity to the Lokichar Fault (Figure 2.3A). By comparison, AFT ages from the Lokichar Basin flexural margin yield older ages (56 ± 5 to 63 ± 10 Ma) and shorter MTL (11.9 μm , with a standard deviation of 2.01 μm).

Apparent ages from the AHe system show no clear relationship to either eU (Figure 2.2C) or grain size (Figure SD2.2). However, the inter-sample spread in AHe results display a spatial relationship with Cenozoic structures. Single grain ages (8.3 ± 0.5 to 137 ± 9 Ma) and intra-sample age dispersion from western basin margin samples increase with distance into the Lokichar Fault footwall (Figure 2.3A). AHe ages from the eastern basin margin are far less dispersed, yielding ages from 36 ± 2 to 49 ± 3 Ma.

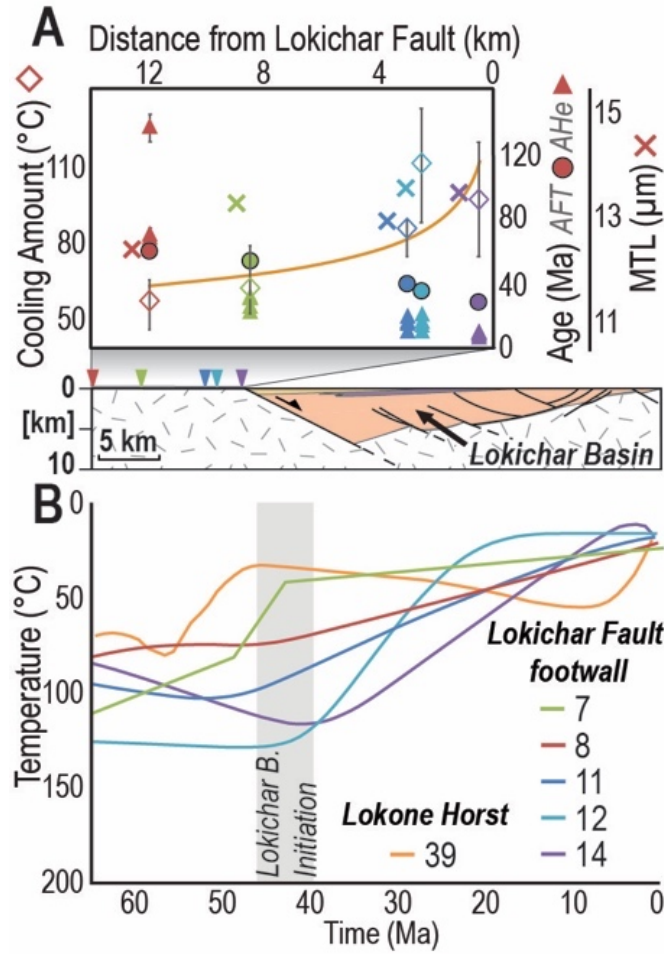


Figure 2.3. A: AFT and AHe ages, MTL and mid Eocene-Recent cooling magnitude (Figure 2.3B) for footwall samples versus distance from the Lokichar Fault, normal to strike. Trendline (power function) of cooling magnitudes in orange. MTL data are slightly offset for clarity. B: Preferred inverse thermal history models. Sample colours as for Figure 2.2.

2.4 Thermochronology of the Lokichar Basin

To quantify thermal histories for basement samples from the Lokichar Basin margins, AFT and single grain ZHe and AHe data were inversely modelled using the software *QTQt* [Gallagher *et al.*, 2012]. Model inputs included thermochronological data, AFT crystal Cl content, present-day surface temperatures, and radiation damage He diffusion kinetics models. Samples without confined AFT data (samples 38 & 41) were omitted during modelling. Preferred thermal history models are presented in Figure 2.3B. Thermal history models of basement samples record distinct phases of cooling and reheating along the Lokichar Basin margins during Cretaceous-Recent time. In this study we focus on the

Cenozoic portion of the thermal histories. The inverse modelling protocol and modelling results are presented in detail in the GSA Data Repository.

Samples with the highest degree of thermal overprinting provide the most reliable thermochronological constraints on the onset of rift-related cooling. Thermochronology data (Figure SD2.3) and thermal histories (Figure 2.3B) from the Lokichar Fault footwall record pronounced Eocene to Miocene cooling, which the most thermally rejuvenated samples (12 and 14) suggest began ~45-40 Ma. The magnitude of observed cooling increases with proximity to the Lokichar Fault (Figure 2.3A), consistent with the trend of decreasing AFT and AHe ages and increasing MTL for samples nearer the fault trace. These relationships, the exposure of highly eroded Precambrian basement along the western basin margin, and the occurrence of ~7 km of Eocene?-Oligocene to mid Miocene syn-rift sediment in the Lokichar Fault hanging wall [Morley *et al.*, 1999], are consistent with observed Eocene-Miocene cooling, interpreted here as resulting from isostatic footwall uplift and denudation during basin formation. The range of recorded cooling magnitudes (~60-110 °C) is equivalent to Eocene-Miocene denudation increasing from ~2.5 to ~4.5 km with proximity to the Lokichar Fault, using a conservative paleogeothermal gradient of 25 °C/km [average of Precambrian East African mobile belts Nyblade, 1990].

Thermal history modelling (Figure 2.3B) indicates that the Lokichar Basin flexural margin (sample 39) experienced reheating between ~45 and 10 Ma. The exposure of Oligocene-mid Miocene sequences on the western Lokone Horst and occurrence of Eocene?-Oligocene to Miocene deposits on the basin's flexural margin [Figure 2.2A; Morley *et al.*, 1999] are consistent with reheating resulting from burial and subsidence, consistent with middle Eocene-late Miocene formation of the Lokichar Basin. Sample 39 experienced subsequent mid Miocene-Recent cooling to surface temperatures, interpreted as uplift and denudation in response to formation of the adjacent, middle Miocene North Kerio Basin [Vétel and Le Gall, 2006].

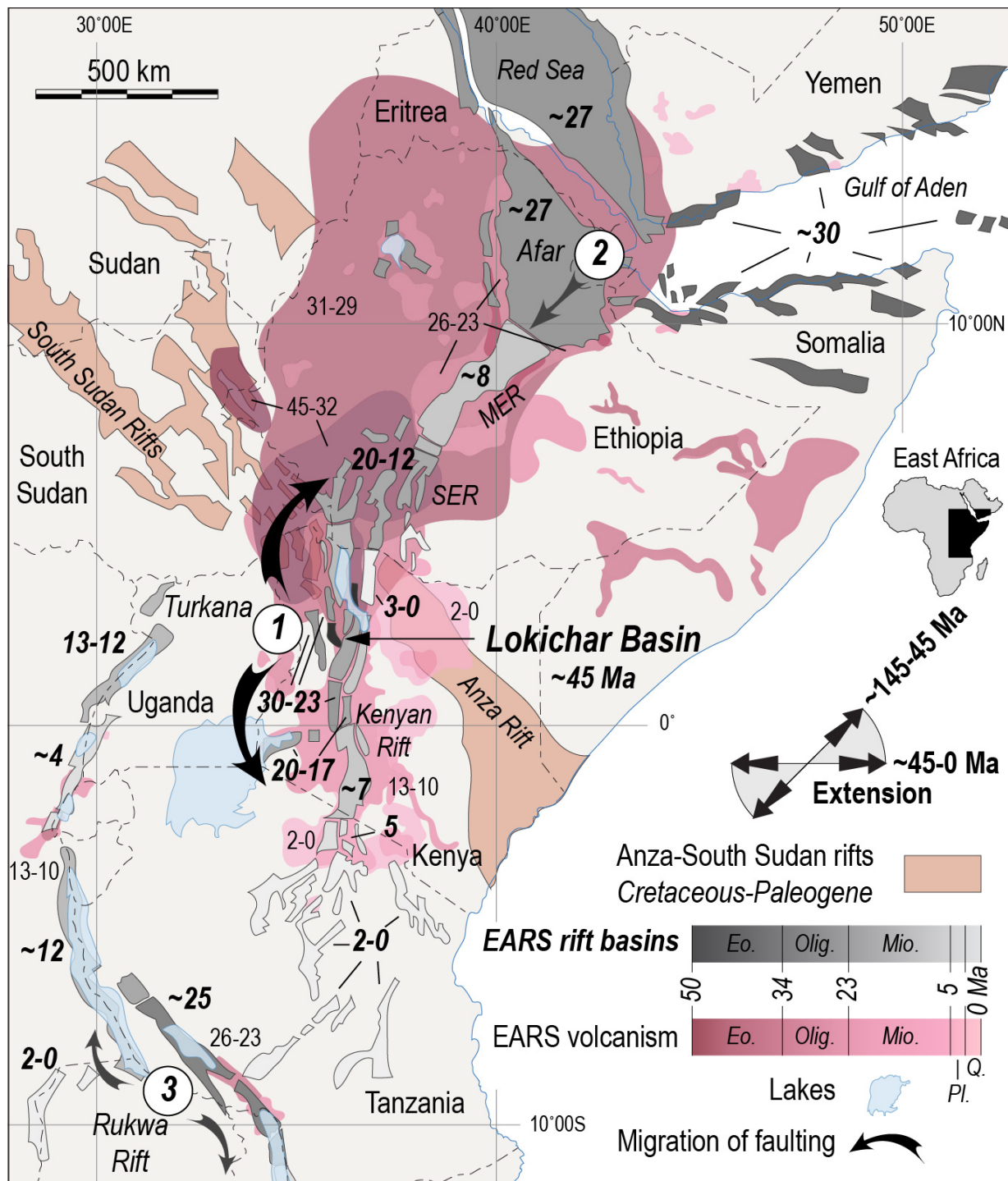


Figure 2.4. Chronology of Cretaceous-Recent rift basin formation and EARS-related volcanism in East Africa. Colours of EARS basins (grey) and volcanics (red) darken with age. EARS volcanism and faulting is suggested to have nucleated within inherently weak lithospheric zones (1: Turkana Depression, 2: Afar, 3: Rukwa Rift) before propagating north and south. Distribution of basins and surface volcanics, and corresponding ages from Purcell et al. [2017] and references therein. MER: Main Ethiopian Rift; SER: Southern Ethiopian Rift.

2.5 Re-evaluation of East African Rift Chronology

We present low-temperature thermochronology data and thermal history modelling that strongly advocate for an Eocene (~45-40 Ma) onset of EARS basin formation in the Turkana Depression, broadly concurrent with commencement of plume-related volcanism in northern Turkana and southern Ethiopia [Figure 2.4; *Morley*, 1994; *Ebinger and Sleep*, 1998]. Rift initiation in the Turkana Depression ca. 45 Ma significantly predates the onset of magmatism and strain elsewhere in the EARS by up to 15 Myr [*Purcell*, 2017; Figure 2.4]. The ~N-S orientation of the Eocene-lower Miocene Lokichar Basin suggests that the Turkana upper crust, at least locally, underwent ~E-W extension during late Paleogene times, oblique to earlier ~NE-SW Anza-South Sudan extension. Indeed, ~E-W extension may have been more widespread in Turkana during the Eocene, as suggested by the significant expansion (1100 m) of Eocene-Oligocene basalts across a N-S fault in eastern Turkana [*Vétel and Le Gall*, 2006] and the detection of two ~N-S basins in the western Turkana subsurface via remote sensing data, interpreted to be Cretaceous or Paleogene in age [*Wescott et al.*, 1999]. However, due to the poor quality of the seismic data and lack of exploration wells, their nature and age remain poorly constrained.

The next manifestations of rifting did not occur until the Oligocene, when voluminous basalt flows were erupted over much of Ethiopia and the surrounding areas ca. 31-29 Ma, followed by extension in the proto-Gulf of Aden (~30 Ma) and -Red Sea areas (~27 Ma) [*Purcell*, 2017]. In the late Oligocene, EARS magmatism and strain then affected the Rukwa Rift sector, far to the south [*Roberts et al.*, 2012]. Subsequent to the initial nucleation of EARS rifting in these discrete zones, extension propagated longitudinally and eventually linked to form a largely continuous rift by the Pliocene [*Purcell*, 2017]. In the case of the Turkana Depression, strain became more widespread locally during the Oligocene [*Morley et al.*, 1999] before migrating both south and north to form the Kenyan Rift [Miocene, possibly late Oligocene; *Vétel and Le Gall*, 2006] and the broadly rifted zone of southern Ethiopia [early Miocene; *Purcell*, 2017]. Meanwhile within Turkana, volcanism and faulting migrated eastwards during the Neogene [e.g. *Morley et al.*, 1999].

2.6 Rift Nucleation in the Turkana Depression

The much earlier occurrence of E-W extensional basin formation in the Turkana region compared to elsewhere in the EARS suggests that a unique set of circumstances facilitated strain nucleation in the Turkana Depression during that time. In other early segments of the EARS onset of extension was assisted by either thermal modification and weakening of the lithosphere by mantle plume activity [e.g. Afar, *Ebinger and Sleep*, 1998], or by pre-existing lithospheric weakness zones inherited from earlier tectonism [e.g. Rukwa Rift, *Roberts et al.*, 2012].

Strain focusing in the Turkana Depression, by contrast, may reflect the interplay of both mantle plume activity [e.g. *Ebinger and Sleep*, 1998] and inherited rheological heterogeneities, such as transverse brittle weakness zones [i.e. Karisia Fault; *Vétel and Le Gall*, 2006] and pre-attenuated lithosphere [*Brune et al.*, 2017]. Furthermore, the pre-existing lithospheric architecture may have caused early focusing of magmatism in the Turkana Depression by channelling upwelling plume material into the region of thin lithosphere beneath the Cretaceous-early Paleogene rifts of eastern South Sudan [*Ebinger and Sleep*, 1998].

2.7 Implications for the Onset of East African Rifting

By constraining the spatio-temporal onset of ~E-W extension in East Africa, this study adds valuable insight into the causal mechanisms for EARS inception. Modern intraplate deformation is considered to be generated by buoyancy forces arising from East African topography [*Stamps et al.*, 2015]. However, there is a lack of strong supporting evidence for regional domal topography in Turkana at any stage during the Cenozoic, despite its significant history of plume-related volcanism [*Vétel and Le Gall*, 2006]. Even if a Paleogene ‘Turkana Dome’ did once exist, as significant Paleogene denudation recorded throughout much of Turkana and the western Anza Rift may suggest (Chapter 3), the presence of a directional far-field stress field is required to develop a non-axisymmetric, linear rift [*Burov and Gerya*, 2014], such as the N-S trending Eocene-Recent basins of the Turkana Depression (i.e. Lokichar Basin). If buoyancy forces did not contribute to the force budget during the onset of Turkana Depression rifting, some other mechanism is still required to have initiated Eocene upper crustal dilation. The apparent synchronicity between the rotation of the East

African paleostress regime from ~NE-SW to ~E-W and a major reorganization of plate motion in the Indian Ocean ca. ~45-42 Ma [*Cande et al.*, 2018] suggest that plate boundary forces may have played an important role during EARS inception. Horizontal basal traction resulting from divergent mantle flow under Africa [*Ghosh and Holt*, 2012], may too have been significant. These observations highlight the importance of directional far-field stress fields during the initiation of magma-assisted intracontinental rifts.

References

- BEICIP. (1987). Geological Map of Kenya. Nairobi: Ministry of Energy and Regional Development.
- Benoit, M. H., Nyblade, A. A., & Pasyanos, M. E. (2006). Crustal thinning between the Ethiopian and East African plateaus from modeling Rayleigh wave dispersion. *Geophysical Research Letters*, 33(13).
- Brune, S., Corti, G., & Ranalli, G. (2017). Controls of inherited lithospheric heterogeneity on rift linkage: Numerical and analog models of interaction between the Kenyan and Ethiopian rifts across the Turkana depression. *Tectonics*, 1–30.
- Buck, W. R. (2004). Consequences of asthenospheric variability on continental rifting. In G. B. Karner, N. D. Taylor, & B. Kohlstedt (Eds.), *Rheology and deformation of the lithosphere at continental margins* (pp. 92–137). Columbia University Press.
- Burov, E., & Gerya, T. (2014). Asymmetric three-dimensional topography over mantle plumes. *Nature*, 513(7516), 85–89.
- Cande, S. C., Patriat, P., & Dymant, J. (2018). Motion between the Indian, Antarctic and African plates in the early Cenozoic. *Geophysical Journal International*, 183, 127–149.
- Ebinger, C. J., & Sleep, N. H. (1998). Cenozoic magnetism throughout east Africa resulting from impact of a single plume. *Nature*, 395, 788–791.
- Farley, K. A., & Stockli, D. F. (2002). (U-Th)/He Dating of Phosphates: Apatite, Monazite, and Xenotime. *Reviews in Mineralogy and Geochemistry*, 48(1), 559–577.
- Gallagher, K. (2012). Transdimensional inverse thermal history modeling for quantitative thermochronology. *Journal of Geophysical Research: Solid Earth*, 117, 1–16.
- Ghosh, A., & Holt, W. E. (2012). Plate Motions and Stresses from Global Dynamic Models. *Science*, 335, 838–843.
- Gleadow, A. J. W., & Duddy, I. R. (1981). A Natural Long-Term Track Annealing Experiment for Apatite. *Nuclear Tracks*, 5(1–2), 169–174.
- Guenther, W. R., Reiners, P. W., Ketcham, R. A., Nasdala, L., & Giester, G. (2013). Helium diffusion in natural zircon: radiation damage, anisotropy, and the interpretation of zircon (U-TH)/He thermochronology. *American Journal of Science*, 313(3), 145–198.
- Morley, C. K. (1994). Interaction of deep and shallow processes in the evolution of the Kenya rift. *Tectonophysics*, 236, 81–91.
- Morley, C. K., Stone, D. M., Harper, R. M., & Wigger, S. T. (1999). Geology and Geophysics of the Western Turkana Basins, Kenya. In C. K. Morley (Ed.), *Geoscience of Rift Systems—Evolution of East Africa: AAPG Studies in Geology No. 44* (pp. 19–54).
- Morley, C. K., Wescott, W. a., Stone, D. M., Harper, R. M., Wigger, S. T., & Karanja, F. M. (1992). Tectonic evolution of the northern Kenyan Rift. *Journal of the Geological Society*, 149(3), 333–348.
- Moucha, R., & Forte, A. M. (2011). Changes in African topography driven by mantle convection. *Nature Geoscience*, 4(10), 707–712. <https://doi.org/10.1038/ngeo1235>
- Purcell, P. G. (2017). Re-imagining and re-imaging the development of the East African Rift. *Petroleum Geoscience*, 36.
- Reiners, P. W. (2005). Zircon (U-Th)/He Thermochronometry. *Reviews in Mineralogy and Geochemistry*, 58, 151–179.
- Roberts, E. M., Stevens, N. J., O'Connor, P. M., Dirks, P. H. G. M., Gottfried, M. D., Clyde, W. C., ... Hemming, S. (2012). Initiation of the western branch of the East African Rift coeval with the eastern branch. *Nature Geoscience*, 5(4), 289–294.
- Stamps, D. S., Iaffaldano, G., & Calais, E. (2015). Role of mantle flow in Nubia-Somalia plate divergence. *Geophysical Research Letters*, 42, 290–296.

- Stockli, D. F. (2005). Application of Low-Temperature Thermochronometry to Extensional Tectonic Settings. *Reviews in Mineralogy and Geochemistry*, 58, 411–448.
- Vétel, W., & Le Gall, B. (2006). Dynamics of prolonged continental extension in magmatic rifts: the Turkana Rift case study (North Kenya). *Geological Society of London Special Publications*, 259, 209–233.
- Wescott, W. A., Wigger, S. T., Stone, D. M., & Morley, C. K. (1999). Geology and Geophysics of the Lotikipi Plain. *Geoscience of Rift Systems-Evolution of East Africa*, (44), 55–66.

Supplementary Data

Methodology

Eight Precambrian crystalline basement rock samples were collected to constrain the low-temperature thermal history of the Lokichar Basin (Figure 2.2A). Sample details, including rock types, coordinates and elevations are listed in Table SD2.1. For simplicity, samples referred to in the text and figures simply by the last digits of their sample code (i.e. 39 instead of TUB13-39) but are listed in full in the data tables.

The AFT methodology is based on the formation of damage trails in apatite, called fission tracks, which develop as a consequence of spontaneous fission of ^{238}U [Wagner and van den Haute, 1992]. Fission tracks are temperature sensitive and begin to anneal when exposed to temperatures above $\sim 60^\circ\text{C}$ over geological timescales. At temperatures above $\sim 110\text{--}120^\circ\text{C}$ fission tracks anneal completely [Gleadow and Duddy, 1981]. Within this temperature sensitivity range, referred to as the partial annealing zone (PAZ), fission tracks anneal progressively to a length corresponding to the maximum temperature they have experienced [Laslett *et al.*, 1987]. It has been shown that the annealing behaviour of fission tracks in apatite is also compositionally controlled, with chlorine-rich apatites being more resistant to annealing [Barbarand *et al.*, 2003]. By combining fission track age and length data the thermal history of a sample within the PAZ can be constrained [Ketcham *et al.*, 2005].

AHe and ZHe thermochronology are based upon the production of ^4He during the α -decay of ^{238}U , ^{235}U , ^{232}Th and, in minerals with low to moderate U and Th (e.g. apatite), ^{147}Sm [Farley and Stockli, 2002]. The diffusion of ^4He is a function of time, temperature, crystal size and accumulated radiation damage [Reiners and Farley, 2001]. In apatite, ^4He diffusion accelerates significantly above $\sim 30^\circ\text{C}$ to nearly instantaneous loss of all ^4He above $\sim 90^\circ\text{C}$ [Reiners *et al.*, 2005]. Within these temperatures, referred to as the apatite partial retention zone, ^4He is partially retained. Zircon exhibits a higher temperature sensitivity, with a zircon partial retention zone of $\sim 130\text{--}210^\circ\text{C}$ [Reiners *et al.*, 2002] for typical igneous zircons.

A wide range of crystallographic and compositional characteristics have been shown to affect He diffusion in apatite and zircon, at times producing intra-sample He age dispersion greater than 50% [e.g. Wildman *et al.*, 2016]. Some of these potential error sources can be avoided during the crystal selection procedure, such as U-Th inclusions [Anderson *et al.*, 2017]. Other

variables accounted for in this study are grain size, morphology and accumulated radiation damage, all of which are considered during data interpretation and thermal history modelling. Here, we use effective uranium concentration ($eU = U + 0.235 \times Th$) as a proxy for α -radiation damage, which considers the decay rate and abundance of radiogenic parent elements. The potential effect grain size may have on resultant He age is assessed by the proxy, equivalent spherical radius (R_s ; $R_s = [3 \cdot R \cdot L] / [2 \cdot \{R + L\}]$) [Beucher *et al.*, 2013], allowing the comparison of 0, 1 and 2 termination grains.

Samples 38 and 41 yielded insufficient apatite and zircon grains for He analysis. For samples that did yield suitable grains for ZHe and AHe analyses, an alpha ejection correction (F_T) was applied to calculate a corrected age (Tables SD2.2 and SD2.3), as suggested by Farley *et al.* [1996]. However, it has been shown that for samples with protracted thermal histories in the PRZ, as is the case for slowly cooled basement terranes, volume diffusion is the dominant mechanism governing He loss. Thus, the application of the F_T correction in these cases will result in inaccurately old ages [Danišik *et al.*, 2008]. Therefore, all ZHe and AHe data discussed in the text and displayed in figures are uncorrected ages. The effects of alpha ejection are, however, accounted for during inverse thermal history modelling (see Thermal History Modelling Protocol). All zircon grains selected for analysis were euhedral with two crystal terminations.

A comprehensive review of the AFT, ZHe and AHe methodologies and technical procedures used in this study can be found in Gleadow *et al.* [2015]. A summary of the AFT, ZHe and AHe analytical protocols are given in the footnotes of Tables SD2.1, SD2.2 and SD2.3, respectively.

Table SD2.1. Lokichar Basin apatite fission track summary table

Sample No.	Rock Type	Longitude/Latitude	Elevation [m]	No. of grains	N _s	ρ _s [10 ⁵ cm ⁻²]	²³⁸ U [ppm ± 1σ]	Cl [wt%]	Dispersion [%]	Pooled age [Ma ± 1σ]	Central age [Ma ± 1σ]	N _{length}	Mean track length [μm ± se]	St.Dev. [μm]
Lokichar Footwall (North to South)														
TUB16-14	Granulitic gneiss	35.65581/2.44495	770	27	316	1.3894	10.89 ± 5.99	0.04	22	28.6 ± 2.2	29.3 ± 2.2	100	13.5 ± 0.2	1.93
Northern transect (West-East)														
TUB16-8	Biotite-Granulite	35.57142/2.28210	837	30	760	2.0687	7.16 ± 3.02	0.03	18	58.4 ± 3.8	60.4 ± 3.2	100	12.4 ± 0.2	1.82
TUB16-7	Biotite-Granulite	35.59524/2.32415	846	30	3492	7.3648	28.8 ± 10.32	0.28	11	53.9 ± 1.8	54.4 ± 1.6	100	13.3 ± 0.2	1.56
Southern transect (West-East)														
TUB16-11	Granulitic gneiss	35.72054/2.17156	907	28	645	1.5397	8.52 ± 4.46	0.11	21	38.1 ± 2.9	40.6 ± 2.4	100	13.0 ± 0.2	2.10
TUB16-12	Granulitic gneiss	35.72213/2.17118	837	25	987	2.2584	16.32 ± 13.87	0.08	20	34.5 ± 2.3	35.8 ± 2.0	100	13.6 ± 0.2	1.75
Lokhone Horst (North to South)														
Northern transect (West-East)														
TUB13-41	Biotite Gneiss	35.958500/2.516400	645	7	59	2.8997	17.5 ± 24.90	0.14	62	46.1 ± 18.1	60.1 ± 17.8	-	- ± -	-
TUB13-39	Biotite Gneiss	35.984500/2.519400	632	23	1621	7.8726	33.4 ± 22.11	0.06	36	53.6 ± 6.2	56.2 ± 4.9	176	11.9 ± 0.2	2.01
Southern horst														
TUB13-38	Garnet Granulitic Gneiss	35.973534/2.412424	621	12	47	0.6504	2.37 ± 1.37	0.01	0	60.0 ± 7.1	62.7 ± 10.0	-	- ± -	-

Notes: Apatite grains were mounted and polished, and fission tracks analysed using the LA-ICP-MS technique following procedures described in *Gleadow et al.* [2015]. Analyses were performed by Samuel C Boone on image sets captured by TrackWorks using a 3.2MP AVT Oscar F-320C camera mounted on a Zeiss AxioImager microscope with a 1000x total magnification and a 100x dry objective (calibration = 0.07 μm by 0.07 μm per pixel). Spontaneous track densities were measured on prismatic internal apatite surfaces after etching with 5M HNO₃ for 20sec at 20°C. Track counts were obtained by automated counting in FastTracks using the 'coincidence mapping' technique of *Gleadow et al.* [2009] followed by manual inspection. Uranium concentrations of each grain were determined by LA-ICP-MS single spot analysis using a New Wave Nd:YAG Laser (λ=213nm with 5Hz @ 45% power, spot size=30μm) connected to an Agilent 7700 mass spectrometer. NIST612 was used as the primary reference material and a sintered Mud Tank Carbonatite apatite as an in-house secondary reference material during LA-ICP-MS analyses, following the protocol of *Boone et al.* [2016]. Data reduction of LA-ICP-MS results were performed using the TraceElementst_IS data reduction scheme in software package Lolite [*Paton et al.*, 2011]. Single grain and pooled ages were calculated according to *Hasebe et al.* [2004]. Central ages were estimated from single grain ages and errors according to the formulas given by *Galbraith* [2005, p.100] using the Newton-Raphson method. All ages are "model" ages obtained using a range factor (Rs) of 7.17μm (average mean track length of Durango and Fish Canyon Tuff standards) and are directly comparable to conventional External Detector Method ages [*Seiler et al.*, 2014]. Confined track lengths (TINTs) were measured as true 3D lengths using *FastTracks* after irradiation by ²⁵²Cf and are corrected for a refractive index of 1.634 for apatite. Chlorine concentrations of individual apatite grains were measured using a Cameca SX50 electron microprobe.

N_s = number of spontaneous tracks counted; ρ_s = spontaneous track density; D_{par} = long axis of track etch pit; N_{length} = number of lengths measured; se = standard error

Table SD2.2. Lokichar Basin Zircon (U-Th)/He Data Summary

Sample No.	Lab #	⁴ He [ncc]	Mass [mg]	U [ppm]	Th [ppm]	Th/U ratio	[eU] (ppm) ^a	Grain length [μm]	Grain half-width [μm]	R _s ^b [μm]	F _r ^c	Uncorrected Age [Ma ± 1σ]	Corrected Age [Ma ± 1σ]	^d Crystal morphology
Lokichar Footwall (North to South)														
TUB16-14	16065	56.844	0.0096	315.7	77.2	0.24	333.8	254.9	54.3	67.1	0.81	116.0 ± 7.2	143.4 ± 8.9	2T
TUB16-14	16066	57.708	0.0114	338.1	84.7	0.25	358.0	275.4	56.2	70.0	0.82	94.6 ± 5.9	115.4 ± 7.2	2T
TUB16-14	16067	91.856	0.0098	993.0	450.6	0.45	1098.9	285.4	49.9	63.7	0.80	55.6 ± 3.4	69.4 ± 4.3	2T
TUB16-14	16068	67.285	0.0093	484.8	98.5	0.20	508.0	286.8	48.1	61.8	0.80	92.5 ± 5.7	115.5 ± 7.2	2T
Northern transect (West-East)														
TUB16-08	16061	70.170	0.0064	970.0	104.8	0.11	994.6	275.9	39.7	52.1	0.77	69.0 ± 4.3	89.9 ± 5.6	2T
TUB16-08	16062	115.630	0.0062	778.3	127.4	0.16	808.2	289.5	37.7	50.0	0.75	140.2 ± 8.7	186.2 ± 11.5	2T
TUB16-08	16063	28.596	0.0087	146.1	48.9	0.33	157.6	267.4	48.9	62.0	0.80	134.5 ± 8.3	168.5 ± 10.4	2T
TUB16-08	16064	83.449	0.0065	1640.6	147.1	0.09	1675.1	251.2	42.9	55.0	0.77	48.6 ± 3.0	62.9 ± 3.9	2T
TUB16-07	16053	88.143	0.0094	319.1	45.5	0.14	329.8	298.9	46.9	60.8	0.79	180.4 ± 11.2	227.7 ± 14.1	2T
TUB16-07	16054	471.020	0.0212	547.8	77.4	0.14	566.0	391.0	61.5	79.8	0.84	263.0 ± 16.3	312.6 ± 19.4	2T
TUB16-07	16055	494.130	0.0321	448.7	54.6	0.12	461.6	446.4	70.9	91.8	0.86	230.6 ± 14.3	267.1 ± 16.6	2T
TUB16-07	16056	355.456	0.0213	391.4	49.3	0.13	403.0	339.8	69.1	86.2	0.86	284.1 ± 17.6	330.1 ± 20.5	2T
Southern transect (West-East)														
TUB16-11	16049	17.279	0.0115	65.9	85.4	1.29	86.0	291.5	53.8	68.1	0.81	114.5 ± 7.1	142.2 ± 8.8	2T
TUB16-11	16050	66.585	0.0166	211.0	24.5	0.12	216.7	377.1	54.7	71.7	0.82	123.8 ± 7.7	150.5 ± 9.3	2T
TUB16-11	16051	30.988	0.0157	103.2	83.5	0.81	122.8	256.3	76.1	88.1	0.84	110.2 ± 6.8	130.7 ± 8.1	2T
TUB16-11	16052	18.313	0.0228	19.8	16.9	0.85	23.8	368.1	67.4	85.5	0.85	229.9 ± 14.3	270.7 ± 16.8	2T
TUB16-12	16057	204.884	0.0165	458.9	49.9	0.11	470.6	270.4	73.7	86.9	0.85	181.4 ± 11.2	212.7 ± 13.2	2T
TUB16-12	16058	200.660	0.0252	224.8	27.0	0.12	231.2	510.3	56.3	76.0	0.84	230.9 ± 14.3	275.4 ± 17.1	2T
TUB16-12	16059	62.638	0.0086	291.7	46.2	0.16	302.5	269.2	48.1	61.2	0.80	155.8 ± 9.7	195.0 ± 12.1	2T
TUB16-12	16060	49.522	0.0112	194.7	35.2	0.18	202.9	414.0	41.3	56.3	0.78	137.8 ± 8.5	176.0 ± 10.9	2T
Lokhone Horst (North to South)														
Northern transect														
TUB13-39	13176	651.167	0.0253	936.9	97.4	0.10	959.8	483.3	58.4	78.1	0.84	182.0 ± 11.3	216.3 ± 13.4	2T
TUB13-39	13177	205.576	0.0121	1361.4	127.3	0.09	1391.3	335.9	49.9	65.2	0.81	80.7 ± 5.0	99.6 ± 6.2	2T
TUB13-39	13178	320.453	0.0126	1456.2	140.6	0.10	1489.3	323.2	53.3	68.7	0.82	113.7 ± 7.0	139.2 ± 8.6	2T
TUB13-39	13179	493.791	0.0192	1157.2	98.3	0.08	1180.3	324.6	64.9	81.2	0.85	149.8 ± 9.3	176.6 ± 11.0	2T
Fish Canyon Tuff standard														
Fish Canyon	13571	9.458	0.0071	333.1	152.5	0.46	369.0	238.5	47.3	-	0.78	23.4 ± 1.4	29.8 ± 1.8	2T
Fish Canyon	13560	11.128	0.0116	231.3	118.7	0.51	259.2	297.8	53.2	-	0.81	24.6 ± 1.5	30.4 ± 1.9	2T
Fish Canyon	13459	6.811	0.0059	300.0	154.8	0.52	336.4	211.7	47.1	-	0.78	21.8 ± 1.4	28.1 ± 1.7	2T
Fish Canyon	16045	13.432	0.0092	367.2	212.4	0.58	417.1	260.8	51.5	-	0.80	23.1 ± 1.4	28.8 ± 1.8	2T
Fish Canyon	16046	8.101	0.0060	355.3	211.2	0.59	404.9	232.6	43.7	-	0.75	20.5 ± 1.3	27.4 ± 1.7	2T
Fish Canyon	16154	11.162	0.0064	430.9	232.5	0.54	485.5	223.7	47.1	-	0.78	23.1 ± 1.4	29.5 ± 1.8	2T
Fish Canyon	16170	5.499	0.0047	316.7	175.4	0.55	357.9	204.9	41.7	-	0.75	20.3 ± 1.3	27.0 ± 1.7	2T
Fish Canyon	16171	4.048	0.0032	325.3	195.3	0.60	371.2	155.0	43.5	-	0.74	20.5 ± 1.3	27.7 ± 1.7	2T
Fish Canyon	16335	9.257	0.0054	434.6	244.6	0.56	492.1	211.1	44.6	-	0.77	21.9 ± 1.4	28.5 ± 1.8	2T

Notes: Clear, euhedral, non-fractured zircons with two terminations were separated manually by hand-picking under an Olympus SZX12 binocular microscope. Zircon grain geometries were imaged and measured microscopically and the loaded into acid-treated platinum capsules. Zircons were then outgassed under vacuum at ~1300°C for 20 minutes, using a semiconductor Coherent Quattro FAP 820 nm diode laser with a fibre-optic coupling to the sample chamber. He volume was determined by isotope dilution using a pure ³He spike, calibrated against an independent ⁴He standard and measured using a Balzers quadrupole (QMS 200 – Prisma) mass spectrometer. A hot blank was run after each gas extraction to verify complete outgassing of the apatite grains. Outgassed zircon grains were transferred from their Pt capsules into Parr bombs, where they were spiked with ²³³U and ²²⁹Th and digested at 240°C for 40 hours in HF. Fish Canyon Tuff zircon (Gleadow et al., 2015) was also run as an ‘unknown’ with each batch zircon samples and served as a check on sample accuracy. Standards and unspiked reagent blanks, were treated identically as unknowns. Samples underwent a second bombing in HCl for 24 hours at 200°C, ensuring dissolution of fluoride salts. After drying, samples were dissolved in HNO₃ solutions and diluted in H₂O to 5% for analysis of ²³⁸U, ²³³U and ²²⁹Th by solution using an Agilent 7700X ICP Mass Spectrometer. For single crystals digested in small volumes (0.3-0.5 ml), U and Th isotope ratios were measured to a precision of up to 2%, but typically better than 1% on an Agilent 7700x ICP-MS. Overall precision for the ZHe method is ~6% or less. α-emission corrected ZHe ages were calculated following the approach of Hourigan et al. (2005).

^a eU = U ppm + Th ppm * 0.235

^b R_s (equivalent spherical radius) = [3•R•L]/[2•(R+L)], where R = grain radius and L = grain length

^c F_r is the α-ejection correction after Farley et al., [1996].

^d Grain morphology - 0T = no crystal terminations, 1T = one crystal termination and 2T = 2 crystal terminations.

Table SD2.3. Lokichar Basin apatite (U-Th-Sm)/He data summary

Sample No.	Lab #	⁴ He [ncc]	Mass [mg]	U [ppm]	Th [ppm]	Sm [ppm]	Th/U ratio	[eU] (ppm) ^a	Grain length [μm]	Grain width [μm]	R _s ^b [μm]	F _T ^c	Uncorrected Age [Ma ± 1σ]	Corrected Age [Ma ± 1σ]	^d Crystal morphology
Lokichar Footwall (North to South)															
TUB16-14	16093	0.124	0.00782	10.5	1.5	371.7	0.14	10.9	188.7	84.2	51.7	0.83	11.5 ± 0.7	14.0 ± 0.9	1T
TUB16-14	16094	0.277	0.02977	7.3	0.5	262.2	0.06	7.4	343.3	92.9	61.4	0.87	9.9 ± 0.6	11.3 ± 0.7	OT
TUB16-14	16095	0.289	0.02243	6.5	12.8	256.7	1.96	9.5	248.8	94.7	59.7	0.85	10.8 ± 0.7	12.7 ± 0.8	OT
TUB16-14	16096	0.098	0.01184	7.5	2.0	223.5	0.27	8.0	186.5	79.5	49.1	0.82	8.3 ± 0.5	10.0 ± 0.6	OT
Northern transect (West-East)															
TUB16-8	16097	2.732	0.04020	7.3	1.6	151.4	0.22	7.7	281.5	119.2	73.8	0.88	70.8 ± 4.4	80.1 ± 5.0	OT
TUB16-8	16098	0.597	0.00965	6.8	1.3	212.3	0.19	7.1	237.5	69.9	45.7	0.80	68.5 ± 4.2	85.7 ± 5.3	2T
TUB16-8	16099	1.657	0.02205	3.9	1.6	117.3	0.40	4.3	263.2	91.3	58.3	0.85	137.0 ± 8.5	160.5 ± 9.9	OT
TUB16-7	16089	0.556	0.01708	10.4	2.8	105.3	0.27	11.1	205.8	90.9	55.8	0.84	23.9 ± 1.5	28.3 ± 1.8	OT
TUB16-7	16090	0.887	0.01521	16.8	2.9	129.8	0.17	17.5	231.5	80.8	51.6	0.84	27.1 ± 1.7	32.4 ± 2.0	OT
TUB16-7	16091	1.231	0.01351	21.7	3.1	100.0	0.14	22.4	253.4	72.8	47.7	0.83	33.2 ± 2.1	40.0 ± 2.5	OT
TUB16-7	16092	1.372	0.00897	35.7	11.1	210.2	0.31	38.3	270.2	69.4	46.1	0.81	32.5 ± 2.0	40.1 ± 2.5	1T
Southern transect (West-East)															
TUB16-11	16081	0.094	0.00781	7.5	1.7	366.3	0.23	7.9	241.4	69.4	45.5	0.81	11.8 ± 0.7	14.6 ± 0.9	1T
TUB16-11	16082	0.186	0.01853	4.1	0.7	301.3	0.18	4.3	272.8	90.4	58.2	0.84	17.6 ± 1.1	21.0 ± 1.3	2T
TUB16-11	16083	0.258	0.01447	8.2	1.4	286.6	0.17	8.5	249.7	97.0	60.9	0.85	16.5 ± 1.0	19.3 ± 1.2	1T
TUB16-11	16084	0.768	0.04819	5.7	1.1	234.9	0.19	6.0	463.6	111.8	74.8	0.88	21.1 ± 1.3	24.0 ± 1.5	2T
TUB16-12	16085	0.130	0.02620	2.3	0.7	152.3	0.30	2.5	246.1	102.9	63.8	0.86	15.2 ± 0.9	17.6 ± 1.1	OT
TUB16-12	16086	0.484	0.03382	9.6	1.6	120.7	0.17	10.0	313.7	103.6	66.7	0.88	11.7 ± 0.7	13.3 ± 0.8	OT
TUB16-12	16087	0.439	0.01293	10.5	7.1	276.4	0.67	12.2	231.4	82.0	52.2	0.81	22.2 ± 1.4	27.3 ± 1.7	2T
TUB16-12	16088	1.309	0.01973	29.8	2.2	263.2	0.08	30.3	264.0	86.2	55.6	0.85	17.8 ± 1.1	20.8 ± 1.3	OT
Lokhone Horst (North to South)															
Northern transect															
TUB13-039	13043	0.863	0.0114	15.8	1.4	138.2	0.09	16.1	283.5	154.0	90.9	0.83	37.9 ± 2.3	45.5 ± 2.8	1T
TUB13-039	13044	3.846	0.0360	23.0	0.8	81.7	0.03	23.2	362.7	198.7	117.0	0.88	37.7 ± 2.3	42.9 ± 2.7	OT
TUB13-039	13045	1.089	0.0167	14.7	1.0	162.1	0.07	14.9	329.8	171.5	102.1	0.85	35.5 ± 2.2	41.7 ± 2.6	1T
TUB13-039	13046	4.752	0.0217	34.8	7.6	184.6	0.22	36.6	347.4	192.2	112.9	0.86	48.6 ± 3.0	56.3 ± 3.5	1T
Southern horst															
TUB13-38	13040	0.377	0.0388	1.8	0.7	41.7	0.39	2.0	275.6	236.8	124.2	0.88	39.1 ± 2.4	44.5 ± 2.8	OT
Durango standard															
Durango	13055	3.974	-	-	-	-	26.42	-	-	-	-	1.00	- ± -	31.3 ± 1.9	-
Durango	13056	4.347	-	-	-	-	26.04	-	-	-	-	1.00	- ± -	31.4 ± 1.9	-
Durango	13073	2.069	-	-	-	-	19.61	-	-	-	-	1.00	- ± -	31.2 ± 1.9	-
Durango	13109	11.244	-	-	-	-	21.41	-	-	-	-	1.00	- ± -	31.3 ± 1.9	-
Durango	15910	31.234	-	-	-	-	19.99	-	-	-	-	1.00	- ± -	32.3 ± 2.0	-
Durango	15952	31.993	-	-	-	-	23.23	-	-	-	-	1.00	- ± -	29.8 ± 1.8	-
Durango	16008	15.041	-	-	-	-	20.72	-	-	-	-	1.00	- ± -	29.7 ± 1.8	-
Durango	16572	2.922	-	-	-	-	21.41	-	-	-	-	1.00	- ± -	31.5 ± 2.0	-
Durango	16587	1.968	-	-	-	-	15.76	-	-	-	-	1.00	- ± -	31.3 ± 1.9	-
Durango	16588	1.255	-	-	-	-	21.82	-	-	-	-	1.00	- ± -	31.8 ± 2.0	-

Notes: Clear, euhedral, non-fractured apatites were separated manually by hand-picking under an Olympus SZX12 binocular microscope. Apatites, in acid-treated platinum capsules, were outgassed under vacuum at ~900°C for 5 minutes, using a semiconductor Coherent Quattro FAP 820 nm diode laser with fibre-optic coupling to the sample chamber. He volume was determined by isotope dilution using a pure ³He spike, calibrated against an independent ⁴He standard and measured using a Balzers quadrupole (QMS 200 – Prisma) mass spectrometer. A hot blank was run after each gas extraction to verify complete outgassing of the apatite grains. The second re-extract contributed less than 0.5% of the total measured ⁴He for all samples. ²³⁸U, ²³⁵U, ²³²Th and ¹⁴⁷Sm content was obtained by total dissolution of outgassed apatites, still in platinum capsules, in HNO₃ and analysed using an Agilent 7700x ICP-MS. Durango apatite [McDowell *et al.*, 2005] and Fish Canyon Tuff zircon [Gleadow *et al.*, 2015] were also run as ‘unknowns’ with each batch of apatite or zircons samples and served as a check on sample accuracy. (U-Th-Sm)/He ages were calculated and corrected for α-emission following the approach of Farley *et al.*, [1996]. Analytical uncertainties, including the α-ejection correction, an estimated 5 μm uncertainty in grain dimension, gas analysis (<1%) and ICP-MS analytical uncertainties, are conservatively assessed at ~6.2%. Accuracy and precision of U, Th and Sm content range up to 2%, but are typically <1%.

^a eU (ppm) = U ppm + 0.235*Th ppm

^b R_s (equivalent spherical radius) = [3•R•L]/[2•(R+L)], where R = grain radius and L = grain length

^c F_T is the α-ejection correction after Farley *et al.*, [1996].

^d Grain morphology - OT = no crystal terminations, 1T = one crystal termination and 2T = 2 crystal terminations.

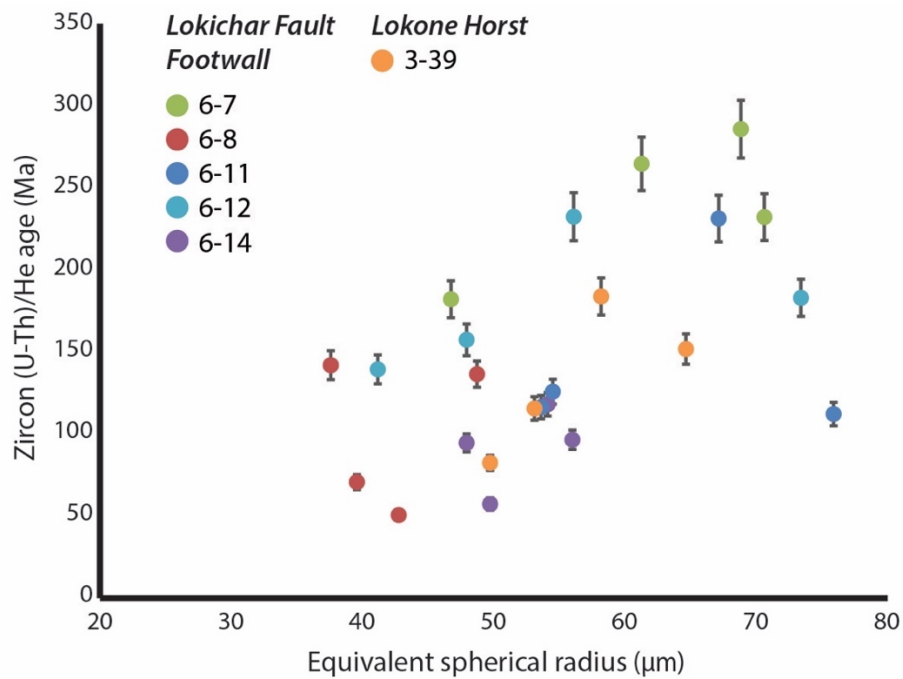


Figure SD2.1. Zircon (U-Th)/He single grain ages versus equivalent spherical radius, where equivalent spherical radius (R_s) is defined by $[3 \cdot R \cdot L] / [2 \cdot (R + L)]$, R = grain radius and L = grain length.

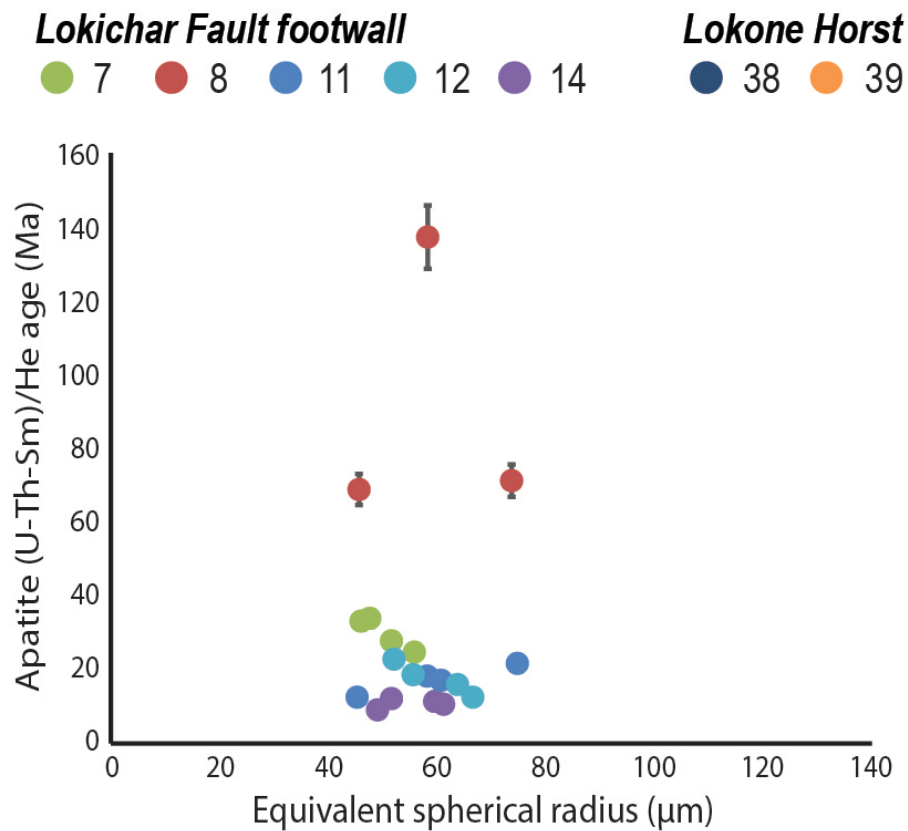


Figure SD2.2. Apatite (U-Th-Sm)/He single grain ages versus equivalent spherical radius, where equivalent spherical radius (R_s) is defined by $[3 \cdot R \cdot L] / [2 \cdot (R + L)]$, R = grain radius and L = grain length.

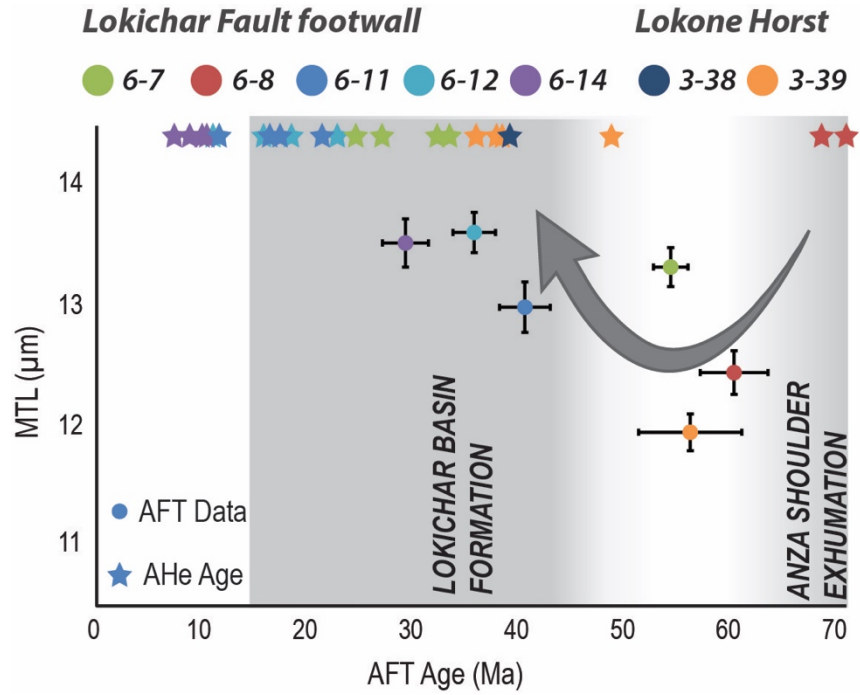


Figure SD2.3. AFT MTL as a function of age. Single grain AHe ages are shown for comparison. AFT data are inferred to form the ascending arm of a so-called ‘boomerang curve’ that is indicative of a set of samples that have experienced a multi-phase thermal history [Green et al., 1989; Gallagher and Brown, 1997]. Such AFT age-MTL relationships are characteristic of an early period of pronounced cooling, such as Late Cretaceous-early Paleogene cooling recorded throughout Kenya and southern Ethiopia related to uplift and denudation of the Anza Rift shoulders and adjacent hinterlands [e.g. Foster and Gleadow, 1996; Spiegel et al., 2007; Balestrieri et al., 2016; Boone et al., 2018], followed by a marked period of Eocene-Miocene cooling through the AFT PAZ to near surface temperatures, interpreted here to represent uplift and denudation of the Lokichar Fault footwall in response to Lokichar Basin formation. Samples in the proximal Lokichar Fault footwall (11, 12 and 14) experienced a greater degree of cooling from temperature above or within the upper PAZ as indicated by relatively young AFT ages (40 ± 2 to 30 ± 2 Ma), long MTL (13.0 - 13.6 μm) and young AHe ages (22 ± 1 to 8.3 ± 0.5 Ma). By contrast, thermochronology data from samples collected from the more distal footwall (7 and 8) or the Lokichar Basin flexural margin (Lokone Horst; 38 and 39) are consistent with thermal histories involving prolonged residence within or reheating into the AFT PAZ, respectively, followed by cooling to surface temperatures during Eocene-Miocene times. This scenario is in agreement with their older AFT ages (54 ± 2 to 60 ± 3 Ma), intermediate MTL (12.4 to 13.3 μm) and older AHe ages (137 ± 9 to 24 ± 2 Ma).

Thermal History Modelling Protocol

In order to quantify thermal histories for basement samples from the footwall of the Lokichar Fault, ZHe, AFT and AHe data were inverse modelled jointly using the modelling software *QTQt* [Gallagher, 2012], with the multi-compositional fission track annealing model of Ketcham *et al.* [2007a]. *QTQt* employs a Bayesian transdimensional Markov chain Monte Carlo (MCMC) sampling method to generate a range of acceptable thermal histories, quantified in terms of a posterior probability distribution [Gallagher *et al.*, 2009; Gallagher 2012]. Confined fission track lengths were normalised for annealing and etching anisotropy using the *c*-axis projection method [Ketcham, 2007b]. Differences in ZHe and AHe grain-size were accounted for by using an alpha ejection correction (F_T) [Farley *et al.*, 1996]. Radiation damage accumulation and annealing models were employed for modelling of AHe data [Gautheron *et al.*, 2009] and ZHe data [ZRDAAM; Guenthner *et al.*, 2013] to account for differences in radiation-modified diffusion kinetics recorded by age-eU relationships. However, these models are unable to account for all known sources of intrasample ZHe and AHe age dispersion, such as chemical zonation [Guenthner *et al.*, 2012] and apatite composition [Gautheron *et al.*, 2013]. Therefore, during modelling, *QTQt* was allowed to resample the He error for single grain ages, effectively decreasing their influence on determining thermal histories, as compared to data from the more robust AFT system [Gallagher, 2012]. For samples with fragmented AHe grains, only 1- and 2-termination grains were modelled in *QTQt* due to current limitations in the implementation of the Brown *et al.* [2013] fragmentation model.

Sample 38 and 41 did not yield confined fission track measurements and were, therefore, omitted during thermal history modelling. For all other samples, thermal history modelling was first performed using solely AFT data, as apatite fission track annealing kinetics are much better understood than He diffusion processes in zircon and apatite [Green and Duddy, 2018]. Thermal history modelling then proceeded using sample AFT and AHe data, when available, before a final modelling run was carried out that also included ZHe data. These results were then compared with one another to ensure consistency. For thermal history modelling of either AFT data only or AFT & AHe data, a temperature range of 0-140 °C and a time range of 0 Ma to twice the value of the oldest thermochronometer age was used for the time-temperature prior range. For thermal history modelling of data that included ZHe results, the prior temperature range was increased to 0-250 °C. During thermal history

modelling of individual samples, 150,000 iterations were run, with the first 50,000 discarded as burn-in [Gallagher *et al.*, 2009]. To ensure satisfactory sampling, the acceptance rates of the proposed model parameters and the posterior sampling chain were checked after each modelling run [Gallagher, 2012]. A time-temperature constraint was also applied to all models, requiring samples to be at a present day surface temperatures of 20 ± 10 °C.

The preferred thermal history model for each sample was that which was both consistent with the local geology and best predicted the observed thermochronology data. Preferred thermal histories are shown in Figure 2.3B and Figure SD2.4.

Thermal history models from basement samples from the Lokichar Basin margins record a succession of cooling and reheating periods that span Cretaceous-Recent times. The Cretaceous and, in some cases, early Paleogene portions of these models are constrained by ZHe data alone, as all AFT and AHe data were completely reset prior to basement samples cooling below the corresponding closure temperatures late Paleogene. In general, thermal history models do a poor job of predicting observed single grain ZHe ages (Figures SD2.4-6), likely resulting from the well-documented inability of the ZRDAAM He diffusion kinetic model to reproduce datasets with significant ZHe age dispersion that correspond to a wide range of zircon eU concentrations [e.g. Orme *et al.*, 2016; Mackintosh *et al.*, 2017]. The observed ages may also be influenced by problems arising from intra-grain zonation of U and Th [Anderson *et al.*, 2017]. Whatever the reason for the discrepancy between observed and predicted ZHe ages, little confidence is placed in portions of the thermal history models constrained solely by ZHe data. Instead, this study focuses on the late Paleogene-Recent portion of modelled thermal histories, which are also constrained by AFT and AHe data. In the majority of cases, thermal history models predicted AHe ages within error of observed values. However, this was not the case for sample 8, which generated thermal histories that poorly reproduced observed AHe data (Figures SD2.6B, D). The inability of current apatite radiation damage He diffusivity models to reproduce AHe ages in disperse datasets, such as sample 8, has often been reported [Green and Duddy, 2018 and references therein]. Thus, these thermal history models are excluded from further discussion.

Thermal history models for basement samples from the Lokichar Fault footwall all record late Paleogene to Miocene-Pliocene cooling that are broadly consistent with one another. The thermal history models (which included all data and no additional time-temperature

constraints) for samples collected < 3 km into the Lokichar Fault footwall (11, 12 and 14) record a pronounced Eocene (~50-40 Ma; Figure SD2.4) to Mio-Pliocene cooling period. This is consistent with their Eocene-early Oligocene AFT ages (29 ± 2 to 41 ± 2 Ma; Table SD2.1), intermediate to long MTL (13.0-13.6 μm ; Table SD2.1) and Miocene single grain AHe ages (8.3 ± 0.5 to 22 ± 1 Ma; Table SD2.3) (see Figure SD2.3). The 7 km of recorded late Paleogene-middle Miocene subsidence in the nearby Lokichar Fault hanging wall [Morley *et al.*, 1999] and the incised nature of exposed Precambrian crystalline basement rocks on the Lokichar western margin suggest that the observed Eocene to Mio-Pliocene cooling in the Lokichar Basin western margin resulted from uplift and denudation of the Lokichar Fault footwall in response to extension.

The more distal Lokichar footwall samples (7 and 8) by contrast, record less late Paleogene-Neogene thermal rejuvenation, as exhibited by their older early Paleogene AFT ages (54 ± 2 to 60 ± 3 Ma; Table DR1), moderate MTL (12.4-13.3 μm ; Table DR1) and older and more disperse AHe ages (28 ± 2 to 161 ± 10 Ma; Table DR3). Accordingly, the late Paleogene-Neogene thermal history of samples 7 and 8 are more poorly constrained by thermochronology data than their counterparts situated more proximal to the Lokichar Fault trace (11, 12 and 14). As a result, the *QTQt* software, whose Bayesian approach favours simpler thermal histories [Gallagher *et al.*, 2012], generated prolonged and monotonic time-temperature paths for samples 7 and 8 that exhibit an earlier onset (latest Cretaceous-early Paleogene) of cooling (Figures SD2.5 and SD2.6).

To test if thermochronology data from samples 7 and 8 were consistent with an Eocene onset of cooling as recorded by thermal history models from the more thermally overprinted, and thus better constrained, Lokichar Fault footwall samples (11, 12 and 14), samples 7 and 8 were remodelled with an additional Eocene (50-40 Ma) time-temperature constraint (Figures SD2.5E-H and SD2.6E-H). The temperature range of the added constraint for sample 7 was 70-100 °C, consistent with thermal history models generated without the Eocene constraint (Figure S2.5A-D). The temperature range for the Eocene constraint was reduced to 60-90 °C for sample 8, whose older AFT and AHe ages and shorter MTL (Tables SD2.1-3) are consistent with cooling from lower temperatures. In all cases, thermal history models generated with the additional Eocene time-temperature constraint for samples 7 and 8 are broadly consistent with those without (Figures SD2.5 and SD2.6), indicating the addition of such a constraint is coherent with the thermochronology data. Moreover, the preferred

thermal history models for samples 7 and 8 which incorporate the added Eocene constraint (Figures 2.3B and SD2.4-6) yield best-fit time-temperature paths that exhibit an Eocene onset of marked cooling, consistent with thermal history modelling results from samples 11, 12 and 14.

Sample 39 was the only basement samples from the Lokichar flexural margin (Lokone Horst) to generate thermal history models (Figures 2.3B and SD2.4). However, due to AFT ages from nearby samples 38 and 41 being within 1σ of the AFT age of sample 39, thermal history modelling results for sample 39 are considered representative of the Lokone Horst. After a period of Paleocene-early Eocene cooling to near surface temperatures, the best-fit time-temperature reconstruction for sample 39 records approximately 23 °C of cooling between the middle Eocene (~45 Ma) and late Miocene (~10 Ma). This is followed by a period of late Miocene-Recent rapid cooling to near-surface temperatures. The extensive exposure of Oligocene-upper Miocene syn-rift sequences throughout the Lokichar Basin flexural margin and onlapping the western Lokone Horst and occurrence of Oligocene strata in the nearby Loperot-1 exploration well (Figure 2.2A) [2950 m; *Morley et al.*, 1999] suggest that the observed reheating period is a result of burial and subsidence during the formation of the Lokichar Basin. While thermal history modelling implies an earlier Eocene (~45 Ma) onset of syn-rift deposition on the Lokichar Basin flexural margin than outcrop may suggest, the presence of a significant thickness of undrilled, syn-rift strata beneath the bottom of the Loperot-1 exploration well [*Morley et al.*, 1999] is consistent with such a scenario. Moreover, thermal history modelling from the western Lokichar Basin margin are also consistent with an Eocene (50-40 Ma) onset of basin formation.

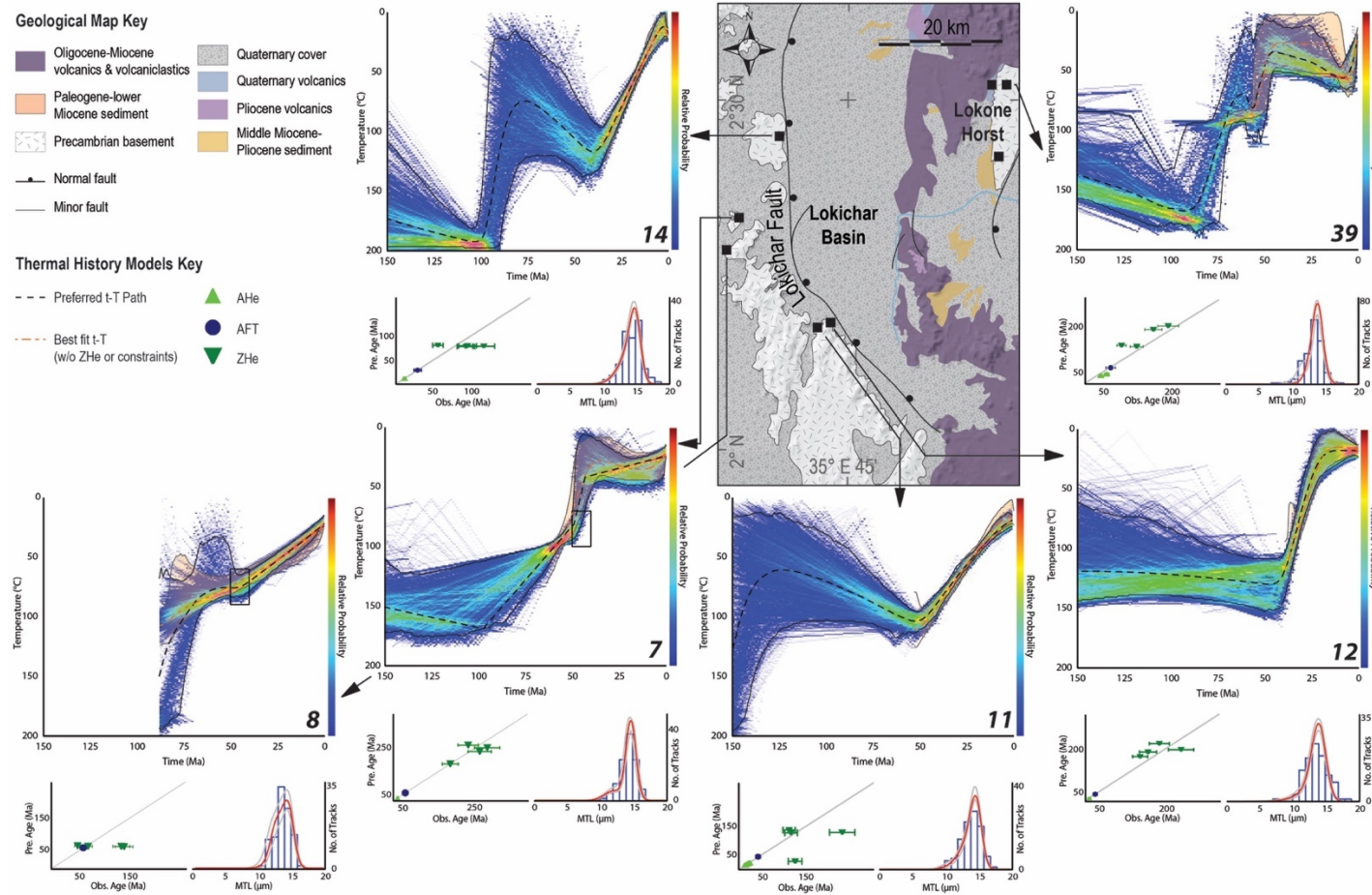


Figure DR4. Preferred individual time-temperature histories of basement samples from the Lokichar Basin margins, determined using the QTQt software [Gallagher, 2012]. Preferred thermal history models include all available data, with the exception of sample 8 where AHe data were omitted. See Thermal History Modelling Protocol for detailed modelling procedure and criteria for selection of preferred thermal histories. Black boxes represent time-temperature constraints (samples 7 and 8). The colour map represents the posterior-probability distribution of accepted time-temperature paths, with warmer colours indicating higher probabilities. The preferred thermal histories (black dashed lines) are the best fit time-temperature paths [weighted mean of the posterior distribution; see Gallagher, 2012 for discussion], except for that of sample 7, which is the maximum posterior model (see Figure SD2.5 caption). Black solid lines represent the 95% confidence interval limits. The data fit for each preferred thermal history model is illustrated by panel inserts showing observed versus predicted thermochronology ages (note that He ages are displayed with resampling errors) and of the observed confined track distributions overlain by the predicted distributions (red curve) and the 95% confidence interval (grey curves) of the predicted distributions. For comparison, the corresponding best fit t-T path (orange dashed line) and 95% confidence interval (orange envelope) of thermal history models generated for each sample without ZHe data or additional

time-temperature constraints (if applied) are also illustrated. Geological map, after BEICIP [1987], showing sample localities and underlain by a digital elevation model [NASA, 2000].

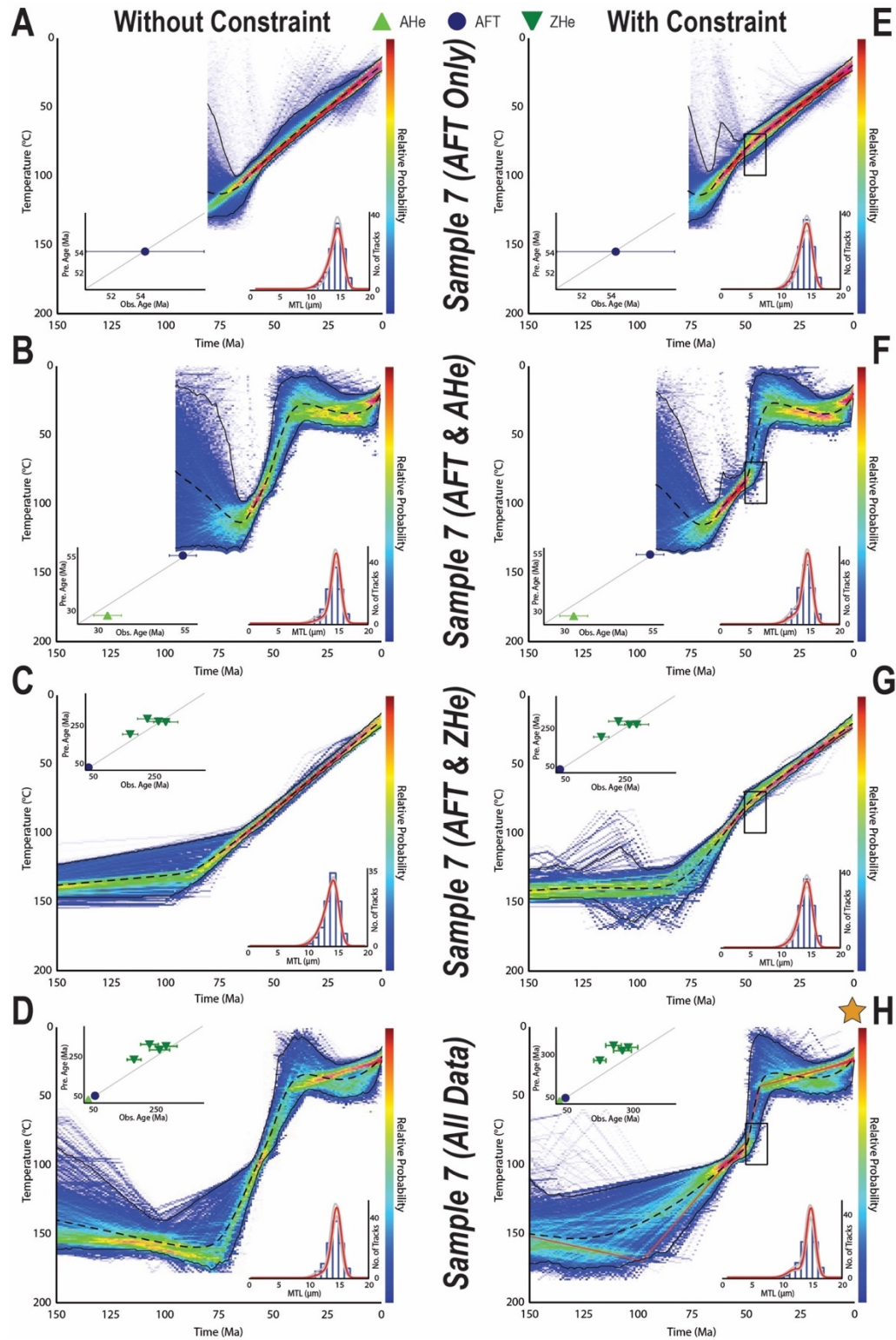


Figure SD2.5. Influence of Eocene constraint on thermal history modelling of sample 7. See Thermal History Modelling Protocol for detailed modelling procedure and criteria for selection of preferred thermal histories. Symbology as for Fig. DR4. The preferred thermal history model for sample 7 (star) was generated using all available thermochronometer data and the Eocene constraint (**H**). In contrast to the sample 7 thermal history model using AFT data only (**A** and **E**) and the preferred thermal histories of other Lokichar Fault footwall samples (Figure SD2.4), the best-fit time-temperature path for **H** exhibits rapid Eocene-Oligocene cooling followed by relative thermal stability until the present day. However, the 95% confidence interval (black envelope) and time-temperature paths of highest probability (the maximum posterior model; red line) are consistent with a thermal history involving a rapid onset of Eocene cooling followed by continued, yet reduced, Oligocene-Recent cooling that is compatible with preferred time-temperature reconstructions from other samples (Fig. DR4) and the local geology. Thus, the maximum

posterior thermal history of **H** is selected as the preferred thermal history model for sample 7 (Figures 2.3B and SD2.4).

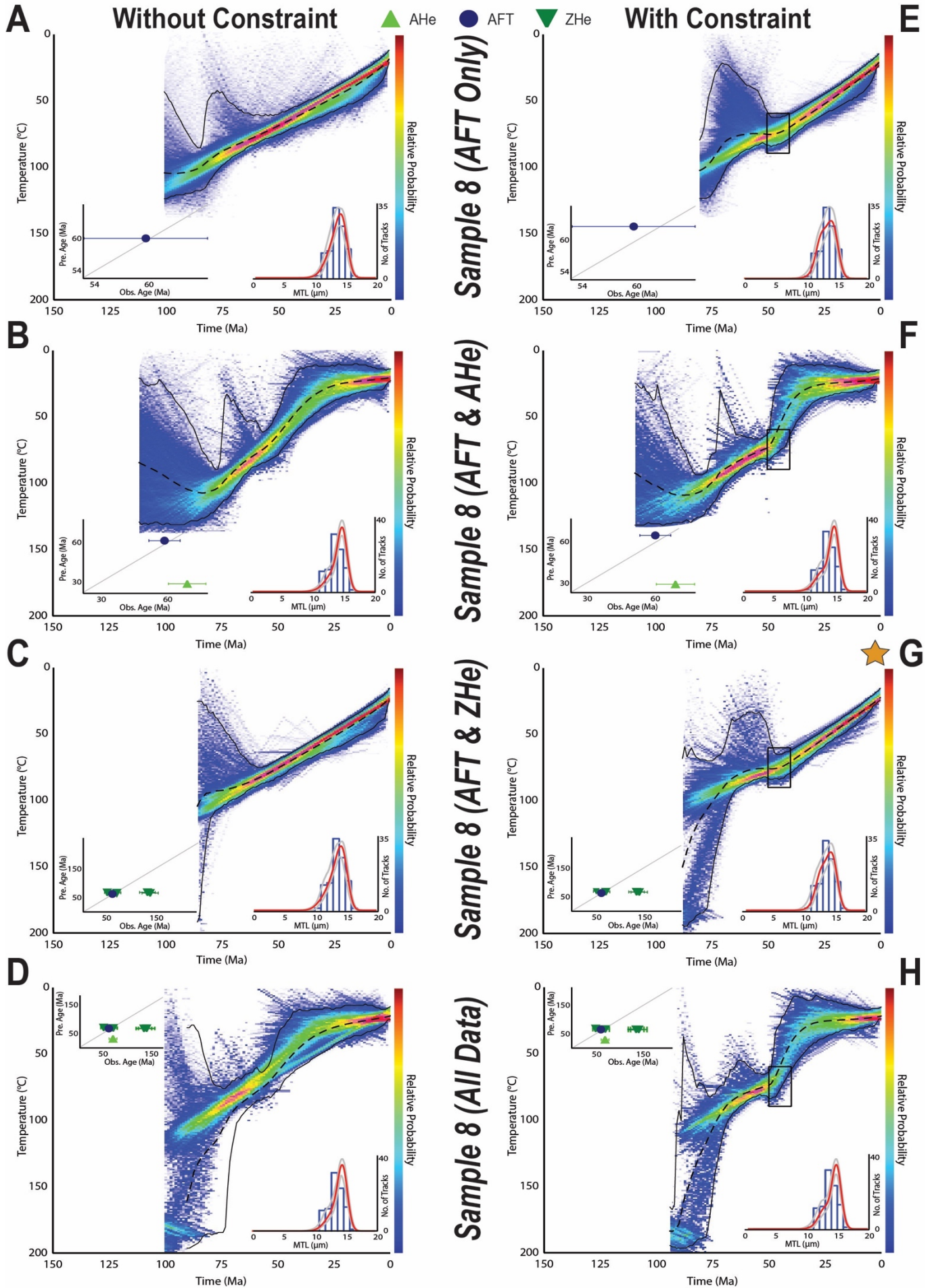


Figure SD2.6. *Influence of Eocene constraint on thermal history modelling of sample 8. See Thermal History Modelling Protocol for detailed modelling procedure and criteria for selection of preferred thermal histories. Symbology as for Figure SD2.4. The preferred thermal history model for sample 8 (star) was generated using AFT and ZHe data and the Eocene constraint (G).*

Supplementary Data References

- Anderson, A. J., Hodges, K. V., & van Soest, M. C. (2017). Empirical constraints on the effects of radiation damage on helium diffusion in zircon. *Geochimica et Cosmochimica Acta*, 218, 308–322. <https://doi.org/10.1016/j.gca.2017.09.006>
- Balestrieri, M. L., Bonini, M., Corti, G., Sani, F., & Philippon, M. (2016). A refinement of the chronology of rift-related faulting in the Broadly Rifted Zone, southern Ethiopia, through apatite fission-track analysis. *Tectonophysics*, 671, 42–55. <https://doi.org/10.1016/j.tecto.2016.01.012>
- Barbarand, J., Carter, A., Wood, I., & Hurford, T. (2003). Compositional and structural control of fission-track annealing in apatite. *Chemical Geology*, 198(1–2), 107–137. [https://doi.org/10.1016/S0009-2541\(02\)00424-2](https://doi.org/10.1016/S0009-2541(02)00424-2)
- BEICIP. (1987). Geological Map of Kenya. Nairobi: Ministry of Energy and Regional Development.
- Beucher, R., Brown, R. W., Roper, S., Stuart, F., & Persano, C. (2013). Natural age dispersion arising from the analysis of broken crystals: Part II. Practical application to apatite (U–Th)/He thermochronometry. *Geochimica et Cosmochimica Acta*, 120, 395–416. <https://doi.org/10.1016/j.gca.2013.05.042>
- Boone, S. C., Seiler, C., Reid, A. J., Kohn, B., & Gleadow, A. (2016). An Upper Cretaceous paleo-aquifer system in the Eromanga Basin of the central Gawler Craton, South Australia: evidence from apatite fission track thermochronology. *Australian Journal of Earth Sciences*, 63(3), 315–331. <https://doi.org/10.1080/08120099.2016.1199050>
- Boone, S. C., Seiler, C., Kohn, B. P., Gleadow, A. J. W., Foster, D. A., & Chung, L. (2018). Influence of Rift Superposition on Lithospheric Response to East African Rift System Extension: Lapur Range, Turkana, Kenya. *Tectonics*, 37, 1–26.
- Brown, R. W., Beucher, R., Roper, S., Persano, C., Stuart, F., & Fitzgerald, P. (2013). Natural age dispersion arising from the analysis of broken crystals. Part I: Theoretical basis and implications for the apatite (U–Th)/He thermochronometer. *Geochimica et Cosmochimica Acta*, 122(120), 478–497. <https://doi.org/10.1016/j.gca.2013.05.041>
- Danišik, M., Sachsenhofer, R. F., Privalov, V. A., Panova, E. A., Frisch, W., & Spiegel, C. (2008). Low-temperature thermal evolution of the Azov Massif (Ukrainian Shield–Ukraine) — Implications for interpreting (U–Th)/He and fission track ages from cratons. *Tectonophysics*, 456, 171–179. <https://doi.org/10.1016/j.tecto.2008.04.022>
- Farley, K. A., & Stockli, D. F. (2002). (U–Th)/He Dating of Phosphates: Apatite, Monazite, and Xenotime. *Reviews in Mineralogy and Geochemistry*, 48(1), 559–577. <https://doi.org/10.2138/rmg.2002.48.15>
- Farley, K. A., Wolf, R. A., & Silver, L. T. (1996). The effects of long alpha-stopping distances on (U–Th)/He ages. *Geochimica et Cosmochimica Acta*, 60(21), 4223–4229. [https://doi.org/10.1016/S0016-7037\(96\)00193-7](https://doi.org/10.1016/S0016-7037(96)00193-7)
- Foster, D. A., & Gleadow, J. W. (1996). Structural framework and denudation history of the flanks of the Kenya and Anza Rifts, East Africa. *Tectonics*, 15(2), 258–271.
- Galbraith, R. F. (2005). *Statistics for fission track analysis*. Boca Raton: Taylor & Francis Group.
- Gallagher, K. (2012). Transdimensional inverse thermal history modeling for quantitative thermochronology. *Journal of Geophysical Research: Solid Earth*, 117. <https://doi.org/10.1029/2011JB008825>
- Gallagher, K., & Brown, R. (1997). The onshore record of passive margin evolution. *Journal of the Geological Society, London*, 154, 451–457. <https://doi.org/10.1144/gsjgs.154.3.0451>
- Gallagher, K., Charvin, K., Nielsen, S., Sambridge, M., & Stephenson, J. (2009). Markov chain Monte Carlo (MCMC) sampling methods to determine optimal models, model

- resolution and model choice for Earth Science problems. *Marine and Petroleum Geology*, 26, 525–535. <https://doi.org/10.1016/j.marpetgeo.2009.01.003>
- Gautheron, C., Barbarand, J., Ketcham, R. A., Tassan-Got, L., van der Beek, P., Pagel, M., ... Fialin, M. (2013). Chemical influence on α -recoil damage annealing in apatite: Implications for (U–Th)/He dating. *Chemical Geology*, 351, 257–267. <https://doi.org/10.1016/j.chemgeo.2013.05.027>
- Gautheron, C., Tassan-Got, L., Barbarand, J., & Pagel, M. (2009). Effect of alpha-damage annealing on apatite (U–Th)/He thermochronology. *Chemical Geology*, 266, 157–170. <https://doi.org/10.1016/j.chemgeo.2009.06.001>
- Gleadow, A. J. W., & Duddy, I. R. (1981). A Natural Long-Term Track Annealing Experiment for Apatite. *Nuclear Tracks*, 5(1–2), 169–174.
- Gleadow, A., Harrison, M., Kohn, B., Lugo-zazueta, R., & Phillips, D. (2015). The Fish Canyon Tuff: A new look at an old low-temperature thermochronology standard. *Earth and Planetary Science Letters*, 424, 95–108. <https://doi.org/10.1016/j.epsl.2015.05.003>
- Gleadow, A. J. W., Gleadow, S. J., Belton, D. X., Kohn, B. P., Krochmal, M. S., & Brown, R. W. (2009). Coincidence mapping – a key strategy for the automatic counting of fission tracks in natural minerals. *Geological Society of London Special Publications, Thermochronological Methods: From Palaeotemperature Constraints to Landscape Evolution Models.*, 324, 25–36. <https://doi.org/10.1144/SP324.2>
- Green, P., & Duddy, I. (2018). Apatite (U–Th–Sm)/He thermochronology on the wrong side of the tracks. *Chemical Geology*.
- Green, P. F., Duddy, I. R., Gleadow, A. J. W., & Lovering, J. F. (1989). Apatite Fission-Track Analysis as a Paleotemperature Indicator for Hydrocarbon Exploration. In N. D. Naeser & T. H. McCulloh (Eds.), *Thermal History of Sedimentary Basins, Methods and Case Histories* (pp. 181–196). New York: Springer-Verlag.
- Guenther, W. R., Reiners, P. W., Ketcham, R. A., Nasdala, L., & Giester, G. (2013). Helium diffusion in natural zircon: radiation damage, anisotropy, and the interpretation of zircon (U–Th)/He thermochronology. *American Journal of Science*, 313(3), 145–198. <https://doi.org/10.2475/03.2013.01>
- Guenther, W. R., Reiners, P. W., & Tian, Y. (2014). Interpreting date-eU correlations in zircon (U–Th)/He datasets: A case study from the Longmen Shan, China. *Earth and Planetary Science Letters*, 403(OCTOBER), 328–339. <https://doi.org/10.1016/j.epsl.2014.06.050>
- Hasebe, N., Barbarand, J., Jarvis, K., Carter, A., & Hurford, A. J. (2004). Apatite fission-track chronometry using laser ablation ICP-MS. *Chemical Geology*, 207(3), 134–145.
- Hourigan, J. K., Reiners, P. W., & Brandon, M. T. (2005). U–Th zonation-dependent alpha-ejection in (U–Th)/He chronometry. *Geochimica et Cosmochimica Acta*, 69(13), 3349–3365. <https://doi.org/10.1016/j.gca.2005.01.024>
- Ketcham, R. A. (2005). Forward and Inverse Modeling of Low-Temperature Thermochronometry Data. *Reviews in Mineralogy and Geochemistry*, 58, 275–314. <https://doi.org/10.2138/rmg.2005.58.11>
- Ketcham, R. A., Carter, A., Donelick, R. A., Barbarand, J., & Hurford, A. J. (2007a). Improved modeling of fission-track annealing in apatite. *American Mineralogist*, 92, 799–810. <https://doi.org/10.2138/am.2007.2281>
- Ketcham, R. A., Carter, A., Donelick, R. A., Barbarand, J., & Hurford, A. J. (2007b). Improved measurement of fission-track annealing in apatite using c-axis projection. *American Mineralogist*, 92(5–6), 789–798. <https://doi.org/10.2138/am.2007.2280>
- Laslett, G. M., Green, P. F., Duddy, I. R., & Gleadow, A. J. W. (1987). Thermal annealing of fission tracks in apatite 2. A quantitative analysis. *Chemical Geology (Isotope Geoscience Section)*, 65, 1–13. [https://doi.org/10.1016/0168-9622\(87\)90057-1](https://doi.org/10.1016/0168-9622(87)90057-1)

- Mackintosh, V., Kohn, B., Gleadow, A., & Tian, Y. (2017). Phanerozoic Morphotectonic Evolution of the Zimbabwe Craton: Unexpected Outcomes From a Multiple Low-Temperature Thermochronology Study. *Tectonics*, 36, 1–24. <https://doi.org/10.1002/2017TC004703>
- McDowell, F. W., McIntosh, W. C., & Farley, K. A. (2005). A precise ^{40}Ar - ^{39}Ar reference age for the Durango apatite (U-Th)/He and fission track dating standard. *Chemical Geology*, 214(3), 249–263.
- Morley, C. K., Stone, D. M., Harper, R. M., & Wigger, S. T. (1999). Geology and Geophysics of the Western Turkana Basins, Kenya. In C. K. Morley (Ed.), *Geoscience of Rift Systems—Evolution of East Africa: AAPG Studies in Geology No. 44* (pp. 19–54).
- NASA. (2000). Shuttle Radar Topography Mission (SRTM) 1 Arc-Second Global. Sioux Falls, SD: Land Processes Distributed Active Archive Center. Retrieved from <http://lpdaac.usgs.gov>
- Orme, D. A., Guenthner, W. R., Laskowski, A. K., & Reiners, P. W. (2016). Long-term tectonothermal history of Laramide basement from zircon U-Pb/He age-eU correlations. *Earth and Planetary Science Letters*, 453. <https://doi.org/10.1016/j.epsl.2016.07.046>
- Paton, C., Hellstrom, J., Paul, B., Woodhead, J., & Hergt, J. (2011). Iolite: Freeware for the visualisation and processing of mass spectrometric data. *Journal of Analytical Atomic Spectrometry*, 26, 2508–2518. <https://doi.org/10.1039/c1ja10172b>
- Reiners, P. W. (2005). Zircon (U-Th)/He Thermochronometry. *Reviews in Mineralogy and Geochemistry*, 58, 151–179. <https://doi.org/10.2138/rmg.2005.58.6>
- Reiners, P. W., & Farley, K. A. (2001). Influence of crystal size on apatite (U-Th)/He thermochronology: an example from the Bighorn Mountains, Wyoming. *Earth and Planetary Science Letters*, 188, 413–420.
- Reiners, P. W., Farley, K. A., & Hickey, H. J. (2002). He diffusion and (U-Th)/He thermochronometry of zircon: Initial results from Fish Canyon Tuff and Gold Butte. *Tectonophysics*, 349(1–4), 297–308. [https://doi.org/10.1016/S0040-1951\(02\)00058-6](https://doi.org/10.1016/S0040-1951(02)00058-6)
- Seiler, C., Kohn, B., & Gleadow, A. (2014). Apatite fission track analysis by LA-ICP-MS: An evaluation of the absolute dating approach. In *International Conference on Thermochronology*.
- Spiegel, C., Kohn, B. P., Belton, D. X., & Gleadow, A. J. W. (2007). Morphotectonic evolution of the central Kenya rift flanks: Implications for late Cenozoic environmental change in East Africa. *Geology*, 35(5), 427–430. <https://doi.org/10.1130/G23108A.1>
- Wagner, G., & Van den Haute, P. (1992). *Fission-Track Dating*. Dordrecht: Kluwer Academic Publishers.
- Wildman, M., Brown, R., Beucher, R., Persano, C., Stuart, F., Gallagher, K., ... Carter, A. (2016). The chronology and tectonic style of landscape evolution along the elevated Atlantic continental margin of South Africa resolved by joint apatite fission track and (U-Th-Sm)/He thermochronology. *Tectonics*, 35(3), 511–545. <https://doi.org/10.1002/2015TC004042>

Tectono-thermal evolution of a long-lived segment of the East African Rift System: Thermochronological insights from the North Lokichar Basin, Turkana, Kenya

Abstract

The Turkana Depression is a structurally complex and long-lived segment of the East African Rift System (EARS), with associated magmatism and strain nucleating there in the late Paleogene. The anomalously wide, ~N-S rift zone defines the topographic lowlands separating the Ethiopian and East African Domes. The atypical architecture and morphology of the Turkana Depression has often been attributed to the influence of an oblique, pre-existing lithospheric heterogeneity speculated to result from earlier Cretaceous-early Paleogene Anza-South Sudan rifting. However, this hypothesized period of earlier rifting is poorly constrained due to the obscuring effects of extensive Cenozoic rifting and volcanism. Similarly, the extent and timing of basin formation during the initial stages of EARS extension in Turkana is not well understood. Seismic reflection studies in Turkana have revealed the presence of older, possibly late Paleogene sub-basins, predating the Neogene onset of major faulting elsewhere in the EARS. One example, the Lothidok Basin, has previously been imaged beneath the late Miocene-Pliocene North Lokichar Basin. Its age, however, is unconstrained due to a lack of well controls, geochronological constraints and outcrop of its basal unit. Here, we present a multiple low-temperature thermochronometer [apatite fission track, apatite (U-Th-Sm)/He and zircon (U-Th)/He] study performed on Precambrian basement samples from the western margin of the overlying North Lokichar Basin. Thermal history modelling reveals a polyphase Late Cretaceous-Recent tectono-thermal evolution providing new insights into pre-EARS tectonism in Turkana and subsequent, late Paleogene ~E-W extension. Pronounced Late Cretaceous-Paleogene denudational cooling challenges the theorized linkage of the Anza-South Sudan Rifts in Turkana, perhaps suggesting later Paleogene tectonism played a more critical role in modifying the lithosphere. Subsequent Oligocene-early Miocene reheating is interpreted as resulting from burial beneath ~200-800 m of overburden, accordant with the proposed formation of the Lothidok Basin and/or coeval emplacement of thick lava flows in the region.

3.1. Introduction

Defining the topographically subdued lowlands separating the Ethiopian and East African Domes, the Turkana Depression marks a long-lived and complex segment of the East African Rift System (EARS). Here, the EARS is comprised of up to four adjacent, ~N-S extensional basins, oblique to the northwest trending, failed Cretaceous-early Paleogene Anza and South Sudanese Rifts Systems to the east and west (Figure 3.1) [Morley *et al.*, 1992; 1999b]. As the EARS traverses the Turkana Depression, the zone of rifting becomes considerably wider (up to 300 km) than the narrow (~80 km) and well expressed rift morphology exhibited by the Ethiopian and Kenyan Rifts to the north and south [Ebinger *et al.*, 2000; Corti, 2009]. This diffusion of extensional strain coincides with the transition to more attenuated crust; crustal thicknesses in Turkana are as low as ~20 km, compared to ~30 and ~35 km beneath the Ethiopian and Kenyan Rifts, respectively [Benoit *et al.*, 2006; Mariita and Keller, 2007; Keranen and Klemperer, 2008]. The anomalous architecture and structural evolution of the Turkana Rift are thought to have been largely controlled by pre-existing rheological heterogeneities resulting from an earlier, but poorly understood period of Cretaceous-early Paleogene rifting [Ebinger *et al.*, 2000; Vétel *et al.*, 2005; Benoit *et al.*, 2006; Brune *et al.*, 2017].

East African rifting nucleated early in the Turkana Depression, marked by intensive flood volcanism in southern Ethiopia and northern Turkana ca. 45 Ma. While the period of tectonism traditionally associated with the EARS is Neogene to Recent, seismic reflection data and remote sensing analysis have suggested that EARS-related extensional basin formation began earlier in the Turkana area, most likely during the late Paleogene [e.g. Morley *et al.*, 1992, Wescott *et al.*, 1999]. One example is the North Lokichar Basin, an ~N-S half graben on the west side of Lake Turkana (Figures 3.1 and 3.2). Bounded by the east-dipping Lokichar Fault on its western margin, the North Lokichar Basin contains an upper succession of westward thickening late Miocene-Pliocene siliciclastics [Boschetto *et al.*, 1992; Morley *et al.*, 1999b], consistent with hanging wall sedimentation in an asymmetric half-graben bound by the Lokichar Fault. Yet, seismic reflection data reveal an eastward thickening sequence in the lower part of the basin, which is estimated to be late Paleogene-middle Miocene in age, and referred to as the Lothidok Basin [Morley, 1999b; Morley *et al.*,

1999b]. The basal portion of the syn-rift Lothidok Basin succession, however, is not exposed and no drill core data exists, so that the onset of basin formation is unconstrained.

Low-temperature thermochronology provides an alternative approach to investigate landscape evolution and basin formation by dating periods of cooling and heating as proxies for rift-related erosion and burial processes [*Fitzgerald, 1992; van der Beek et al., 1994; Ehlers et al., 2001; Braun, 2005; Stockli, 2005*]. Such an approach has been used extensively throughout the EARS to determine the age of extensional basin formation and constrain how rifting propagated over time [e.g. *van der Beek et al., 1998; Bauer et al., 2013; Mortimer et al., 2016*].

Here, we present apatite fission track (AFT), apatite (U-Th-Sm)/He (AHe) and zircon (U-Th)/He (ZHe) analyses from 10 basement samples collected from the North Lokichar Basin margin. Inverse thermal history modelling of these data reveals a complex Cretaceous-Recent thermal evolution, which provides important new insights into Cretaceous-Paleogene tectonism in southern Turkana and the early stages of EARS development during late Paleogene time. These data are complemented by detrital AFT, AHe and ZHe analyses from a syn-rift conglomerate sample in the Late Miocene-Pliocene Turkwel Beds, constraining maximum North Lokichar Basin subsidence.

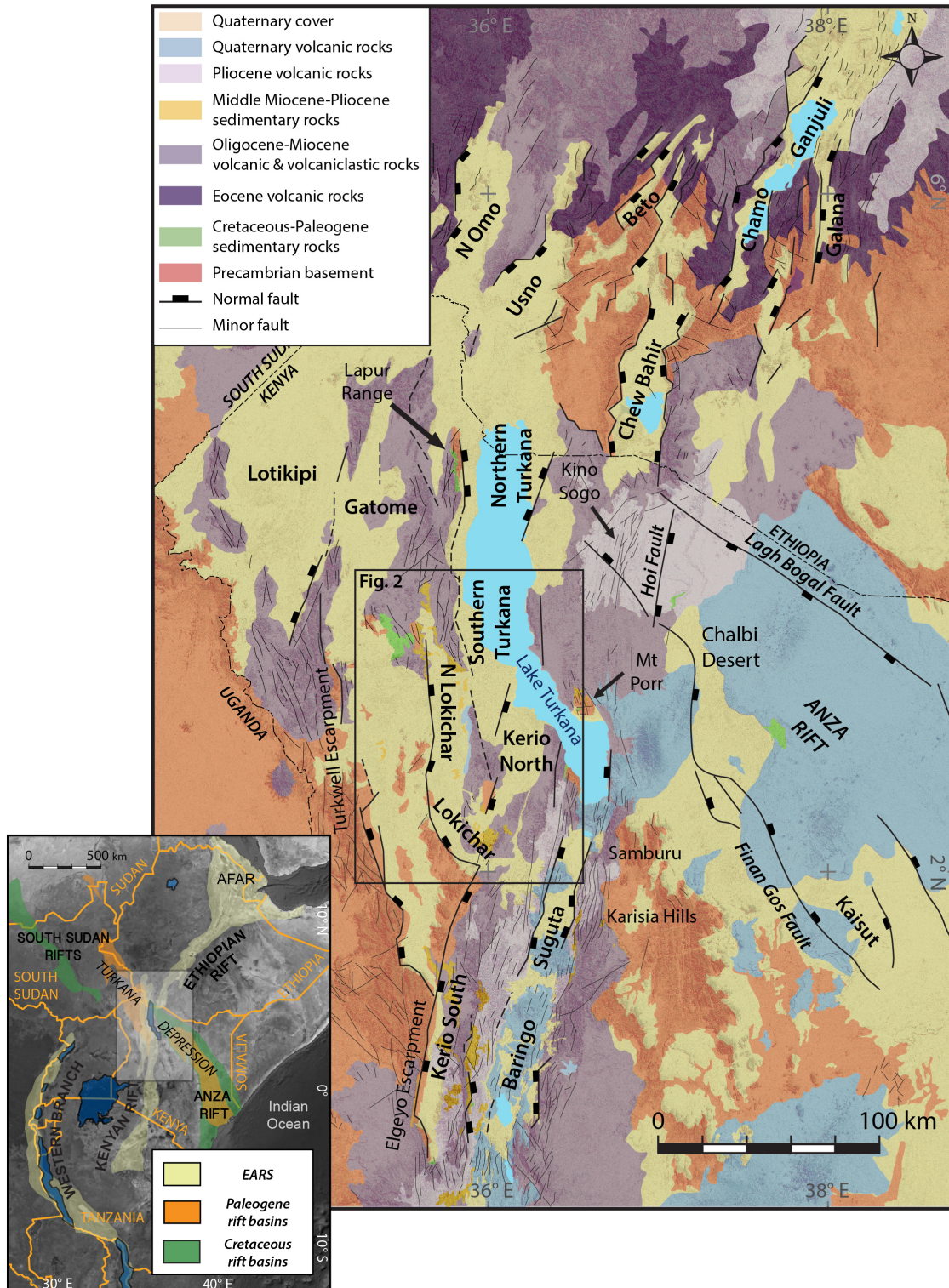


Figure 3.1. Simplified geological map of the Turkana Depression, underlain by a digital elevation model [NASA, 2000], illustrating the location of Paleogene-Pliocene extensional basins, localities referred to in the text and the extent of Figure 3.2, adapted from Kazmin [1972], GMRB-Sudan [1981], Davidson [1983], BEICIP [1987], Morley et al. [1999a], Vétel and Le Gall [2006], Lehto et al. [2014], Philippon et al. [2014], Balestrieri et al. [2016]. Inset displays regional map of East Africa, showing the location of the Turkana Depression in relation to the distribution of Cretaceous, Paleogene and Neogene-Recent (EARS) rift basins, after Hendrie et al. [1994].

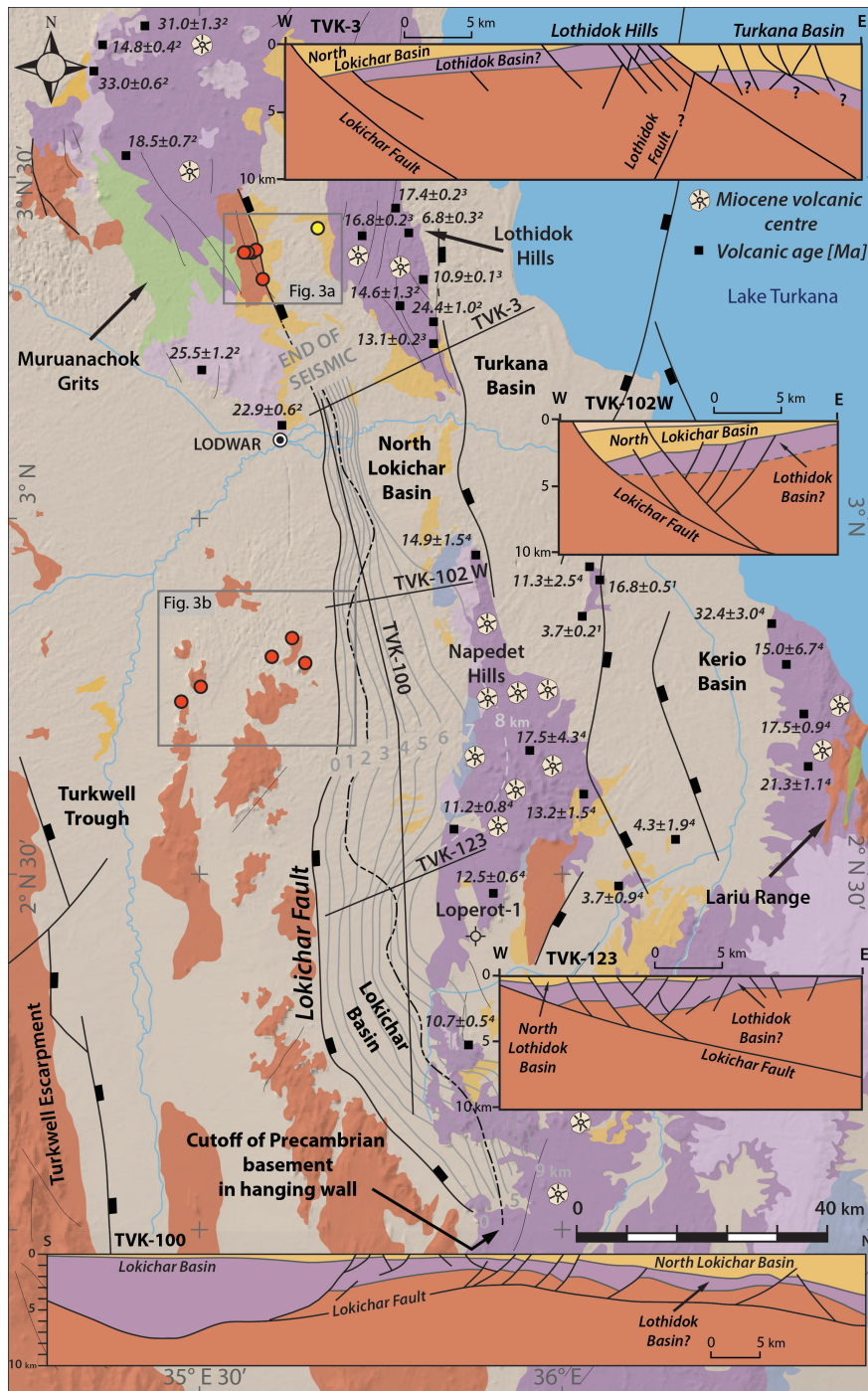


Figure 3.2. Simplified geological map of the North Lokichar Basin region, after BEICIP [1987], underlain by a digital elevation model [NASA, 2000]. Map legend as for Figure 3.1. The location of Miocene volcanic centres [after Morley, 1999b; Vétel and Le Gall, 2006] and volcanic rock ages are also displayed. Source of geochronological ages of volcanic rocks (black boxes) are indicated by superscript: (1) Patterson [1970], (2) Zanettin et al. [1983], (3) Boschetto et al. [1992], and (4) Morley et al. [1992, 1999b]. Location of seismic lines TVK-3, TVK-102W, TVK-123 and TVK-100 shown and the corresponding interpretations, adapted from Morley et al. [1999b]. The Lokichar Fault plane is illustrated as a series of depth contours in kilometres as determined from seismic reflection data [Morley 1999b]. Red (basement rocks) and yellow (sedimentary rocks) circles indicate low-temperature thermochronology sample locations. The extent of Figures 3.3a and 3.3b are shown in grey boxes.

3.2. Tectonic and Geological Setting

3.2.1 Cretaceous-early Paleogene rifting

The Anza Rift extends about 600 km NW from the Kenyan coast to Lake Turkana (Figure 1). Bound between the Lagh Bogal and Finan Gos fault systems, the asymmetric Anza graben hosts highly faulted depocentres up to 7 km thick [Bosworth and Morley, 1994; Morley *et al.*, 1999a; Vétel and Le Gall, 2006]. These consist of Lower Cretaceous-Cenozoic lacustrine shales and fluvio-deltaic sandstones, with an overall trend of increasing thickness of Late Cretaceous and Palaeogene section towards the southeast [Bosworth and Morley, 1994; Morley *et al.*, 1999a]. In the Chalbi desert, the wells drilled in the Anza Graben exhibit an unconformity between the Late Cretaceous section and Miocene sediments (Morley *et al.*, 1999a). Some extensional reactivation occurred during the Miocene to accommodate deposition of over 2000 m of section in some areas (e.g. Bellatrix-1 well, Morley *et al.*, 1999a). Vitrinite reflectance data from exploration wells, shows a jump in values between the Miocene and Cretaceous section, suggesting about 2,400 m of Late Cretaceous-Palaeogene section may have been deposited and eroded prior to the Miocene [Morley *et al.*, 1999a]. Estimated bulk extension across the Anza is 60 km [Dindi, 1994]. To the west, the similarly NW-SE trending Abu Gabra-Muglad, Melut and Bahr el Arab basins represent a series of parallel Cretaceous-Paleogene basins, known collectively as the South Sudan Rifts, that narrow as they approach the South Sudanese-Kenyan border [Browne and Fairhead, 1983]. The geometry of this termination however, remains unclear due to a general data gap in eastern South Sudan and the obscuring effect of subsequent Paleogene-Recent tectonics and volcanism in northern Kenya [Guiraud and Bosworth, 1997]. Despite this uncertainty, the compatible geometries, coeval evolution and similar gravity signals of the South Sudan and Anza rifts suggest a possible link between the two systems in Turkana [Schull, 1988; Ibrahim *et al.*, 1991; Bosworth, 1992; Ebinger and Ibrahim, 1994; Hendrie *et al.*, 1994]. The kinematically linked global Gondwana restorations of Reeves *et al.* [2016] illustrate how through-going motions along the Central African Rift System are needed to accommodate spreading in both the Atlantic and Indian oceans.

However, others argue that the Anza and South Sudan Rifts were initially disconnected [Vétel and Le Gall, 2006], instead suggesting a termination of the Anza Graben at a syn-rift faulted boundary along the Hoi Fault (Figure 3.1). This is consistent with the scarcity of Cretaceous-

lower Paleogene deposits in the Turkana region. In the Lapur Range of northern Turkana, exposure of Upper Cretaceous-Eocene Lapur Sandstone provides the sole confirmed example of syn-Anza Rift deposition in the region [Tiercelin *et al.*, 2012b]. The Lapur Sandstone unconformably overlies Precambrian basement and interfingers with the overlying Turkana Volcanics [37-17 Ma; Zanettin *et al.*, 1983; Morley *et al.*, 1992]. However, their connection to Anza rifting remains dubious due to their limited thickness (500 m) and lack of rift-related faulting [Morley *et al.*, 1992]. A poorly understood syn-rift series revealed on seismic data underlying Oligocene-Miocene volcanics of the Lotikipi and Gatome area may represent a Cretaceous [Vétel and Le Gall, 2006] or Paleogene [Wescott *et al.*, 1999] extension of the South Sudan rift trends. Scattered sequences of undated arkosic grits in the Muruanachok Hills, Lariu Range, and Mt Porr areas (Figures 3.1 and 3.2) may also represent Cretaceous-lower Paleogene sedimentation. Unlike the later EARS, the Cretaceous-early Paleogene Anza-South Sudan rift systems were largely amagmatic [Guiraud and Bosworth, 1997], although some intrusions of Cretaceous age are present [Morley *et al.*, 1999a]. Thus, the stratigraphic position of these scattered sandstone sequences, unconformably overlying Precambrian basement and underlying Miocene or Oligocene volcanic flows, suggests that they may be Cretaceous?-Paleogene in age [Walsh and Dodson, 1969; Wescott *et al.*, 1993; Tiercelin *et al.*, 2004]. However, the lack of dateable volcanic-clastic material or biostratigraphic markers has prevented these units from being dated precisely.

3.2.2 Late Paleogene-Recent East African Rift System (EARS)

In the Turkana segment of the EARS, earliest volcanism preceded deposition in half-graben basins beginning with the emplacement of tholeiitic flood basalts in southern Ethiopia and northern Turkana between 45 and 33 Ma [Bellieni *et al.*, 1981; Zanettin *et al.*, 1983; Ebinger *et al.*, 1993; Wescott *et al.*, 1999; George and Rogers, 2002]. The temporal lag between the Eocene onset of extensive volcanism in northern Turkana and southern Ethiopia and significant surface-rifting advocates melt generation via upwelling of a Paleogene asthenospheric thermal anomaly [Hendrie *et al.*, 1994; Morley, 1994; Ebinger *et al.*, 2000; Pik *et al.*, 2008]. In general, magmatic activity within Turkana migrated southwards and eastwards over time. A significant phase of volcanism affected the area north of Lodwar between 26-19 Ma [Zanettin *et al.*, 1983; Morley *et al.*, 1999b]. Between 17 and 13 Ma, the Napudet Hills and Lariu Range (Figure 3.2) were densely intruded by N50° to N170° trending dikes [Morley *et al.*, 1992; Wescott *et al.*, 1993; Tiercelin *et al.*, 2004], coeval with a major volcanic phase in south-eastern and eastern Turkana and the central Kenyan Rift [Walsh and

Dodson, 1969; *BEICIP*, 1987]. By the Pliocene, magmatism, while widely distributed, became focussed on the eastern side of Lake Turkana, where fissure-type lavas covered much of the Kino-Sogo Plateau and the eastern Anza Graben [*Haileab et al.*, 2004; *McDougall and Brown*, 2009]. Extension due to the emplacement of dikes was volumetrically insignificant (5-8%) [*Morley et al.*, 1992; *Ebinger et al.*, 2000].

The earliest manifestation of strain was the development of the late Eocene?-Oligocene to Miocene Lokichar Basin [*Morley*, 1999b; *Morley et al.*, 1999b; *Vétel and Le Gall*, 2006]. Bounded by the east-dipping Lokichar Fault, the Lokichar Basin hosts a ~7 km thick series of fluvial and lacustrine sediments capped by a 300 m thick basaltic sequence dated between 12.5 and 10.7 Ma [*Morley et al.*, 1992, 1999b]. However, the exact nature and age of the oldest strata of the Lokichar Basin is unknown due to a lack of either surface exposure or exploration well data from the deeper parts of the basin. To date, paleocurrent indicators have only been observed in late Oligocene-early Miocene Lokichar Basin strata, consistent with material being transported from the south and southeast [*Williamson and Savage*, 1986; *Morley et al.*, 1992, 1999b; *Wescott et al.*, 1993; *Tiercelin et al.*, 2004]. Formation of the Lokichar Basin was followed by the development of the late Paleogene-late Miocene Lothidok Basin, discussed below (Section 3.2.3). Marked changes in thickness (from 1200 to 100 m) of Eocene-Miocene lavas in the Kino-Sogo area suggest syn-volcanic extension also affected the eastern side of Lake Turkana [*Vétel et al.*, 2005]. Extension during this early phase of rifting is thought to have been ~E-W, based on the predominant ~N-S orientation of major basin bounding faults and the occurrence of N-S striking Oligocene dikes in the Lapur Range area [*Vétel and Le Gall*, 2006]. Beginning in the Miocene, strain progressively migrated eastwards to the neighbouring Turkana and North Kerio Basins [*Morley et al.*, 1999b; *Vétel and Le Gall*, 2006], while simultaneously radiating to the north and south into the southern Ethiopian and central Kenyan rifts [*Nyblade and Brazier*, 2002; *Bonini et al.*, 2005; *Rooney et al.*, 2007]. Locally, this was manifested by northward propagation of the Lokichar Fault and the associated development of the North Lokichar Basin (Section 3.2.3) [*Talbot et al.*, 2004; *Vétel and Le Gall*, 2006; *Tiercelin et al.*, 2012a]. The Kino Sogo Fault Belt (< 3 Ma), a densely spaced minor fault swarm on the east side of Lake Turkana, represents the youngest structural zone within the Kenyan Rift [*Vétel et al.*, 2005].

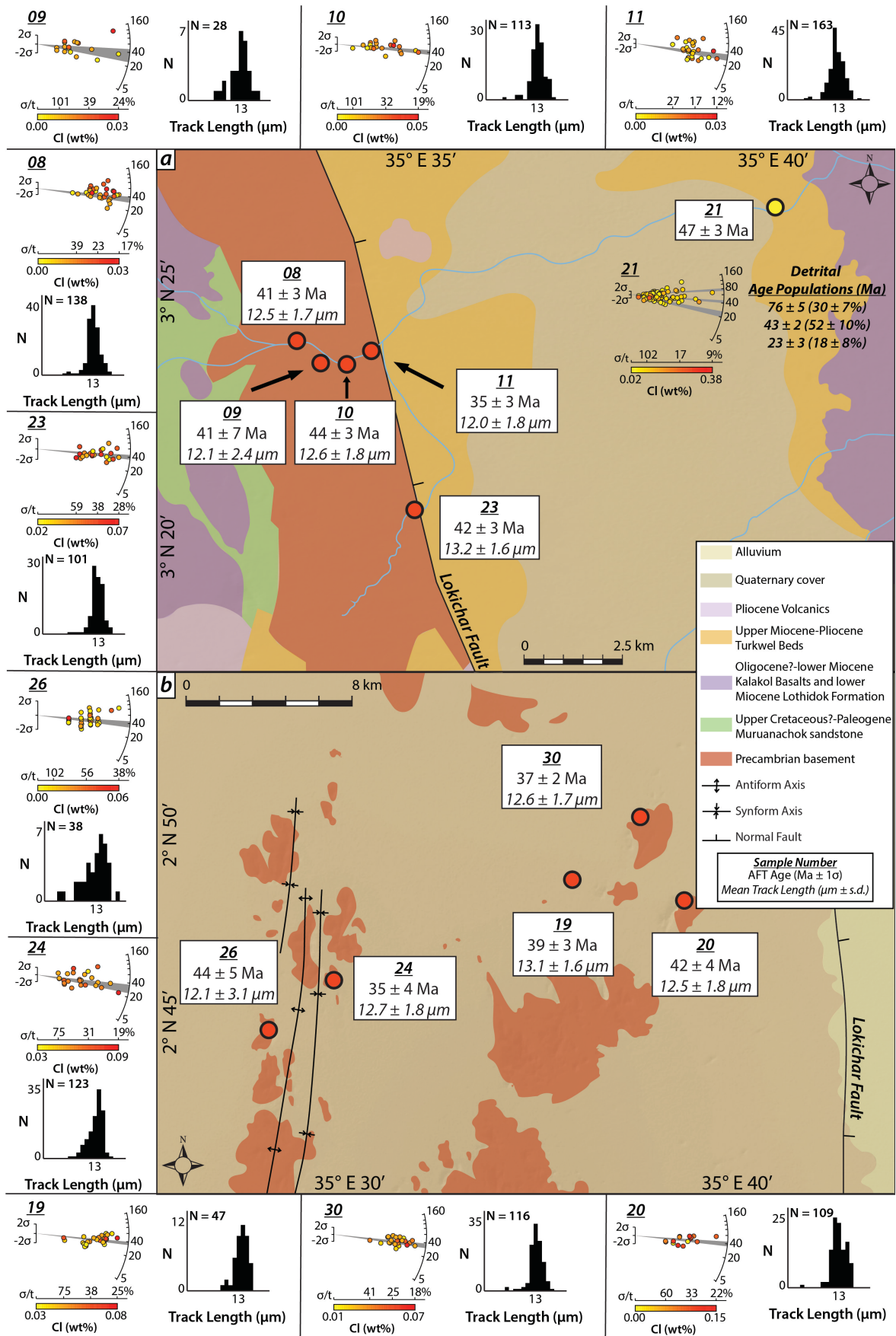


Figure 3.3 (Previous page). Geological maps of the study areas showing locations of basement (red circles) and sedimentary (yellow circles) rock samples from the northern (a) and southern (b) transects across the footwall of the Lokichar Fault that marks the western boundary of the North Lokichar Basin. AFT central ages and MTL for each sample are presented in white boxes. The corresponding confined fission track histograms and radial plots displaying single grain AFT age distributions for each sample are displayed around the border of the maps. The radial plot [Galbraith, 1990] and histogram were constructed using Radial-Plotter [Vermeesch, 2009]. The age for each point on a radial plot is obtained by projecting a line from the origin through the single-grain age (circles) to the radial age axis. All points have the same size error bar corresponding to the axis about the origin on the left showing $\pm 2\sigma$. The further a point plots to the right of the origin, the more precise the individual grain age, as seen by the horizontal precision axis directly below the plot. The colour assigned to each individual point shifts from yellow to red with increasing Cl content (wt%). AFT results for sample 21, a conglomerate from the late Miocene-Pliocene Turkwel Beds, consist of three single grain AFT age populations, as determined by the automatic mixture model function of RadialPlotter [Vermeesch, 2009], that are represented by grey bars in its radial plot. Geological unit boundaries [after Walsh and Dodson, 1969] are underlain by a digital elevation model [NASA, 2000].

3.2.3 North Lokichar Basin

The North Lokichar Basin is a ~N-S trending half-graben in the central Turkana Depression (Figures 3.1 and 3.2). It is characterized by an outcropping, upper syn-rift package that thickens westward into the Lokichar Fault and an underlying, deeper sequence in a sub-basin called the Lothidok Basin, which shows thickening towards the east as revealed by reflection seismic [Morley, 1999b; Morley *et al.*, 1999b]. Sedimentation in the Lothidok Basin is thought to have been governed by the development of a half-graben via modest extension along the inferred west-dipping Lothidok normal fault [Vétel and Le Gall, 2006]. The volcanic nature of the critical Lothidok Hills area and the resulting poor quality of seismic data prevented the identification of the Lothidok Fault [Morley *et al.*, 1999b]. However, the identification of a west-dipping normal and associated rotated strata overlain by a 17.7 Ma lava flow corroborate a period of pre-North Lokichar extensional strain [Boschetto, 1988; Boschetto *et al.*, 1992].

In the overlying North Lokichar Basin, the Lokichar Fault footwall is formed by crystalline basement with very little surface relief (tens of meters) and its outcrop coincides well with the location of the fault as determined from seismic data. Extending over ~150 km, it also serves as the master fault of the adjacent Lokichar Basin to the south (Figures 3.1 and 3.2). Lokichar Fault geometry is well constrained by an extensive series of dip seismic lines acquired by Amoco and reveals significant changes in fault dip (12-70°) and substantial

variation in horizontal displacement (5-16 km) along strike (Figure 3.2) [Morley *et al.*, 1992, 1999b; Morley, 1999b].

To the east, in the Napedet and Lothidok Hills, the North Lokichar Basin is bounded by a flexural margin composed of outcropping Oligocene-Quaternary volcanics and Neogene sediment [Morley *et al.*, 1999b]. The flexural margin also acts as the footwall of an east-dipping fault bounding the western margin of the neighbouring late Miocene-Pliocene southern Turkana Basin (Figures 3.1 and 3. 2). Movement on this fault resulted in isostatic uplift and erosion of the flexural margin, exposing late syn-rift North Lokichar Basin infill.

3.2.3.1 Stratigraphy

Due to the lack of wells in the North Lokichar Basin, its stratigraphy has been inferred primarily by correlating to exposed syn-rift rocks in the flexural margin, particularly in the Lothidok Hills (Figure 3.2). The Lothidok Hills stratigraphy has been divided into three main groups: the Kalakol Basalts, Lothidok Formation and the Turkwel Beds [Boschetto *et al.*, 1992].

The upper Miocene-Pliocene Turkwel Beds, consist of fluvial and lacustrine sediments dipping gently to the west (8-10°) and reaching a minimum thickness of 60 m in the Lothidok Hills [Boschetto *et al.*, 1992]. This unit defines the young, westward-thickening package and represents syn-tectonic sedimentation in the North Lokichar Basin, in the hanging wall of the Lokichar Fault [Morley *et al.*, 1999b].

Underlying the Turkwel Beds unconformably, the Lothidok Formation [18-13 Ma; Boschetto *et al.*, 1992; Morley *et al.*, 1999b] is composed of volcanoclastic and pyroclastic units and some lava flows. It is subdivided into lower and upper units marked by a change in paleocurrent direction from NE-SW to SW-NE, respectively, and the appearance of basement derived material in the upper package detritus [Boschetto *et al.*, 1992].

The Kalakol Basalts [28-18 Ma; Boschetto *et al.*, 1992], whose base is not exposed in outcrop, are the principal unit exposed in the Lothidok Hills. They consist of a series of at least 20 lava flows between 4 to 60 m thick, occasionally intercalated with conglomerates, conglomeratic sandstones, mudstones, volcani- and pyroclastics, reaching a total thickness of at least 785 m [Boschetto *et al.*, 1992; Morley *et al.*, 1999b]. Collectively with the Lothidok

Formation, these represent the older, eastward thickening sequence (Lothidok Basin) that cumulatively makes up ~1400 m of section.

Further south, the middle Miocene volcanic complexes of the Napadet Hills, consisting of plugs, dikes, volcanoclastics and lava flows, define the North Lokichar Basin's flexural margin. On the western side of the hills, the volcanic rocks are overlain by a series of gently dipping (8-10° W) upper Miocene-Pliocene fluvial and lacustrine sediment equivalent to the Turkwel Beds [Morley *et al.*, 1999b].

3.2.4 Previous low-temperature thermochronology studies

Most previous low-temperature thermochronology studies in East Africa are based on AFT data and have revealed a series of regional basement cooling events that began in the Early Mesozoic. The oldest AFT ages (>250 Ma) are recorded in the Tanzanian Craton of SW Kenya and interpreted as recording Early Mesozoic monotonic cooling related to craton exhumation [Wagner *et al.*, 1992; Kasanzu, 2016; Kasanzu *et al.*, 2016]. A pervasive Late Cretaceous-early Paleogene cooling phase, recorded by AFT studies in southern Ethiopia [Philippon *et al.*, 2014; Balestrieri *et al.*, 2016] and throughout central and eastern Kenya [Foster and Gleadow, 1992, 1993, 1996; Torres Acosta *et al.*, 2015], is interpreted as the result of rift-related denudation of the Anza Rift shoulders and adjacent hinterlands. Coeval cooling of similar magnitude was also recorded in NE Tanzania and correlated to changes in regional stress fields in the African plate and erosion of the Indian Ocean passive margin [Noble *et al.*, 1997].

Perhaps surprisingly, few Neogene AFT ages related to EARS faulting and footwall erosion have been recorded in East Africa to date. Instead, only a minor degree of basement cooling (between 10-40 °C) in central Kenya has been inferred from thermal history modelling of AFT and AHe data [Spiegel *et al.*, 2007; Torres Acosta *et al.*, 2015], suggesting the magnitude of Kenya Rift flank exhumation was insufficient to expose rocks which had cooled through the AFT temperature sensitivity range (~60-120 °C). By contrast, Miocene-Pliocene AFT ages from northern Turkana and southern Ethiopia clearly record the onset of pronounced denudational cooling between 20-14 Ma associated with flexural isostatic footwall uplift in response to EARS-extension in the region [Balestrieri *et al.*, 2016; Boone *et al.*, 2018; Philippon *et al.*, 2014]. AHe data from the western margin of the Chew Bahir Basin were interpreted as also recording a period of rift-related cooling that began in the early

Miocene associated with the onset of extensional deformation in southern Ethiopia [Pik *et al.*, 2008]. Boone *et al.* [2018] suggested that increased Neogene denudational cooling recorded by Turkana Depression and southern Ethiopia basement rocks was in part due to increased flexural isostatic footwall uplift in response to EARS extension resulting from the relatively low effective elastic thickness of the lithosphere in these regions [Hendrie *et al.*, 1994; Pérez-Gussinyé *et al.*, 2009].

3.3. Sample Collection and Methodology

Eleven samples were collected to constrain the low-temperature thermal history of the North Lokichar Basin (Figures 3.2 and 3.3). Sample details, including rock types, coordinates and elevations are listed in Table 3.1. For simplicity, samples will be referred to in the text and figures simply by the last two digits of their sample code (i.e. 23 instead of TUB13-23) but are listed in full in the data tables.

Field sampling was conducted along two pseudo-transects into the Lokichar footwall perpendicular to the strike of the Lokichar Fault. The northern transect, approximately 30 km north of the town of Lodwar (Figure 3.2), consists of four gneissic and granulitic basement rocks from the Neoproterozoic-Early Paleozoic Mozambique Belt, collected every ~500 m along an ~E-W oriented, ephemeral river bed, or laga, that incises the footwall. Along the same laga, a cobble conglomerate belonging to the upper Miocene-Pliocene Turkwel Beds was collected (sample 21) from the Lokichar Basin, ~10 km east of the Lokichar Fault in the hanging wall. An additional basement sample (sample 23) was collected along the Lokichar fault scarp ~4 km south of the laga. The southern transect, 30 km south of Lodwar, consists of five gneisses and granulites from the Mozambique Belt basement. Here, samples were collected at larger intervals resulting in a ~ENE-WSW transect beginning ~6 km west of the Lokichar Fault trace and extending nearly 30 km into the exposed Lokichar Fault footwall.

The AFT methodology is based on the formation of damage trails in apatite, called fission tracks, which develop as a consequence of spontaneous fission of ^{238}U [Fleischer *et al.*, 1975; Wagner and Van den Haute, 1992]. Fission tracks are temperature sensitive and begin to anneal when exposed to temperatures above ~60 °C over geological timescales. At temperatures above ~110-120 °C fission tracks anneal completely [Gleadow and Duddy,

1981]. Within this temperature sensitivity range, referred to as the partial annealing zone (PAZ), fission tracks anneal progressively to a length corresponding to the maximum temperature they have experienced [Laslett *et al.*, 1987]. It has been shown that the annealing behaviour of fission tracks in apatite is also compositionally controlled, with chlorine-rich apatites being more resistant to annealing [Barbarand *et al.*, 2003; Donelick *et al.*, 2005]. By combining fission track age and length data the thermal history of a sample within the PAZ can be constrained [Gallagher, 1995; Ketcham, 2005].

AHe and ZHe thermochronology are based upon the production of ^4He during the α -decay of ^{238}U , ^{235}U , ^{232}Th and, in minerals with low to moderate U and Th (e.g. apatite), ^{147}Sm [Farley and Stockli, 2002]. The diffusion of ^4He is a function of time, temperature, crystal size and accumulated radiation damage [Farley, 2000; Reiners and Farley, 2001; Shuster *et al.*, 2006]. In apatite, ^4He diffusion accelerates significantly above $\sim 30^\circ\text{C}$ to nearly instantaneous loss of all ^4He above $\sim 90^\circ\text{C}$ [Reiners *et al.*, 2005; Reiners and Brandon, 2006]. Within these temperatures, referred to as the apatite partial retention zone (AHe PRZ), ^4He is partially retained. Zircon exhibits a higher temperature sensitivity, with a zircon partial retention zone (ZHe PRZ) of $\sim 130\text{--}210^\circ\text{C}$ [Reiners *et al.*, 2002; Stockli, 2005; Biswas *et al.*, 2007] for typical igneous zircons.

A wide range of crystallographic and compositional characteristics have been shown to affect He diffusion in apatite and zircon, at times producing intra-sample He age dispersion greater than 50% [e.g. Wolfe and Stockli, 2010; Wildman *et al.*, 2016]. Some of these potential error sources can be avoided during the crystal selection procedure, such as U-Th inclusions [Danišik *et al.*, 2017]. Other variables accounted for in this study are grain size, morphology and accumulated radiation damage, all of which are considered during data interpretation and thermal history modelling. Here, we use effective uranium concentration ($e\text{U} = \text{U} + 0.235 \times \text{Th}$) as a proxy for α -radiation damage, which considers the decay rate and abundance of radiogenic parent elements. The potential effect grain size may have on resultant He age is assessed by the proxy, equivalent spherical radius (R_s ; $R_s = [3 \cdot R \cdot L] / [2 \cdot \{R + L\}]$) [Beucher *et al.*, 2013], allowing the comparison of 0, 1 and 2 termination grains.

Samples 26 and 30 yielded insufficient apatite grains for He analysis, while samples 19, 20, 26 and 30 did not produce zircon. For samples that did yield suitable grains for ZHe and AHe analyses, an alpha ejection correction (F_T) was applied to calculate a corrected age (Tables

3.2 and 3.3), as suggested by *Farley et al.* [1996]. However, it has been shown that for samples with protracted thermal histories in the PRZ, as is the case for slowly cooled basement terranes, volume diffusion is the dominant mechanism governing He loss. Thus, the application of the F_T correction in these cases will result in inaccurately old ages [*Meesters and Dunai*, 2002; *Danišik et al.*, 2008]. Therefore, all ZHe and AHe data discussed in the text and displayed in figures are uncorrected ages. The effects of alpha ejection are, however, accounted for during inverse thermal history modelling (see Section 3.6). All zircon grains selected for analysis were euhedral with two crystal terminations.

A comprehensive review of the AFT, ZHe and AHe methodologies and technical procedures used in this study can be found in *Gleadow et al.* [2015]. A summary of the AFT, ZHe and AHe analytical protocols are given in the footnotes of Tables 3.1, 3.2 and 3.3, respectively.

3.4. Results

AFT, ZHe and AHe results are presented in Tables 3.1, 3.2, 3.3, respectively. Simplified geological maps of the northern and southern North Lokichar Basin showing sample localities, AFT ages and mean track lengths (MTL) are displayed in Figure 3.3. Pooled and central AFT ages are indistinguishable within analytical uncertainties and all age values displayed in figures and discussed below are central ages. Chlorine concentrations of all samples in this study are very low (mean between 0.01 and 0.09 wt%) and do not exhibit any clear correlation with AFT age, with the possible exceptions of samples 8-11 and 23 discussed below.

All thermochronology ages are reported to one decimal place in the tables but rounded to whole numbers in the text, except for ages with 1σ errors < 1 Ma. Single grain ZHe and AHe ages are reported with a conservative 6.2% (1σ) analytical error. Results are discussed by thermochronometer in order of the traditionally predicted temperature sensitivities from high to low (i.e. ZHe > AFT > AHe). Within these subsections, data are grouped by northern and southern transects.

Table 3.1. North Lokichar Basin AFT Summary Table

Sample No.	Rock Type	Longitude/Latitude	Elevation [m]	No. of grains	N_s	ρ_s [10^5 cm^{-2}]	^{238}U [ppm $\pm 1\sigma$]	Cl [wt%]	Dispersion [%]	Pooled age [Ma $\pm 1\sigma$]	Central age [Ma $\pm 1\sigma$]	N_{length}	Mean track length [$\mu\text{m} \pm \text{se}$]	St.Dev. [μm]
Northern Transect (West-East)														
Footwall														
TUB15-08	Granulite	35.55368/3.39828	519	30	1476	3.7460	19.7 $\pm$ 12.6	0.01	30	36.0 $\pm$ 3.6	41.4 $\pm$ 2.9	138	12.5 $\pm$ 0.2	1.71
TUB15-09	Granulite	35.55841/3.39482	511	17	105	1.4010	8.5 $\pm$ 8.6	0.01	50	35.6 $\pm$ 9.7	40.7 $\pm$ 7.3	28	12.1 $\pm$ 0.5	2.38
TUB15-10	Granulite	35.56157/3.39605	516	21	315	3.8061	17.5 $\pm$ 12.3	0.02	10	41.0 $\pm$ 4.0	43.9 $\pm$ 3.3	113	12.6 $\pm$ 0.2	1.77
TUB15-11	Granulitic-Gneiss	35.56619/3.39834	516	28	2716	4.6235	29.7 $\pm$ 16.2	0.01	48	31.7 $\pm$ 4.5	34.8 $\pm$ 3.4	163	12.0 $\pm$ 0.1	1.81
TUB13-23	Biotite-Gneiss	35.57744/3.36037	520	23	217	0.7483	3.6 $\pm$ 1.5	0.04	0	40.1 $\pm$ 3.4	41.5 $\pm$ 3.3	101	13.2 $\pm$ 0.2	1.64
Hanging wall														
TU13-21	Cobble Conglomerate	35.65062/3.43485	461	86	3560	2.5406	12.3 $\pm$ 19.1	0.07	45	44.6 $\pm$ 3.5	46.8 $\pm$ 2.8	-	-	-
Southern Transect (West-East)														
Footwall														
TUB15-26	Biotite-Granulitic Gneiss	35.45953/2.74783	694	35	97	0.4199	2.4 $\pm$ 0.5	0.02	0	39.1 $\pm$ 4.4	43.9 $\pm$ 4.6	38	12.1 $\pm$ 0.5	3.10
TUB15-24	Granulitic-Gneiss	35.49224/2.76305	761	25	242	1.7350	12.6 $\pm$ 11.6	0.06	47	31.2 $\pm$ 6.4	35.4 $\pm$ 4.3	123	12.7 $\pm$ 0.2	1.80
TUB15-19	Biotite-Granulitic Gneiss	35.59058/2.81264	674	29	340	0.8741	4.6 $\pm$ 1.0	0.05	0	38.4 $\pm$ 2.1	39.2 $\pm$ 2.5	47	13.1 $\pm$ 0.2	1.58
TUB13-30	Biotite-Hornblende Gneiss	35.61449/2.84087	661	24	732	5.0785	30.2 $\pm$ 9.9	0.03	0	35.7 $\pm$ 1.8	36.9 $\pm$ 1.7	116	12.6 $\pm$ 0.2	1.68
TUB15-20	Biotite-Granulitic Gneiss	35.63718/2.80434	662	15	144	0.8405	4.1 $\pm$ 0.5	0.09	0	40.3 $\pm$ 3.6	42.0 $\pm$ 3.9	109	12.5 $\pm$ 0.2	1.84

Notes: Apatite grains were mounted and polished, and fission tracks analysed using the LA-ICP-MS technique following procedures described in *Gleadow et al.* [2015]. Analyses were performed by Samuel C Boone on image sets captured by TrackWorks using a 3.2MP AVT Oscar F-320C camera mounted on a Zeiss AxioImager microscope with a 1000x total magnification and a 100x dry objective (calibration = 0.07 μm by 0.07 μm per pixel). Spontaneous track densities were measured on prismatic internal apatite surfaces after etching with 5M HNO_3 for 20sec at 20°C. Track counts were obtained by automated counting in FastTracks using the 'coincidence mapping' technique of *Gleadow et al.* [2009] followed by manual inspection. Uranium concentrations of each grain were determined by LA-ICP-MS single spot analysis using a New Wave Nd:YAG Laser ($\lambda=213\text{nm}$ with 5Hz @ 45% power, spot size=30 μm) connected to an Agilent 7700 mass spectrometer. NIST612 was used as the primary reference material and a sintered Mud Tank Carbonatite apatite as an in-house secondary reference material during LA-ICP-MS analyses, following the protocol of *Boone et al.* [2016]. Data reduction of LA-ICP-MS results were performed using the TraceElement_{IS} data reduction scheme in software package Lolite [*Paton et al.*, 2011]. Single grain and pooled ages were calculated according to *Hasebe et al.* [2004]. Central ages were estimated from single grain ages and errors according to the formulas given by *Galbraith* [2005, p.100] using the Newton-Raphson method. All ages are "model" ages obtained using a range factor (R_s) of 7.17 μm (average mean track length of Durango and Fish Canyon Tuff standards) and are directly comparable to conventional External Detector Method ages [*Seiler et al.*, 2014]. Confined track lengths (TINTs) were measured as true 3D lengths using *FastTracks* after irradiation by ^{252}Cf and are corrected for a refractive index of 1.634 for apatite. Chlorine concentrations of individual apatite grains were measured using a Cameca SX50 electron microprobe.

N_s = number of spontaneous tracks counted; ρ_s = spontaneous track density; N_{length} = number of lengths measured; se = standard error

Table 3.2. North Lokichar Basin Zircon (U-Th)/He Data Summary

Sample No.	Lab #	⁴ He [ncc]	Mass [mg]	U [ppm]	Th [ppm]	Th/U ratio	[eU] (ppm) ^a	Grain length [μm]	Grain half-width [μm]	R _s ^b	F _T ^c	Uncorrected Age [Ma ± 1σ]	Corrected Age [Ma ± 1σ]	^d Crystal morphology
Northern Transect (West-East)														
Footwall														
TUB15-08	13148	110.499	0.0095	1629.4	404.9	0.25	1724.6	370.4	40.6	54.9	0.78	42.6 ± 2.6	55.0 ± 3.4	2T
TUB15-08	13149	90.978	0.0094	1424.3	308.7	0.22	1496.9	308.2	45.7	59.7	0.79	42.0 ± 2.6	53.1 ± 3.3	2T
TUB15-08	13150	61.940	0.0070	1347.1	319.0	0.24	1422.1	254.7	44.5	56.9	0.78	39.7 ± 2.5	50.9 ± 3.2	2T
TUB15-08	13151	71.058	0.0110	840.6	180.9	0.22	883.1	308.8	50.2	64.7	0.81	48.1 ± 3.0	59.7 ± 3.7	2T
TUB15-09	13142	38.219	0.0109	425.5	123.1	0.29	454.5	316.2	48.9	63.6	0.80	50.6 ± 3.1	63.0 ± 3.9	2T
TUB15-09	13143	43.891	0.0099	513.3	133.8	0.26	544.7	250.0	56.2	68.9	0.81	54.2 ± 3.4	66.6 ± 4.1	2T
TUB15-09	13144	35.640	0.0109	347.1	108.3	0.31	372.6	263.2	56.9	70.2	0.82	58.7 ± 3.6	71.8 ± 4.5	2T
TUB15-09	13145	31.150	0.0056	411.2	200.2	0.49	458.2	221.7	43.6	54.7	0.77	75.9 ± 4.7	99.0 ± 6.1	2T
TUB15-10	13160	197.099	0.0084	3630.9	388.4	0.11	3722.2	241.1	52.2	64.4	0.80	41.7 ± 2.6	51.8 ± 3.2	2T
TUB15-10	13161	230.730	0.0195	1551.9	549.6	0.35	1681.1	343.1	62.6	79.4	0.84	48.4 ± 3.0	57.5 ± 3.6	2T
TUB15-10	13162	103.156	0.0082	1951.3	629.9	0.32	2099.3	339.6	40.4	54.1	0.77	37.6 ± 2.3	49.0 ± 3.0	2T
TUB15-10	13163	191.700	0.0128	2212.3	616.3	0.28	2357.1	317.7	54.2	69.4	0.82	42.4 ± 2.6	51.9 ± 3.2	2T
TUB15-11	13184	59.880	0.0058	1520.7	359.2	0.24	1605.1	217.0	46.7	57.6	0.78	41.1 ± 2.5	53.0 ± 3.3	2T
TUB15-11	13185	209.912	0.0157	1998.9	507.6	0.25	2118.2	341.0	56.0	72.1	0.83	42.7 ± 2.6	51.6 ± 3.2	2T
TUB15-11	13186	63.086	0.0081	1058.7	262.2	0.25	1120.4	305.9	42.3	55.8	0.78	44.1 ± 2.7	56.7 ± 3.5	2T
TUB15-11	13187	67.216	0.0059	1486.1	439.7	0.30	1589.4	265.3	43.0	55.6	0.76	44.4 ± 2.8	58.7 ± 3.6	2T
TUB13-23	13190	32.724	0.0110	149.1	27.1	0.18	155.5	396.8	41.8	56.8	0.78	121.3 ± 7.5	154.7 ± 9.6	2T
TUB13-23	13191	23.393	0.0082	178.8	36.4	0.20	187.4	371.7	37.7	51.3	0.76	94.6 ± 5.9	124.5 ± 7.7	2T
TUB13-23	13192	24.453	0.0100	216.9	34.3	0.16	224.9	294.6	49.3	63.3	0.80	71.1 ± 4.4	88.5 ± 5.5	2T
TUB13-23	15777	12.170	0.0046	202.9	87.3	0.43	223.4	202.5	41.8	52.0	0.75	72.2 ± 4.5	96.3 ± 6.0	2T
TUB13-23	15779	29.795	0.0078	366.0	23.4	0.06	371.5	223.7	53.6	64.9	0.80	67.2 ± 4.2	84.0 ± 5.2	2T
Hanging Wall														
TUB13-21	13156	49.786	0.0082	115.8	26.6	0.23	122.0	306.6	42.6	56.1	0.78	305.2 ± 18.9	391.4 ± 24.3	2T
TUB13-21	13157	50.335	0.0069	147.0	27.6	0.19	153.5	267.2	42.3	54.8	0.77	293.7 ± 18.2	379.0 ± 23.5	2T
TUB13-21	13158	15.916	0.0127	136.7	47.0	0.34	147.7	325.3	52.3	67.6	0.81	56.5 ± 3.5	69.5 ± 4.3	2T
TUB13-21	13159	24.239	0.0078	84.6	96.1	1.14	107.2	236.7	50.7	62.6	0.79	184.7 ± 11.5	234.4 ± 14.5	2T
Southern Transect														
Footwall														
TUB15-24	13194	321.882	0.0244	899.5	287.2	0.32	967.0	425.5	62.2	81.4	0.84	93.7 ± 5.8	110.9 ± 6.9	2T
TUB15-24	13195	167.880	0.0203	735.7	194.6	0.26	781.4	442.2	55.0	73.4	0.83	71.7 ± 4.4	86.5 ± 5.4	2T
TUB15-24	13196	253.887	0.0185	591.0	203.2	0.34	638.8	380.6	57.6	75.1	0.83	144.4 ± 9.0	173.6 ± 10.8	2T
TUB15-24	13197	80.175	0.0157	344.9	171.8	0.50	385.3	372.1	53.2	69.8	0.82	88.2 ± 5.5	107.7 ± 6.7	2T
TUB15-24	15750	363.643	0.0286	671.9	216.5	0.32	722.8	453.7	65.6	86.0	0.85	121.6 ± 7.5	142.6 ± 8.8	2T
TUB15-24	15752	142.112	0.0154	643.0	188.4	0.29	687.2	311.2	61.0	76.5	0.84	91.5 ± 5.7	109.4 ± 6.8	2T
Fish Canyon Tuff standard														
Fish Canyon	13571	9.458	0.0071	333.1	152.5	0.46	369.0	238.5	47.3	-	0.78	23.4 ± 1.4	29.8 ± 1.8	2T
Fish Canyon	13560	11.128	0.0116	231.3	118.7	0.51	259.2	297.8	53.2	-	0.81	24.6 ± 1.5	30.4 ± 1.9	2T
Fish Canyon	13459	6.811	0.0059	300.0	154.8	0.52	336.4	211.7	47.1	-	0.78	21.8 ± 1.4	28.1 ± 1.7	2T
Fish Canyon	13831	9.814	0.0102	238.2	146.7	0.62	272.7	246.1	58.8	-	0.81	23.6 ± 1.5	29.1 ± 1.8	2T
Fish Canyon	14052	7.573	0.0049	410.1	203.4	0.50	457.9	195.9	46.9	-	0.76	21.1 ± 1.3	27.7 ± 1.7	2T
Fish Canyon	15792	9.666	0.0054	447.6	249.5	0.56	506.3	209.9	44.9	-	0.77	22.3 ± 1.4	28.9 ± 1.8	2T

Notes: Clear, euhedral, non-fractured zircons with two terminations were separated manually by hand-picking under an Olympus SZX12 binocular microscope. Zircon grain geometries were imaged and measured microscopically and the loaded into acid-treated platinum capsules. Zircons were then outgassed under vacuum at ~1300°C for 20 minutes, using a semiconductor Coherent Quattro FAP 820 nm diode laser with a fibre-optic coupling to the sample chamber. He volume was determined by isotope dilution using a pure ³He spike, calibrated against an independent ⁴He standard and measured using a Balzers quadrupole (QMS 200 – Prisma) mass spectrometer. A hot blank was run after each gas extraction to verify complete outgassing of the apatite grains. Outgassed zircon grains were transferred from their Pt capsules into Parr bombs, where they were spiked with ²³³U and ²²⁹Th and digested at 240°C for 40 hours in HF. Fish Canyon Tuff zircon (Gleadow et al., 2015) was also run as an ‘unknown’ with each batch zircon samples and served as a check on sample accuracy. Standards and unspiked reagent blanks, were treated identically as unknowns. Samples underwent a second bombing in HCl for 24 hours at 200°C, ensuring dissolution of fluoride salts. After drying, samples were dissolved in HNO₃ solutions and diluted in H₂O to 5% for analysis of ²³⁸U, ²³³U and ²²⁹Th by solution using an Agilent 7700X ICP Mass Spectrometer. For single crystals digested in small volumes (0.3-0.5 ml), U and Th isotope ratios were measured to a precision of up to 2%, but typically better than 1% on an Agilent 7700x ICP-MS. Overall precision for the ZHe method is ~6% or less. α-emission corrected ZHe ages were

^a eU = U ppm + Th ppm * 0.235

^b R_s (equivalent spherical radius) = [3•R•L]/[2•(R+L)], where R = grain radius and L = grain length

^c F_T is the α-ejection correction after Farley et al., [1996].

^d Grain morphology - 0T = no crystal terminations, 1T = one crystal termination and 2T = 2 crystal terminations.

Table 3.3. North Lokichar Basin Apatite (U-Th-Sm)/He Data Summary

Sample No.	Lab #	⁴ He [ncc]	Mass [mg]	U [ppm]	Th [ppm]	Sm [ppm]	Th/U ratio	[eU] (ppm) ^a	Grain length [μm]	Grain width [μm]	R _s ^b	F _T ^c	Uncorrected Age [Ma ± 1σ]	Corrected Age [Ma ± 1σ]	^d Crystal morphology
Northern Transect (West-East)															
Footwall															
TUB15-08	13065	2.885	0.0264	17.2	5.9	100.9	0.34	18.6	379.3	183.0	185.2	0.85	47.7 ± 3.0	56.1 ± 3.5	2T
TUB15-08	13066	2.495	0.0169	20.9	8.9	103.0	0.43	23.0	348.5	166.6	169.1	0.84	52.3 ± 3.2	62.1 ± 3.8	1T
TUB15-08	13067	1.110	0.0134	19.4	7.0	122.8	0.36	21.0	302.0	161.3	157.7	0.84	32.0 ± 2.0	38.4 ± 2.4	1T
TUB15-08	13068	0.626	0.0096	10.3	18.6	83.1	1.81	14.7	239.46	126.4	124.1	0.79	36.1 ± 2.2	45.7 ± 2.8	0T
TUB15-09	13014	0.058	0.0067	2.3	1.3	250.2	0.57	2.6	171.8	124.8	108.4	0.78	24.3 ± 1.5	31.1 ± 1.9	0T
TUB15-09	13015	0.027	0.0040	5.0	4.2	321.7	0.84	6.0	167.2	107.9	98.4	0.72	8.7 ± 0.5	12.1 ± 0.7	2T
TUB15-09	13016	0.047	0.0037	12.6	4.6	295.2	0.37	13.7	179.8	99.5	96.1	0.72	7.5 ± 0.5	10.4 ± 0.6	2T
TUB15-09	13017	0.048	0.0047	5.6	3.6	265.1	0.64	6.4	162.9	107.0	96.8	0.75	12.5 ± 0.8	16.7 ± 1.0	0T
TUB15-10	13069	0.370	0.0062	19.4	10.1	224.7	0.52	21.8	187.53	147.2	123.7	0.80	22.1 ± 1.4	27.6 ± 1.7	1T
TUB15-10	13070	0.656	0.0092	13.0	18.4	204.0	1.41	17.3	210.19	132.1	121.7	0.79	33.1 ± 2.1	41.7 ± 2.6	0T
TUB15-10	13071	2.069	0.0133	9.5	15.8	148.0	1.66	13.2	202.78	161.4	134.8	0.82	94.6 ± 5.9	115.6 ± 7.2	0T
TUB15-10	13072	1.712	0.0150	28.9	10.2	292.2	0.35	31.3	275.95	183.2	165.1	0.85	29.6 ± 1.8	34.9 ± 2.2	1T
TUB15-11	13010	0.313	0.0227	9.1	5.2	78.2	0.57	10.3	286.3	177.7	164.5	0.85	10.8 ± 0.7	12.7 ± 0.8	0T
TUB15-11	13011	0.255	0.0085	5.3	4.5	40.5	0.85	6.4	181.0	136.3	116.6	0.79	38.4 ± 2.4	48.6 ± 3.0	0T
TUB15-11	13012	0.087	0.0154	6.9	6.0	41.8	0.87	8.3	228.6	163.9	143.2	0.83	5.5 ± 0.3	6.6 ± 0.4	0T
TUB15-11	13013	0.326	0.0140	14.9	9.7	50.6	0.65	17.2	251.7	148.8	140.3	0.82	11.1 ± 0.7	13.5 ± 0.8	0T
TUB13-23	13019	0.115	0.0196	3.7	0.7	9.0	0.19	3.9	260.7	173.1	156.0	0.85	12.6 ± 0.8	14.8 ± 0.9	0T
TUB13-23	13020	0.036	0.0209	2.1	0.8	11.7	0.38	2.3	256.7	179.9	158.6	0.85	6.2 ± 0.4	7.3 ± 0.4	0T
TUB13-23	13021	0.582	0.0323	4.5	0.7	12.6	0.16	4.7	330.1	197.3	185.2	0.87	31.5 ± 2.0	36.1 ± 2.2	0T
Hanging wall															
TU13-21	13031	0.136	0.0186	1.5	9.5	106.4	6.33	3.7	270.6	181.9	163.2	0.82	15.4 ± 1.0	18.7 ± 1.2	2T
TU13-21	13032	0.273	0.0107	4.2	31.9	141.8	7.60	11.7	274.3	136.9	137.0	0.78	17.7 ± 1.1	22.8 ± 1.4	2T
TU13-21	13033	0.104	0.0179	2.9	1.8	6.4	0.62	3.3	272.9	161.6	152.3	0.84	14.4 ± 0.9	17.2 ± 1.1	0T
TU13-21	13034	0.308	0.0263	4.4	6.1	54.0	1.39	5.8	524.0	164.2	187.5	0.84	16.2 ± 1.0	19.2 ± 1.2	1T
Southern Transect (West-East)															
Footwall															
TUB15-26	12988	0.121	0.0241	1.7	3.2	18.1	1.88	2.5	313.1	174.9	168.3	0.85	16.4 ± 1.0	19.4 ± 1.2	0T
TUB15-26	12999	0.151	0.0244	1.7	3.3	20.7	1.94	2.5	294.5	181.7	168.5	0.85	20.3 ± 1.3	24.0 ± 1.5	0T
TUB15-26	13000	0.048	0.0126	1.5	4.2	23.0	2.80	2.5	195.9	160.0	132.1	0.81	12.4 ± 0.8	15.2 ± 0.9	0T
TUB15-26	13001	0.150	0.0333	1.7	3.9	27.1	2.29	2.6	296.3	211.3	185.0	0.86	14.2 ± 0.9	16.4 ± 1.0	0T
TUB15-24	12993	0.152	0.0066	8.9	2.1	12.9	0.24	9.4	256.6	101.4	109.0	0.77	20.0 ± 1.2	26.0 ± 1.6	0T
TUB15-24	12994	0.107	0.0029	9.4	5.4	11.2	0.57	10.7	206.2	74.8	82.3	0.68	28.3 ± 1.8	41.5 ± 2.6	0T
TUB15-24	12995	0.548	0.0084	18.3	3.0	29.3	0.16	19.0	313.1	121.4	131.3	0.80	28.2 ± 1.7	35.4 ± 2.2	1T
TUB15-24	12996	0.058	0.0125	2.1	1.3	14.7	0.62	2.4	183.5	164.8	130.3	0.82	15.4 ± 1.0	18.7 ± 1.2	0T
TUB15-19	12981	0.198	0.0140	2.7	2.3	32.6	0.85	3.2	279.2	141.4	140.8	0.82	34.9 ± 2.2	42.6 ± 2.6	0T
TUB15-19	12982	0.171	0.0192	3.0	2.2	33.5	0.73	3.5	219.5	186.3	151.2	0.85	20.5 ± 1.3	24.3 ± 1.5	0T
TUB15-19	12983	0.067	0.0066	4.2	3.7	44.6	0.88	5.1	150.1	132.8	105.7	0.78	16.1 ± 1.0	20.7 ± 1.3	0T
TUB15-19	12984	0.360	0.0222	3.5	2.1	37.0	0.60	4.0	246.8	189.1	160.6	0.85	33.1 ± 2.1	38.8 ± 2.4	0T
TUB15-20	12985	0.101	0.0062	5.3	3.2	58.0	0.60	6.1	235.0	124.3	122.0	0.78	21.8 ± 1.4	27.9 ± 1.7	1T
TUB15-20	12986	0.067	0.0088	4.9	4.5	58.2	0.92	6.0	215.9	161.8	138.8	0.82	10.5 ± 0.6	12.8 ± 0.8	1T
TUB15-20	12987	0.067	0.0105	4.6	2.5	42.9	0.54	5.2	219.5	137.8	127.0	0.81	10.0 ± 0.6	12.4 ± 0.8	0T
TUB15-20	12988	0.081	0.0152	4.0	1.3	38.3	0.33	4.3	239.1	159.0	143.2	0.84	10.0 ± 0.6	12.0 ± 0.7	0T
Durango standard															
Durango	13055	3.974	-	-	-	-	26.42	-	-	-	-	1.00	31.3 ± 1.9	31.3 ± 1.9	-
Durango	13056	4.347	-	-	-	-	26.04	-	-	-	-	1.00	31.4 ± 1.9	31.4 ± 1.9	-
Durango	13073	2.069	-	-	-	-	19.61	-	-	-	-	1.00	31.2 ± 1.9	31.2 ± 1.9	-
Durango	13109	11.244	-	-	-	-	21.41	-	-	-	-	1.00	31.3 ± 1.9	31.3 ± 1.9	-
Durango	15910	31.234	-	-	-	-	19.99	-	-	-	-	1.00	32.3 ± 2.0	32.3 ± 2.0	-
Durango	15952	31.993	-	-	-	-	23.23	-	-	-	-	1.00	29.8 ± 1.8	29.8 ± 1.8	-
Durango	16008	15.041	-	-	-	-	20.72	-	-	-	-	1.00	29.7 ± 1.8	29.7 ± 1.8	-

Notes: Clear, euhedral, non-fractured apatites were separated manually by hand-picking under an Olympus SX12 binocular microscope. Apatites, in acid-treated platinum capsules, were outgassed under vacuum at ~900 °C for 5 minutes, using a semiconductor Coherent Quattro FAP 820 nm diode laser with fibre-optic coupling to the sample chamber. He volume was determined by isotope dilution using a pure ³He spike, calibrated against an independent ⁴He standard and measured using a Balzers quadrupole (QMS 200 – Prisma) mass spectrometer. A hot blank was run after each gas extraction to verify complete outgassing of the apatite grains. The second re-extract contributed less than 0.5% of the total measured ⁴He for all samples. ²³⁸U, ²³⁵U, ²³²Th and ¹⁴⁷Sm content was obtained by total dissolution of outgassed apatites, still in platinum capsules, in HNO₃ and analysed using an Agilent 7700x ICP-MS. Durango apatite [McDowell *et al.*, 2005] and Fish Canyon Tuff zircon [Gleadow *et al.*, 2015] were also run as ‘unknowns’ with each batch of apatite or zircons samples and served as a check on sample accuracy. (U-Th-Sm)/He ages were calculated and corrected for α-emission following the approach of Farley *et al.*, [1996]. Analytical uncertainties, including the α-ejection correction, an estimated 5 μm uncertainty in grain dimension, gas analysis (<1%) and ICP-MS analytical uncertainties, are conservatively assessed at ~6.2%. Accuracy and precision of U, Th and Sm content range up to 2%, but are typically <1%.

^a eU (ppm) = U ppm + 0.235*Th ppm

^b R_s (equivalent spherical radius) = [3•R•L]/[2•(R+L)], where R = grain radius and L = grain length

^c F_T is the α-ejection correction after Farley *et al.*, [1996].

^d Grain morphology - 0T = no crystal terminations, 1T = one crystal termination and 2T = 2 crystal terminations.

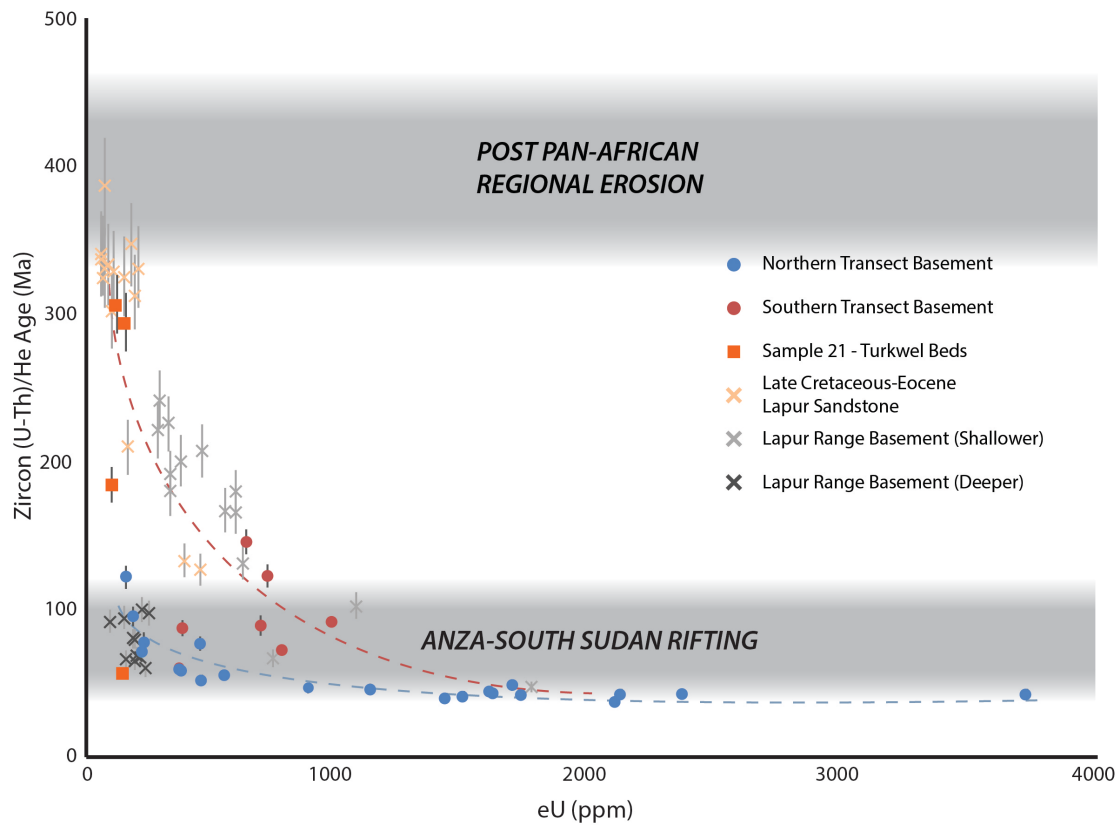


Figure 3.4. Plot of ZHe age versus eU content. ZHe results from the Lapur Range of northern Turkana (Chapter 5) [Boone et al., 2018] are also shown for comparison. Dashed lines portray the distinct age-eU trends displayed by northern (blue) and southern (red) transect samples, suggesting these two regions may have experienced different Mesozoic to early Cenozoic thermal histories. For zircons that share a similar tectonic history and have a similar range of eU content, older ages suggest longer residence in the ZHe PRZ, while younger grains indicate comparatively faster cooling and/or from deeper crustal levels, substantially resetting their ages. These trends are in general agreement with the spread of ZHe data from the Lapur Range, where deeper basement rocks recorded a greater degree of Cretaceous cooling than shallower basement samples and overlying sedimentary rocks. Approximate timing of regional tectonothermal events that affected the study area are illustrated in grey.

3.4.1 Zircon (U-Th)/He (ZHe) results

3.4.1.1 Northern Transect

Basement samples collected within the Lokichar Fault footwall, north of Lodwar, yield single grain ZHe ages ranging between 38 ± 2 Ma and 121 ± 8 Ma (21 grains) that negatively correlate with eU content (155.5 – 3722.2 ppm; Table 3.2 and Figure 3.4). Analysed zircons exhibit no correlation between He age and grain size. Samples 8-11 yielded Late Cretaceous-Eocene single grain ZHe ages (38 ± 2 to 76 ± 5 Ma) with relatively little intrasample

dispersion. By contrast, sample 23, collected ~4 km to the south of samples 8-11, yielded an older, more disperse range of Cretaceous single grain ages (67 ± 4 to 121 ± 8 Ma).

Single grain ZHe ages from the Turkwel Beds conglomerate collected in the Lokichar Fault hanging wall (sample 21) range between 57 ± 4 Ma and 305 ± 19 Ma (4 grains) and are independent of grain size (Table 3.2). eU concentrations for these grains are relatively low (107.2 – 153.5 ppm).

3.4.1.2 Southern Transect

Sample 24 was the only one from the southern transect to yield suitable zircons for analysis. Single grain ZHe ages range from 72 ± 4 Ma to 144 ± 9 Ma (6 grains) with intermediate to high eU concentrations (385.3 – 967.0 ppm; Table 3.2). These show no correlation to crystal size.

3.4.2 Apatite fission track (AFT) results

3.4.2.1 Northern Transect

AFT analyses from the northern Lokichar Fault footwall record middle Eocene to middle Oligocene ages (44 ± 3 to 35 ± 3 Ma; Table 3.1). Samples that yielded a significant number of confined track measurements (> 100) have moderate MTL (12.0 to 13.2 μm) that positively correlates with chlorine content (0.01 to 0.04 wt%), while their modest standard deviations (1.6 to 1.8 μm) are negatively correlated to Cl content. Only 28 confined tracks were measured for sample 9, yielding a MTL of 12.1 μm , consistent with other northern transect basement samples, while producing a significantly higher standard deviation of 2.38 μm .

Sample 21 yielded a central AFT age of 47 ± 3 Ma that predates its upper Miocene-Pliocene depositional age. Single grain AFT ages vary significantly (45% dispersion), ranging between 8 ± 5 and 180 ± 27 Ma. Statistically, AFT single grain ages can be separated into three single grain AFT age populations [23 ± 3 Ma ($17.9 \pm 7.7\%$), 43 ± 2 Ma ($51.6 \pm 9.6\%$), and 76 ± 5 Ma ($30.4 \pm 7.2\%$)], as determined by the mixture model function of *RadialPlotter* [Vermeesch, 2009] (Figure 3.3a).

3.4.2.2 Southern Transect

The five samples collected from the exposed southern Lokichar Fault footwall produced reproducible Eocene AFT ages (44 ± 5 to 35 ± 4 Ma; Table 3.1). The youngest of these, sample 24 (35 ± 4 Ma), exhibits a relatively large spread of single grain ages (47% dispersion, compared to 0% for all other southern transect samples; Table 3.1 and Figure 3.3b). Samples with statistically robust MTL values yield moderate lengths (12.5 to 12.7 μm) with intermediate standard deviations (1.7 to 1.8 μm). On the other hand, samples 26 and 19 only produced 38 and 47 length measurements, respectively. Nonetheless, their MTL are consistent with other southern transect samples (12.1 to 13.1 μm and standard deviations of 3.1 and 1.6 μm , respectively).

3.4.3 Apatite (U-Th-Sm)/He (AHe) results

3.4.3.1 Northern Transect

Single grain AHe ages from the northern Lokichar Fault footwall are highly dispersed, ranging from 5.5 ± 0.3 Ma to 95 ± 6 Ma (Table 3.3). Samples 8 (4 grains) and 23 (3 grains) yielded single grain AHe ages that appear to positively correlate with equivalent spherical radius, while sample 11 exhibits a potential negative correlation (Figure 3.5). Strong correlations between AHe age and eU content are not present except for a positive trend in sample 23. Samples 9 and 10 appear to show weakly negative age-eU correlations (Figure 3.5).

Single grains from TU13-21, which yielded Neogene AHe ages (14.4 ± 0.9 to 18 ± 1 Ma; 4 grains), show no correlation with either eU or R_s .

3.4.3.2 Southern Transect

Apatites from the southern transect yield a narrower range of AHe ages (10.0 ± 0.6 to 35 ± 2 Ma; Table 3.3). No relationship between AHe age and either eU content or R_s is apparent for southern transect samples apart from samples 19 and 24, which perhaps exhibit a slight correlation with R_s and, in the case of sample 24, eU (Figure 3.5).

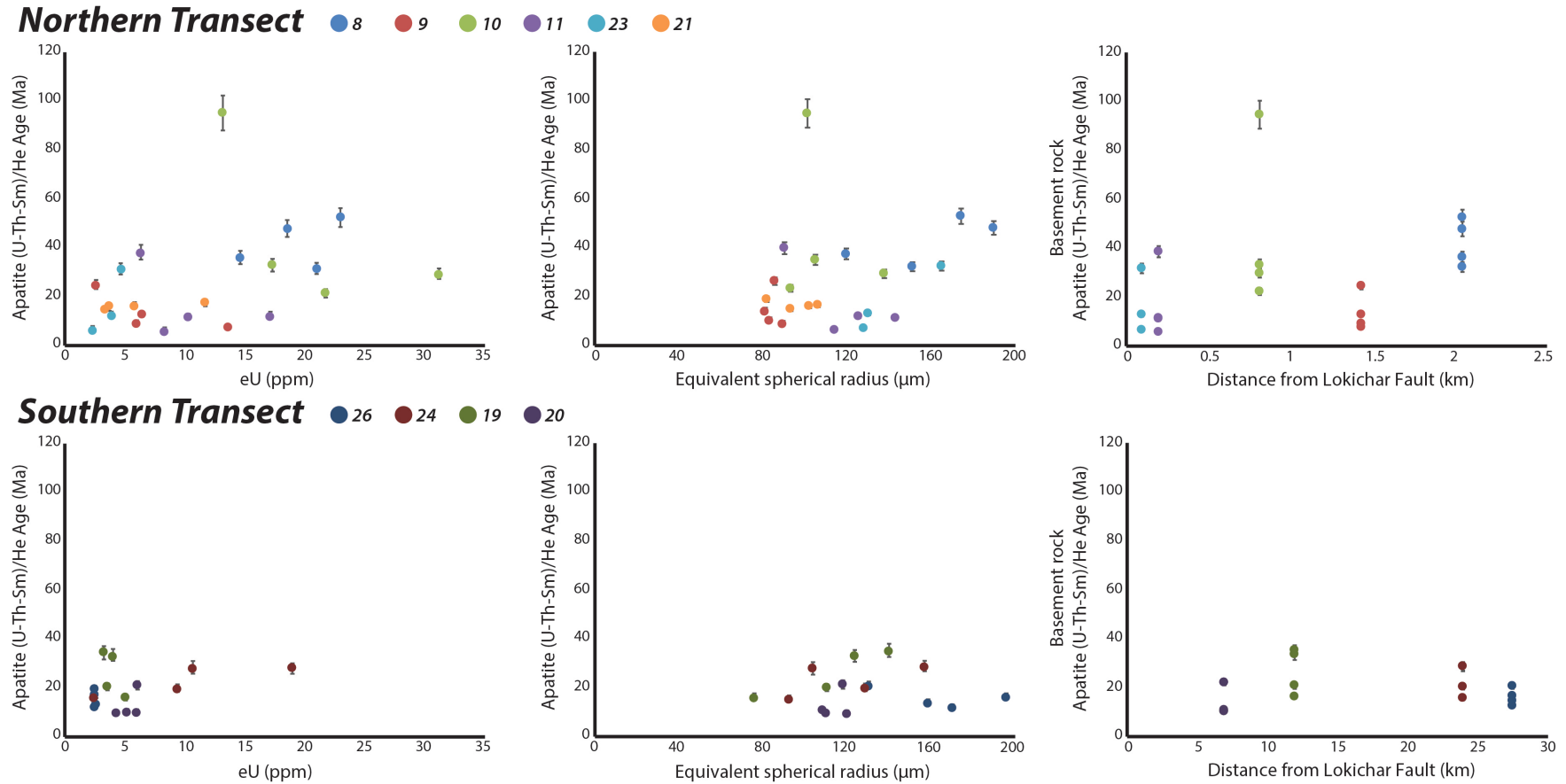


Figure 3.5. Plots of AHe age versus eU content (left column), AHe age versus equivalent spherical radius (middle column) and basement rock AHe age (sample 21 omitted) versus orthogonal distance from the Lokichar Fault (right column) for northern (top row) and southern (bottom row) transect samples. Apatite eU concentrations are generally low. Single grain ages predominantly show no clear relationship with eU content and grains size. Some possible exceptions where single grain ages may exhibit weak correlations with eU and grain size are discussed in Section 3.4.3. In general, basement rock single grain AHe ages decrease and intra-sample AHe age skewness becomes more positive with increasing proximity to the Lokichar Fault (Section 3.5.3).

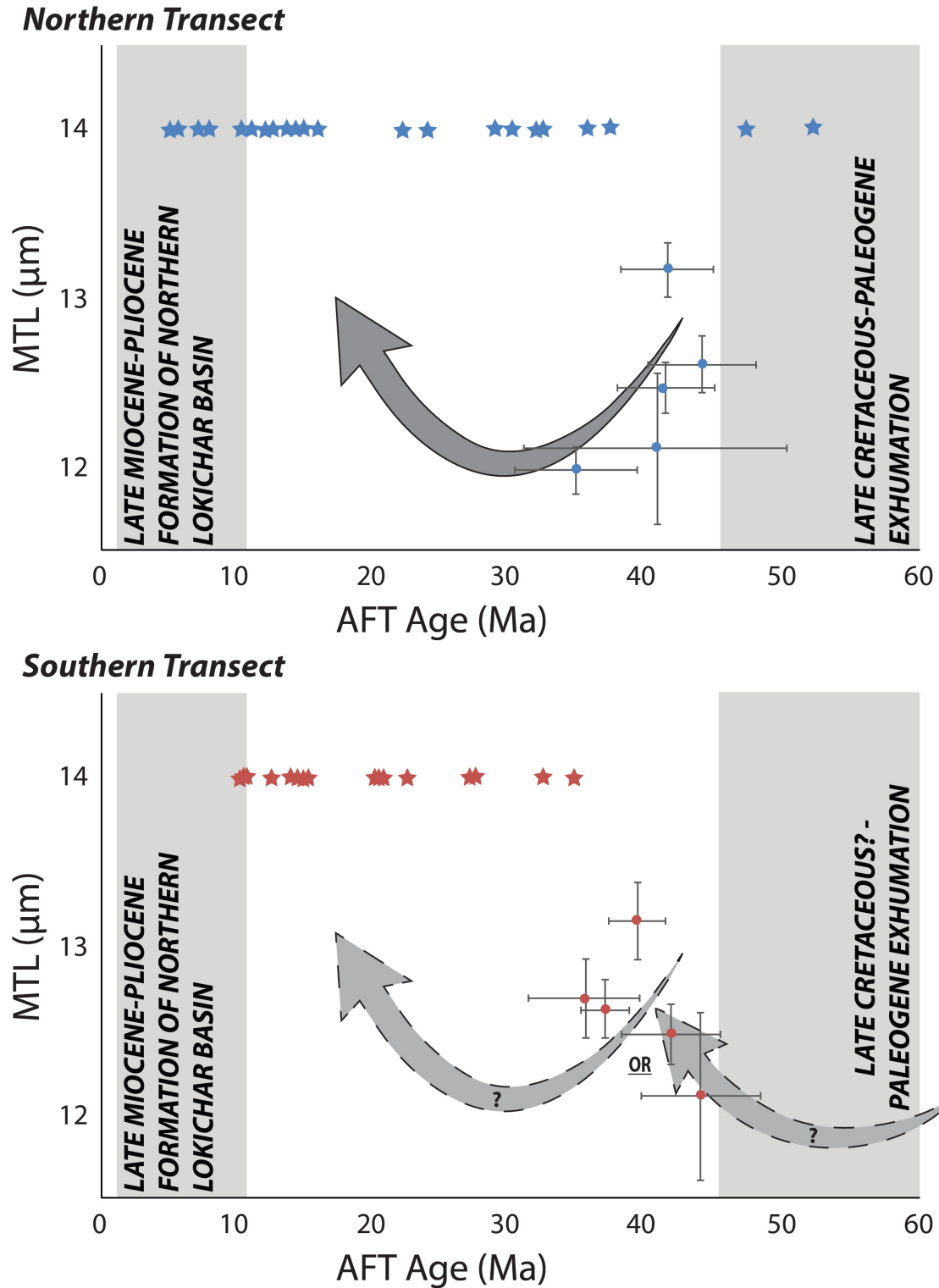


Figure 3.6. Boomerang plots of AFT age versus MTL (circles) for northern and southern transect basement rock samples. Single grain AHe ages are displayed as stars. Note, one northern transect AHe grain plots off the graph due to its old age (95 ± 6 Ma) and is thus not shown. Grey bars illustrate approximate duration of tectonothermal events recorded by AFT and AHe data. Grey arrows represent the boomerang curves which AFT data partially form. Southern transect AFT data may represent the descending arm of a boomerang curve (left arrow), suggesting these samples experienced a similar thermal history to that recorded by northern transect samples. Alternatively, they form the ascending portion of a boomerang curve (right arrow) indicating that prior to late Cenozoic cooling these samples experienced an older Mesozoic cooling phase than that recorded by northern transect samples did. See Section 3.5.2 for description of a boomerang plot and data interpretation.

3.5. Discussion

ZHe, AFT and AHe analyses from the North Lokichar Basin yield complex datasets that exhibit varying degrees of inter- and intra-sample age dispersion related to the competing effects of thermochronometer-specific systematic controls and spatio-temporal relationships between samples. Therefore, before exploring the geological implications of these data, the effects of chemical, crystallographic and radiation controls on radiometric daughter product retention must first be discussed.

3.5.1 Zircon (U-Th)/He age dispersion and radiation-enhanced He diffusion

ZHe results from North Lokichar Basin samples exhibit varying degrees of intra-sample single grain age dispersion that show negative correlations with eU content, while showing no discernible relation to grain size. Negative ZHe age-eU correlations have been observed in several thermochronology investigations of basement terranes with protracted cooling histories [e.g. *Nasdala et al.*, 2004; *Guenther et al.*, 2014, 2017; *Powell et al.*, 2016; *Mackintosh et al.*, 2017] and have been interpreted as the result of intra-grain differences in radiation-damage accumulation [*Guenther et al.*, 2013]. In this model, the degree of radiation damage and He diffusivity coevolve during a sample's thermal history [*Ketcham et al.*, 2013]. Therefore, variations in eU content and thus accumulated radiation damage between zircons that share a similar thermal history result in a range of He diffusivities and temperature sensitivities specific to each grain. Consequently, each individual zircon acts as its own thermochronometer, recording a different He age from the same time-temperature history [*Guenther et al.*, 2017]. At low radiation damage levels, zircon He diffusivity decreases with increasing eU content. Yet, once a grain passes a certain radiation damage threshold He diffusivity begins to increase (or temperature sensitivity decrease) with increasing eU [*Guenther et al.*, 2013], as is the case for ZHe data presented here.

ZHe age-eU relationships in this study can be visualised more clearly when zircons from multiple samples are grouped together, effectively increasing the range of eU content spanned. To do this, it must be assumed that samples grouped together share a similar thermal history. This is easily done for basement samples from the northern transect of the North Lokichar Basin, due to their proximity (within a ~4 km radius), similar elevation (511-520 m) and the lack of crustal structures between them. However, the structural relationship

between the northern and southern transects is not sufficiently clear due to Quaternary sediment and alluvium obscuring the underlying basement architecture, while seismic lines interpreted in the published literature lie primarily to the east of this area [Morley *et al.*, 1992, 1999b]. Therefore, northern and southern transect data are interpreted separately.

Northern transect zircons span a wide range of eU content (155.5-3722.2 ppm), with the lowest eU grain corresponding to the oldest ZHe age (121 ± 7 Ma; sample 23). As expected for zircons that experienced significant radiation damage, ZHe ages decrease with increased eU concentrations. However, the slope of this trend quickly reduces and essentially plateaus above ca. 1000 ppm eU values, corresponding to ZHe ages of 48 ± 3 to 37 ± 2 Ma. In contrast, zircons from sample 24 (southern transect; 6 grains) are of intermediate eU concentrations (385.3-967.0 ppm) and plot above the age-eU trend of northern transect samples (ZHe ages of 144 ± 9 to 72 ± 4 Ma). This relationship possibly suggests that the southern North Lokichar Basin footwall might have experienced a somewhat different pre-Oligocene thermal evolution than its northern counterpart. Possible scenarios include: 1) coeval denudation from different structural depths, where older samples were exhumed from shallower crustal levels, or at different cooling rates, and 2) the southern Lokichar footwall experienced cooling through the ZHe PRZ at an earlier time than the northern portion of the footwall. However, given the low number of ZHe ages from the southern transect, it is not yet clear if these results correspond to a systematic change in regional denudation patterns.

3.5.2 Apatite fission track data

AFT results are summarised in AFT single grain age v. MTL plots for northern and southern transect basement samples (Figure 3.6). These so-called boomerang plots are useful for illustrating the relationship between AFT ages and MTL for a set of samples that have experienced a multi-phase thermal history [Green *et al.*, 1989; Gallagher and Brown, 1997]. AFT analyses from northern transect basement rocks form the descending arm of a boomerang curve. Such AFT age-MTL relationships are characteristic of an early period of pronounced cooling followed by a prolonged period of residence within or reheating into the AFT PAZ, consistent with the intermediate MTL (12.0 to 13.2 μm) and moderate standard deviations (1.6 to 1.8 μm) of northern transect samples (excluding sample 9; see above). Confined track distributions (Figure 3.3) also point to a thermal perturbation of some kind, evident by their small tails extending towards shorter track lengths. A subsequent period of cooling and exhumation postdating the youngest AFT central age (35 ± 3 Ma) is then

required to expose samples with substantially shortened MTL at the surface. This is consistent with the 14 Oligocene-Miocene single grain AHe ages that make up the youngest and largest portion of the widely dispersed northern transect AHe dataset, discussed further below (Section 3.5.3). In general, AFT age and MTL decrease with decreasing Cl content (Table 3.1), as expected for annealing kinetics that are partially controlled by crystal chemistry [Barbarand *et al.*, 2003].

The AFT age-MTL trend for southern transect samples is less clear. AFT results appear to form the ascending portion of a boomerang plot, which might suggest that the older, inherited cooling event reset by subsequent thermal perturbation is different than, and perhaps predates, that recorded by northern transect samples (right arrow; Figure 3.6). This would then also explain the older ZHe ages obtained for southern transect sample 24. Alternatively, southern transect samples may have experienced a similar cooling history as northern transect samples with respect to the AFT system (left arrow; Figure 3.6); in this case, the perceived ascending trend would only be a product of the small sample size ($n = 5$). This interpretation would favour the older ZHe ages from sample 24 resulting from either slower cooling during Late Cretaceous-Paleogene time or cooling from lower maximum paleotemperatures. In either case, a period of cooling at some point in Oligocene-Recent times was required to exhume rocks with substantially reduced MTL and Eocene AFT ages. This is consistent with latest Eocene-early Miocene southern transect AHe ages, all of which are younger than southern transect AFT results for the same samples.

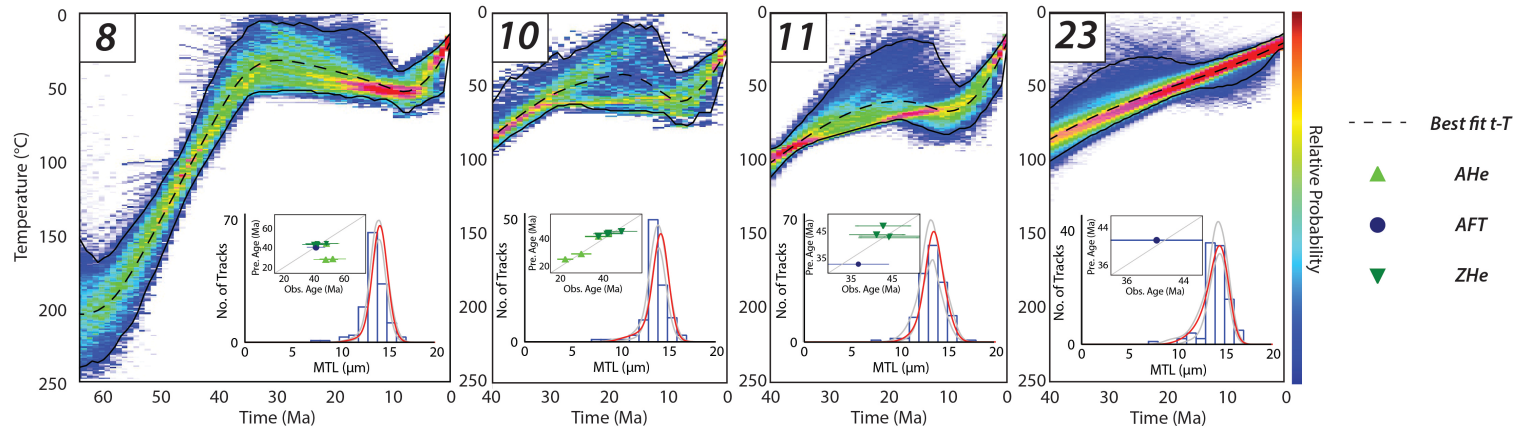
3.5.3 Crystallographic and radiation damage effects on apatite (U-Th-Sm)/He age dispersion

AHe results from both northern and southern transects yield largely dispersed age ranges that do not consistently correlate with either eU or R_s (Figure 3.5). This suggests that the results are influenced by the competing effects of grain size and radiation damage [Brown *et al.*, 2013], U, Th, and Sm zonation [Flowers and Kelley, 2011] and apatite chemistry [Gautheron *et al.*, 2013; Mbongo Djimbi *et al.*, 2015], which are difficult to quantify using the analytical procedure employed. Analysis of fragmented apatite grains likely further convolutes these relationships by producing ages that differ from their whole grain equivalents [Wildman *et al.*, 2016]. One termination (1T) fragments can produce older or younger ages than the whole grain age, while no termination (0T) grains are most likely to yield older ages [Brown *et al.*, 2013].

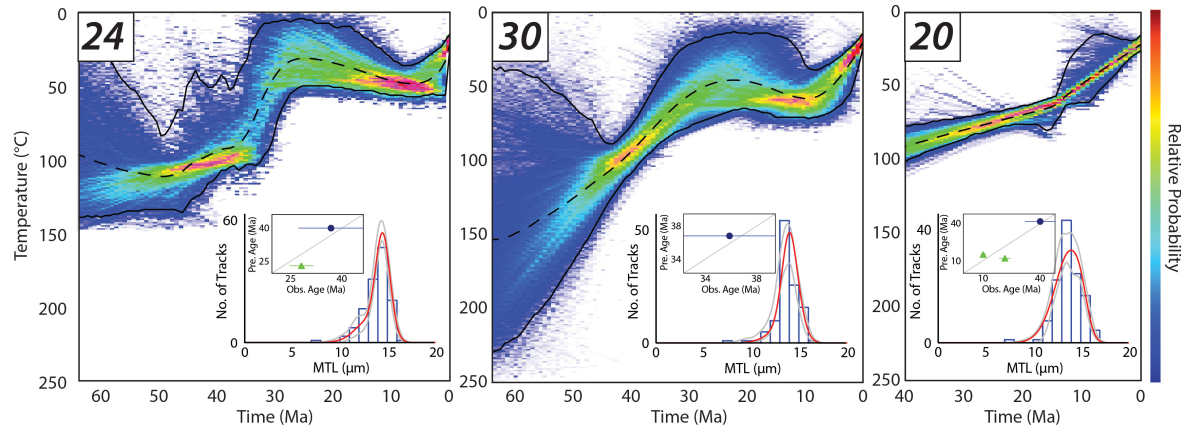
While we are unable to identify a single systematic source for the observed age discrepancy, inter-aliquot dispersion itself provides important insights into the sample's thermal history [Flowers and Kelley, 2011; Brown *et al.*, 2013; Wildman *et al.*, 2016], with dispersion increasing in samples that have experienced prolonged residence in the AHe PRZ [Fitzgerald *et al.*, 2006; Brown *et al.*, 2013]. By comparison, samples that underwent rapid cooling yield lower inter-grain age dispersion [Flowers and Kelley, 2011]. Additionally, the skewness of aliquot AHe ages has been shown to change from negative (skewed towards older ages) to positive (skewed towards younger ages) as time spent in the AHe PRZ decreases, eventually approaching a symmetric single grain age distribution around the sample mean age for samples that rapidly cooled through the AHe PRZ [Fitzgerald *et al.*, 2006; Brown *et al.*, 2013].

Highly dispersed single grain AHe ages from northern transect samples (95 ± 6 to 5.5 ± 0.3 Ma) are moderately skewed towards younger late Miocene times, consistent with protracted residence in or reheating into the PRZ as indicated by AFT age-MTL relationships (Figure 3.6). By contrast, AHe ages from southern transect samples (35 ± 2 to 10.0 ± 0.6 Ma) exhibit a slightly higher degree of skewness towards younger ages (late Oligocene-Miocene), indicating that these samples experienced cooling from higher paleotemperatures, more rapid cooling rates, and/or shorter residence in the AHe PRZ than their northern transect counterparts. Similar trends can be seen within the northern and southern transects, where intra-sample AHe age skewness becomes more positive the closer a sample is to the Lokichar Fault (Figure 3.5). This is consistent with samples residing in the more proximal Lokichar Fault footwall having cooled through the AHe PRZ more rapidly. Regardless of the issues in assigning a single root cause for AHe age dispersion, it is still possible to use these data for thermal history modelling where they are complemented by data from other low-temperature thermochronometers (Section 3.6).

a) Northern Transect



Southern Transect



b) 20: Forward Model

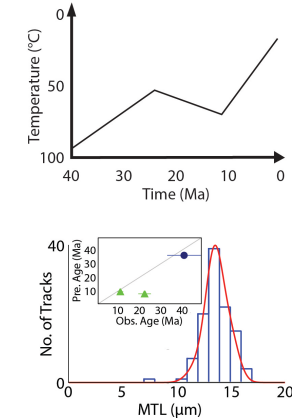


Figure 7. Thermal history models for northern and southern transect basement samples determined using the QTQt software. 150,000 iterations were run during thermal history modelling of individual samples, with the first 50,000 discarded as burn-in [Gallagher et al., 2009]. The data fit for each thermal history model is illustrated by panel inserts showing observed versus predicted thermochronology ages (note that He ages are displayed with resampling errors – see text) and of the observed confined track distributions overlain by the predicted distributions (red curve) and the 95% confidence interval (grey curves) of the predicted distributions. **a)** Inverse thermal history models. The thermal history model t-T paths for sample 8 over the time span of 65–40 Ma is representative of all northern transect samples (See Supplementary Data A–C) and is not repeated for samples 10, 11 and 23. The colour map represents the posterior-probability distribution of accepted time-temperature paths, with warmer colours indicating higher probabilities. Black solid lines represent the 95% confidence interval limits. Refer to Section 3.6 for the thermal history modelling protocol. **b)** Forward thermal history model for sample 20.

3.6. Thermal History Reconstructions

In order to quantify thermal histories for basement samples from the footwall of the northern Lokichar Fault, ZHe, AFT and AHe data were inverse modelled jointly using the modelling software *QTQt* [Gallagher, 2012], with the multi-compositional fission track annealing model of *Ketcham et al.* [2007a]. *QTQt* utilizes a Bayesian transdimensional Markov chain Monte Carlo (MCMC) sampling method to generate a range of acceptable thermal histories, quantified in terms of a posterior probability distribution [Gallagher *et al.*, 2009; Gallagher, 2012]. The resulting expected thermal history models (weighted mean of the posterior distribution) [see Gallagher, 2012 and Wildman *et al.*, 2016 for discussion] are shown in Figure 3.7. Additional information about the thermal history modelling protocol can be found in the caption of Figure 3.7 and in Supplementary Data A-C. Samples with less than 100 confined fission track lengths were omitted during thermal history modelling. Confined fission track lengths were normalised for annealing and etching anisotropy using the *c*-axis projection method [Ketcham *et al.*, 2007b]. Differences in ZHe and AHe grain-size were accounted for by using an alpha ejection correction (F_T) [Farley *et al.*, 1996]. Radiation damage accumulation and annealing models were employed for modelling of AHe data [Gautheron *et al.*, 2009] and ZHe data [ZRDAAM; Guenthner *et al.*, 2013] to account for differences in radiation-modified diffusion kinetics recorded by age-eU relationships. However, these models are unable to account for all known sources of intrasample ZHe and AHe age dispersion, such as chemical zonation [Guenthner *et al.*, 2013] and apatite composition [Gautheron *et al.*, 2013]. Therefore during modelling, *QTQt* was allowed to resample the He error for single grain ages, effectively decreasing their influence on determining thermal histories, as compared to data from the more robust AFT system [Gallagher, 2012]. For samples with fragmented AHe grains, only 1- and 2-termination grains were modelled in *QTQt* due to current limitations in the implementation of the *Brown et al.* [2013] fragmentation model.

As apatite fission track annealing kinetics are well understood, samples were first modelled using solely their AFT data. Thermal history modelling then proceeded using sample AFT and AHe data, when available, before a final modelling run was carried out that also included ZHe data. These results were then compared with one another to ensure consistency.

A single time-temperature constraint was applied to all models, requiring samples to be at present day surface temperatures of 20 ± 10 °C.

3.6.1 Time-temperature histories

Representative thermal history models that reproduce observed thermochronometer ages are displayed in Figure 3.7. Additional thermal history models not included in Figure 3.7 are presented in Supplementary Data A-C.

For samples 23 and 24, ZHe ages predicted by inverse thermal history modelling were inconsistent with observed values (Supplemental Data C), possibly due to the inability of ZRDAAM to reproduce their relatively large spread in ZHe ages corresponding to a wide range of zircon eU concentrations. Similar discrepancies have been reported for other dispersed ZHe age datasets [Orme *et al.*, 2016; Mackintosh *et al.*, 2017]. Alternatively, these ages may also be influenced by problems arising from intra-grain zonation of U and Th [Anderson *et al.*, 2017]. Whatever the reason, these poorly constrained t-T reconstructions are excluded from further discussion. Instead, thermal history models for samples 23 and 24 using only AFT and AHe data (where available) are used for making geological interpretations.

3.6.1.1 Latest Cretaceous-Paleogene cooling

Basement samples throughout the study area record a period of pronounced latest Cretaceous-Paleogene cooling commencing between ~70-40 Ma. Prior to this, thermal history models allow for a potential period of reheating or residence at elevated temperatures that is poorly constrained by the data.

In general, best fit t-T paths exhibit monotonic latest Cretaceous-Paleogene cooling from elevated temperatures (~200-150 °C) ca. ~70-40 Ma to temperatures between ~60 and 20 °C by the late Oligocene-earliest Miocene (~30-20 Ma). Thermal history models that include ZHe data exhibit latest Cretaceous-Paleogene cooling rates of up to 6.5 °C /Ma. However, the degree and rate of cooling indicated by thermal history models which include ZHe data may be overestimated in some cases due to the lower closure temperature for grains with significant radiation damage [Nasdala *et al.*, 2004; Guenthner *et al.*, 2013].

Time-temperature reconstructions for samples that were modelled without ZHe data (20, 23, 24 and 30) exhibit a consistent period of cooling during the latest Cretaceous-Paleogene, although these are only constrained below the closing temperature of the AFT system (less than ~ 110 °C) resulting in reduced cooling rates ($2.5\text{--}3.3$ °C /Ma). In contrast to the linear latest Cretaceous-Paleogene cooling behaviour illustrated by thermal history models for other samples, the best fit t-T history for sample 24 indicates a two-stage Paleogene cooling phase.

3.6.1.2 Late Oligocene-early Miocene reheating

Beginning in the late Oligocene-early Miocene ($\sim 30\text{--}20$ Ma), thermal history models indicate that the basement underwent a period of mild reheating or near-isothermal residence at temperatures of $\sim 80\text{--}30$ °C. On average, this period is marked by an increase in temperature of $\sim 10\text{--}20$ °C to reach maximum Neogene temperatures of $\sim 50\text{--}70$ °C at around 10–6 Ma. The degree of reheating is quite subtle and very close to the lower limits of the temperature sensitivities of the AFT and AHe systems. Nevertheless, the strong consistency between thermal history models from samples 8, 10, 11, 24 and 30 in both timing and amplitude indicates that this observed reheating phase is most likely real. Although this reheating trend is not well developed in the thermal history models of sample 23, it is within the 95% confidence limits of the time-temperature models and thus not precluded by the data. Neogene reheating is not particularly well supported by the inverse thermal history model for sample 20, which appears to have undergone continued gradual cooling to near-surface temperatures by the present day, although some degree of reheating is permissible even in this case (faint blue paths in Figure 3.7a). However, a forward thermal history model of sample 20 (Figure 3.7b) that involved Paleogene ($\sim 45\text{--}23$ Ma) and Late Miocene-Recent ($\sim 11\text{--}0$ Ma) cooling periods separated by a period of Oligocene-early Miocene reheating reproduced observed AFT and MTL data and the youngest of two AHe single ages. This suggests sample 20 results may be consistent with the t-T reconstruction produced by neighbouring samples.

3.6.1.3 Late Miocene-Recent cooling in the northern Lokichar Fault footwall

In the late Miocene (between 10–6 Ma), basement rocks in the Lokichar Fault footwall transitioned to a period of moderate to rapid cooling ($3\text{--}5$ °C/Ma), displayed by best fit t-T paths in inverse thermal history models for all samples (Figure 7a). This includes samples 23 and 20, although best fit models for these samples suggest cooling at a slightly slower rate

than for other samples. In any case, by the present day all samples had reached near-surface temperatures.

3.7. Tectonic Implications

Low-temperature thermochronology data and thermal history modelling from the North Lokichar Basin reveal a thermal history spanning latest Cretaceous to Recent times. Observed periods of cooling and reheating are interpreted to be associated with a combination of regional tectonism, volcanism and local basin formation processes, discussed below.

3.7.1 Paleo-geothermal gradients in Turkana

When interpreting low-temperature thermal history models to constrain geological processes in the upper crust, estimates of paleo-geothermal gradients must be used to approximate denudation and burial rates [Gallagher and Brown, 1999; Gleadow and Brown, 2000]. Heat flow measurements of the Turkana upper crust however, are scarce. The sole measurement from the Loperot-1 well suggests that the present-day geothermal gradient is elevated in Turkana (42 ± 5 °C/km) [Morley *et al.*, 1999b]. 3D thermal modelling of the Kenyan lithosphere also suggests that surface heat flow in Turkana is high [80-90 mW/m²; Sippel *et al.*, 2017], likely due to a combination of the region's highly attenuated crust (~20 km) [Benoit *et al.*, 2006], and the presence of an elongate mantle thermal anomaly beneath East Africa [e.g. Nyblade, 2011; Hansen *et al.*, 2012].

However, it is unclear when exactly the elevated thermal regime of the Turkana crust developed and how heat flow in the region has changed over time. This stems in large part from the currently poor understanding of the chronology of Turkana crustal and lithospheric attenuation. The increase in ambient heat flow in the upper crust would likely have been gradual following a conductive lag time between the initiation of heating at the base of the crust and the onset of any associated elevated surface heat flow, further complicating paleo-geothermal gradient estimation [Lachenbruch and Sass, 1977; Wheildon *et al.*, 1994; Sippel *et al.*, 2017]. Moreover, heat flow studies in central Kenya have shown that short-wavelength temperature patterns in rifts can vary significantly due to local perturbations caused by magmatic and hydrothermal activity [Nyblade, 1990; Wheildon *et al.*, 1994], both of which are well documented to have affected Turkana throughout much of the late Paleogene-

Pleistocene [e.g. *Zanettin et al.*, 1983; *BEICIP*, 1987; *Renaut et al.*, 2002; *McDougall and Brown*, 2009]. Despite these uncertainties, some first-order temporal constraints can be tentatively made from the tectonic, magmatic and topographic histories of the region. *Hendrie et al.* [1994] concluded that the ~35-40 km of late Eocene?-Oligocene to Recent extension can account for the degree of crustal thinning observed in Turkana. While, the ~45 Ma onset of EARS-related volcanism in NW Turkana and southern Ethiopia [*Zanettin et al.*, 1983; *Ebinger et al.*, 1993, 2000] and the proposed Oligocene-middle Miocene development of the Ethiopian and East African Domes in response to thermal uplift and plume-related magmatic build-up [*Smith and Mosley*, 1993; *Hendrie et al.*, 1994; *Morley*, 1994; *Pik et al.*, 2003; *Wichura et al.*, 2010, 2011; *Moucha and Forte*, 2011; *Pik*, 2011] suggest a late Paleogene arrival of a mantle thermal anomaly beneath East Africa. Therefore, when interpreting time-temperature reconstructions for North Lokichar basement samples, a moderate Cretaceous-early Paleogene geothermal gradient of 25 °C/km is used, consistent with the average geothermal gradient for Archean-Proterozoic mobile belts of East Africa [*Nyblade*, 1990]. For late Paleogene-Recent time, a broader range of geothermal gradient (25-45 °C/km) is employed to allow for spatial and temporal variations in heat flow in Turkana.

3.7.2 Latest Cretaceous-early Paleogene denudation in central Turkana

Basement rocks in the Lokichar Fault footwall, North Lokichar Basin, experienced a period of latest Cretaceous-Paleogene cooling, as recorded by Cretaceous-Eocene ZHe ages, Eocene AFT ages and thermal history modelling results. The presence of Oligo-Miocene volcanoclastics unconformably overlying basement rocks in the North Lokichar Basin [*Boschetto et al.*, 1992; *Morley et al.*, 1999b] suggests that this period of cooling was most likely a result of erosional denudation. This is consistent with other stratigraphic relationships, in particular the Cretaceous?-Paleogene Muruanachok Grits, that outcrop in several isolated hills just west of the study area, are bound by unconformities. The grits overlie Precambrian basement and are, in places, unconformably overlain by late Oligocene-Miocene volcanics [*Walsh and Dodson*, 1969; *Morley et al.*, 1992; *Tiercelin et al.*, 2012a]. Elsewhere, the volcanics lie directly on basement rocks, with the absence of grits explained either by non-deposition or erosion.

A coeval period of Late Cretaceous-early Paleogene cooling of similar magnitude has been observed in multiple AFT studies from southern Ethiopia and throughout much of Kenya and interpreted as reflecting denudation of the Anza Rift flanks and the adjacent hinterlands

[*Foster and Gleadow, 1992, 1993, 1996; Spiegel et al., 2007; Philippon et al., 2014; Torres Acosta et al., 2015; Balestrieri et al., 2016*]. Abundant basement-derived clastic sediments of Upper Cretaceous-lower Paleogene age are preserved in the SE part of the Anza Graben and South Sudanese Rifts [*Schull, 1988; Morley et al., 1999a*], which probably represent material removed during this period of erosion. While the SE Anza and South Sudan rifts were subsiding during the Late Cretaceous-early Paleogene, the North Lokichar region underwent a marked period of denudational cooling (up to 60 °C), as recorded by low-temperature thermochronology (this study). This denudation may well have been extensive in central Turkana, extending across to the NW part of the Anza Graben where it is estimated about 2400 m of Late Cretaceous-Palaeogene section was eroded sometime prior to the Miocene [*Morley et al., 1999a*].

Extensive Late Cretaceous-early Palaeogene exhumation observed in Turkana and the NW part of the Anza Graben seems incompatible with a through-going structural connection between the South Sudanese and Anza Rifts in Turkana, as previously proposed [*Schull, 1988; Ibrahim et al., 1991*]. The occurrence of the 500 m thick Upper Cretaceous-lower Paleogene Lapur Sandstone fluvial sequence in the Lapur Range does, however, indicate some amount of accommodation space was being created at that time, at least locally, in northern Turkana. Some authors [*Wescott et al., 1999; Tiercelin et al., 2012b*] have tentatively correlated the Lapur Sandstone to thick piles of sediment identified in the subsurface of the adjacent northern Turkana, Gatome and Lotikipi Basins, suggesting that Upper Cretaceous-lower Paleogene sedimentation may have been far more extensive laterally (> 100 km E-W) than surface exposures would suggest. Due to the lack of exploration wells and scarcity of subsurface data however, this correlation remains speculative. Nonetheless, Lapur basement rocks record a very limited degree of Late Cretaceous-early Paleogene subsidence for such a well-developed rift, like the Anza [$\beta = \sim 2$; *Reeves et al., 1987; Morley et al., 1999a*] to have extended through this region [*Boone et al., 2018*]. In contrast, the Lapur Sandstone shows no evidence for rift-related faulting in outcrop and has been interpreted as representing fluvial deposits that fed axially into the Anza Rift [*Morley et al., 1992*]. *Vétel and Le Gall* [2006] argued against a connection between the Anza and South Sudan Rifts, pointing out the lack of Cretaceous-early Paleogene deposits in the Kino-Sogo region and the high-gradient isobath pattern of the pre-Cretaceous basement in the hanging wall of the Hoi fault (Figure 3.1). Similarly, the substantial degree of Late Cretaceous-early Paleogene denudational cooling recorded in the North Lokichar region (this study) is consistent with the

Anza Rift having been disconnected from the Cretaceous rifts further west, at least during the Late Cretaceous-Paleogene period. Importantly, thermochronology data presented herein do not constrain the tectonothermal evolution of southern Turkana prior to Late Cretaceous time. As the early stage of rifting in the Anza and South Sudan systems occurred during the Neocomian, Aptian-Albian [Schull, 1988; Bosworth and Morley, 1994; Morley *et al.*, 1999a], it is then possible there was a phase of Early Cretaceous extension in Turkana that was subsequently destroyed by Late Cretaceous-Paleogene exhumation and erosion. Indeed, the 2,400 m of removed Late Cretaceous-Paleogene section in the NW Anza Graben (Chalbi Desert) indicated by vitrinite reflectance data [Morley *et al.*, 1999a] is consistent with such an interpretation. However, no evidence currently exists to suggest that Early Cretaceous subsidence also occurred on the west side of Lake Turkana. Instead, thermochronology data from the Lapur Range [Boone *et al.*, 2018] record pronounced Early Cretaceous denudational cooling, consistent with the Anza Rift terminating to the east of Lake Turkana.

The observations regarding the limited extent of early rifting in Northern Turkana challenges the often invoked assumption that the continuation of the Cretaceous-early Paleogene Anza-South Sudan rift system through Turkana is responsible for the region's anomalous crustal architecture, atypical Cenozoic tectonic evolution with respect to the greater EARS and its present-day morphology [e.g. Ebinger *et al.*, 2000; Benoit *et al.*, 2006; Fishwick and Bastow, 2011; Brune *et al.*, 2017]. Indeed, the highly attenuated crust of Turkana and its much greater degree of total extension [35-40 km; Morley *et al.*, 1992; Morley, 2002] compared to the southern Ethiopian [< 25 km; Ebinger *et al.*, 1993] and Kenyan Rifts [8-10 km; Morley *et al.*, 1992] do require an earlier period of Turkana rifting predating the Neogene onset of major extension elsewhere in the EARS to avoid significant spatial problems [Hendrie *et al.*, 1994]. However, our findings suggest that later Paleogene tectonism may have played a more significant role in modifying the Turkana crust than previously thought.

3.7.3 Continued Paleogene denudation in Turkana

Thermal history models presented in previous studies from the Anza Rift shoulders and adjacent hinterlands of central and eastern Kenya and southern Ethiopia record a pronounced reduction in cooling ca. 50 Ma (see references above). By contrast, time-temperature reconstructions from the North Lokichar region (this study) suggest that the central Turkana basement continued cooling until much later into the Oligocene (~30 Ma). The degree of continued Paleogene cooling is substantial, with northern transect t-T reconstructions

illustrating up to 120 °C of cooling between ~50-30 Ma (Figure 3.7). Southern transect thermal history models similarly suggest significant, albeit slightly less, post 50 Ma cooling (~80-90 °C). Although, the difference between observed Paleogene cooling in northern and southern North Lokichar basement rocks may simply stem from the exclusion of data from the higher-temperature ZHe system for time-temperature modelling of southern transect samples (Section 3.6.1).

It is possible that the concurrent formation of fault-bounded depocentres in the neighbouring Lokichar Basin (upper Eocene?-Oligocene to Miocene) and possibly in the poorly understood Muruanachok Basin (Late Cretaceous?-Oligocene) may have resulted in a degree of erosion of the intermediate North Lokichar area. However, there is perhaps a problem of timing. Seismic reflection data suggest that the major expansion of Lokichar Basin syn-rift infill did not occur until later in late Oligocene-early Miocene times [Morley, 1999b; Morley *et al.*, 1999b] and fault plane analysis in the undated Muruanachok Grits indicates NE-SW extension compatible with the earlier Cretaceous paleostress field [Vétel and Le Gall, 2006]. In addition, paleocurrent indicators from the Lokichar Basin, albeit only observed in late Oligocene-early Lokichar Basin strata, are consistent with material being transported from the south and southeast [Morley *et al.*, 1999b; Tiercelin *et al.*, 2004]. Regardless, the considerable magnitude of middle Eocene-early Oligocene cooling (~120 °C to 80 °C, equivalent to ~2-5 km of denudation) recorded over a wide region of central Turkana (~1600 km²) and the lack of major Eocene-early Oligocene structures proximal to basement samples suggest a more regional causal mechanism.

One possible explanation for continued denudational cooling in central Turkana between ~50-30 Ma may be isostatic and/or dynamic uplift in response to the impingement or passing of a mantle plume beneath Turkana at that time, as predicted by some plume migration models [e.g. Morley, 1994, 2002; George *et al.*, 1998; Furman *et al.*, 2006; Rogers, 2006; Furman, 2007]. Thermal uplift above such a thermal anomaly has been invoked to have possibly caused inversion in the Upper Cretaceous(?)-Paleogene Muruanachok Basin, immediately west of the North Lokichar Basin [Morley *et al.*, 1999b], and along the Turkwell escarpment bounding the western Turkana margin [Morley *et al.*, 1992] (Figures 3.1 and 3.2). Similarly, Tiercelin *et al.* [2012b] noted the synchronism between the abrupt termination of Lapur Sandstone deposition and the onset of volcanism in southern Ethiopia and northern Turkana between 45-35 Ma, suggesting this may be related to uplift associated with the

arrival of a mantle plume. The impingement of such an asthenospheric upwelling beneath the Ethiopian lithosphere is considered responsible for the late Eocene-Oligocene onset of doming of the Ethiopian Plateau [*Ebinger and Sleep*, 1998; *Pik et al.*, 2003; *Sembroni et al.*, 2016]. Similarly, uplift above an asthenospheric thermal anomaly is thought to have caused the formation of the East African Dome during Miocene-Recent times [*Pik*, 2011; *Wichura et al.*, 2011]. In contrast with the Ethiopian and East African Plateaux however, strong supportive evidence for regional domal uplift in Turkana does not currently exist [*Vétel and Le Gall*, 2006]. Though, this may simply stem from our poor understanding of regional paleodrainage patterns during the Paleogene, scarcity of well data from Paleogene-aged basins in the greater Turkana area and the inherent difficulty in constraining long-wavelength paleotopography [*Pik*, 2011; *Olen et al.*, 2012].

Benoit et al. [2006] showed that the present-day 1.8 km of differential elevation between the Turkana Depression and the surrounding domes can be attributed to the inability of the highly attenuated Turkana crust to support regional scale topography. However, if significant crustal thinning began later in the Paleogene as opposed to in the Cretaceous as previously assumed (Section 3.7.2), Turkana may have been able to isostatically support a degree of domal topography at that time. Even despite its relatively subdued modern-day topography, numerical modelling suggests that the Turkana region has experienced a >1 km increase in dynamic topography since ca. 30 Ma [*Moucha and Forte*, 2011]. Thus, isostatic uplift and/or the development of dynamic topography in Turkana associated with an asthenospheric thermal anomaly are conceivable during the Paleogene. Nevertheless, without an improved understanding of the Paleogene crustal and lithospheric structure of Turkana and the geographic extent of denudational cooling and sedimentation at that time, such a scenario remains hypothetical.

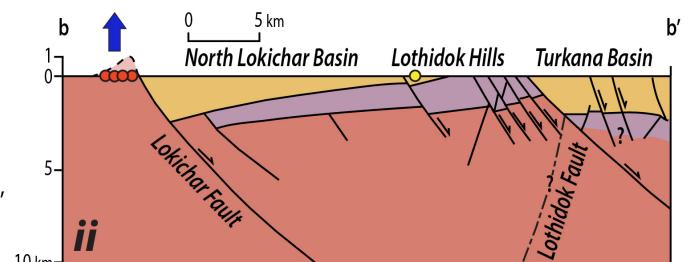
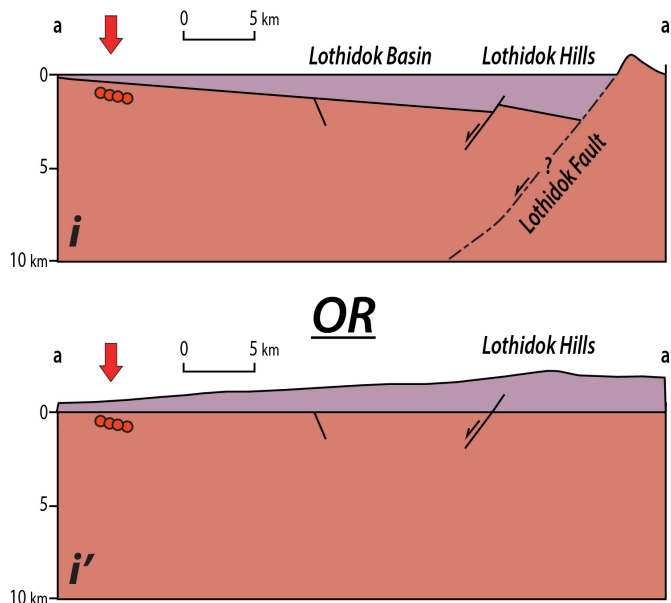
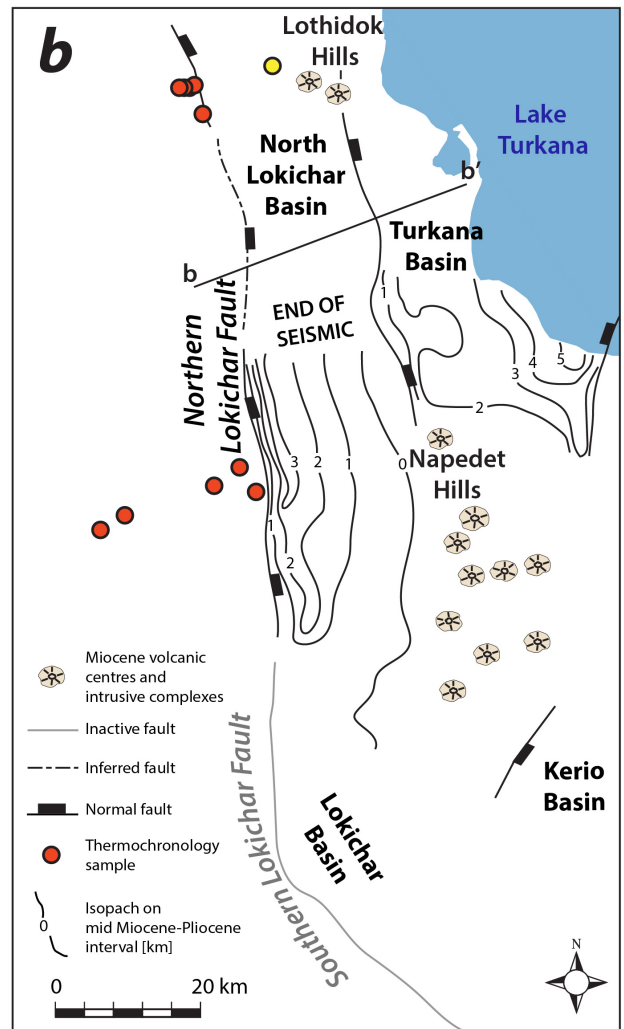
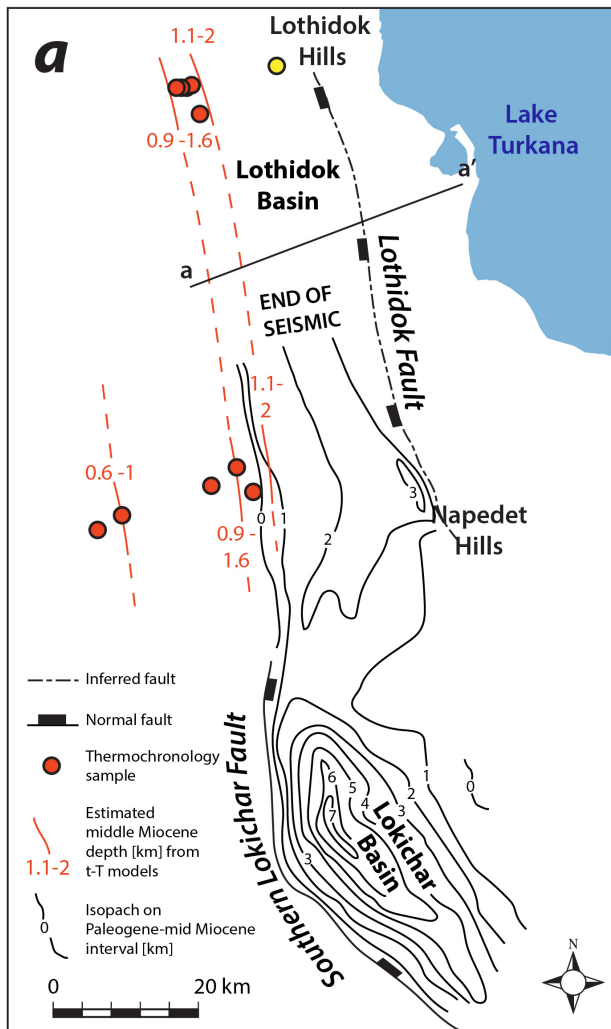


Figure 8 (Previous page). *Isopach maps and schematic cross-sections of Lothidok and North Lokichar Basin evolution for (a) Paleogene-middle Miocene and (b) middle Miocene-Pliocene times. Isopachs are based on seismic reflection data [Morley et al., 1992, 1999b]. Estimated middle Miocene depths of basement samples (red circles) are determined from their corresponding thermal history models (Figure 3.7a), using the maximum middle Miocene paleotemperature of their best fit t-T path, a paleo-surface temperature of 20 °C and a geothermal gradient range of 25-45 °C/km (see Section 3.7.1). The yellow circle represents sample 21, an upper Miocene-Pliocene Turkwel Bed cobble conglomerate. The location of outcropping middle Miocene volcanic centres and igneous intrusives in the Napedet Hills are shown [after Morley, 1999b; Vétel and Le Gall, 2006]. Schematic cross sections, based on seismic line TVK-3 (see Figure 3.2), illustrate the interpreted evolution of the Lothidok-North Lokichar Basin and the relative position of northern transect basement rocks sampled (colour code for geological units as for Figure 3.1). Arrows above cross sections indicate thermal state of basement rocks. Thermal history models suggest that basement samples experienced a period of late Oligocene-middle Miocene reheating (red arrow – left panel) interpreted as a result of burial beneath ~0.2-0.8 km of overburden during the development of the west-facing Lothidok Basin half-graben (cross-section i). However, late Oligocene-middle Miocene reheating due to burial beneath volcanic flows without the creation of accommodation space by normal faulting cannot be excluded (cross-section i'; see Section 3.7.3 for further discussion). Following middle Miocene change in basin polarity (cross-section ii), basement samples began to cool (blue arrow – right panel) due to isostatic uplift of the Lokichar Fault footwall and erosion during formation of the middle Miocene-Pliocene North Lokichar Basin.*

3.7.4 Oligocene-middle Miocene reheating

Beginning between ~30-20 Ma, AFT and AHe data and thermal history models indicate that basement rocks in the study area went through a period of gradual reheating that may have lasted until the late Miocene. By this time, basement rock samples reached temperatures of between 40 °C and 70 °C (Figure 3.7a), with maximum middle Miocene paleo-temperatures increasing eastwards (Figure 3.8a).

The Oligocene-middle Miocene also marked a period of significant magmatic activity across much of Turkana (Figures 3.1 and 3.2) [McDougall and Brown, 2009]. More locally in the Lothidok Hills area and the region north of the Muruanachok Hills, at least 20 individual basalt flows ranging from 4 to 60 m thick were emplaced during late Oligocene-Miocene times (Figure 3.2) [Boschetto et al., 1992; Morley et al., 1999b; McDougall and Brown, 2009]. Later in the middle Miocene, a wide range of igneous rocks were emplaced in the Napedet Hills further south, including lava flows, dike swarms, vent complexes and pyroclastics [Morley et al., 1999b].

In certain cases, the emplacement of intrusions can reheat surrounding or underlying country rock to adequate temperatures to reset low-temperature thermochronometers [e.g. Tagami

and Shimada, 1996; Reiners, 2005]. However, heat transfer of this kind occurs primarily via conduction [Philpotts and Ague, 1990; Spear, 1993] and is thus dependent on time, the geometry of the magmatic body and the distance from the intrusion [Ehlers, 2005]. For instance, country rock located 50 m from a 100 m thick intrusion with an initial temperature of 700 °C will re-equilibrate back to its initial background temperature within 1 Ma. However, an increase in the distance between the country rock in question and the intrusion to just 100 m, results in a drastically reduced thermal equilibration time of < 20 years [Ehlers, 2005]. As the nearest mapped Oligocene-middle Miocene eruptive or intrusive centres are located in the Lothidok and Napedet Hills and north of the Muruanachok Hills (Figure 3.2), >15 km away from any basement sample, we would argue that conductive heat transfer from magmatic intrusions cannot explain the ~30-15 Ma reheating phase recorded by low-temperature thermochronology.

The thermal effect of lava flows on low-temperature thermochronometers is similarly limited by proximity, occurring over a similar length scale to individual flow thickness [Gallagher *et al.*, 1994; Gunnell *et al.*, 2003]. He diffusion experiments have shown that the susceptibility of the low temperature sensitivity AHe system to transitory thermal events, such as lava flows, is limited to just the outer 3 cm of an exposed rock [Wolf *et al.*, 1998; Mitchell and Reiners, 2003; Reiners *et al.*, 2007]. In the case of the North Lokichar region, thermal history models suggest that prior to the onset of Oligocene-middle Miocene reheating basement samples resided at temperatures between 30 °C and 60 °C, corresponding to paleodepths of between ~200 m to 1600 m. Thus, basement samples would have been shielded from conductive heat transfer from Oligocene-middle Miocene lava flows by hundreds of metres of overlying strata.

Furthermore, in studies where transitory heating mechanisms, such as lava flows [e.g. Gallagher *et al.*, 1994; Gunnell *et al.*, 2003], hydrothermal fluids [e.g. Seiler *et al.*, 2009] or heated groundwater [Boone *et al.*, 2016], have been invoked to explain the rejuvenation of low-temperature thermochronology systems in basement rocks, observed reheating patterns are highly localised and, despite their close proximity, are characterised by significantly reset samples being interspersed with unaffected samples. By contrast, basement samples from the North Lokichar region, spread out over an ~1600 km² area, consistently record a period of Oligocene-middle Miocene reheating. Thus, a more long-wavelength heating mechanism must be inferred.

A more plausible explanation for the observed reheating episode is an increase in temperature due to burial and the associated downward advection of basement rock samples through to higher isotherms [Ehlers *et al.*, 2001; Ehlers, 2005]. Such a scenario is consistent with the sequence of Kalakol Basalts (28-18 Ma) and the Lothidok Formation (18-13 Ma) preserved in the proximal Lokichar Fault hanging wall (lines TVK-3 and TVK-102W; Figure 3.2) [Morley, 1999b]. The so-called Lothidok Basin reaches a maximum thickness of up to ~1400 m beneath the Lothidok Hills [Morley, 1999b].

Morley *et al.* [1999b] suggested that subsidence and sedimentation in the Lothidok Basin were likely controlled by a west-dipping, basin-bounding normal fault buried beneath the Lothidok and Napadet Hills (cross-section i; Figure 3.8), consistent with the east-thickening wedge shape of the basal package (TVK-3; Figure 3.2). While most of the faulting observed in the Lothidok Hills is thought to postdate the middle Miocene, Boschetto [1988] identified at least one west-dipping fault that predates the deposition of the Lothidok Formation. A Kalakol Basalt flow dated at 17.7 Ma overlies the fault and thus post-dates it. However, due to the volcanic-dominated nature of the section, surface topography and high intensity faulting, seismic data quality in the critical Lothidok Hills area was poor (TVK-3, Figure 3.2), preventing identification of the proposed Lothidok Fault [Morley, 1999b; Morley *et al.*, 1999b]. If the Kalakol Basalts and Lothidok Formation were indeed deposited within a west-facing half-graben, basement samples from this study would have been located in the shallow sub-surface of the Lothidok Basin flexural margin (Figure 3.8a). Thus, the relatively small degree of late Oligocene-early Miocene reheating (~10-20 °C, equivalent to ~0.2-0.8 km of burial) recorded in thermal history models seems reasonable, as these basement samples would have resided ≥ 15 km west of the basin-bounding fault where the total strata thickness would have been less (Figure 8a). Even the westernmost thermal history model (sample 24) experienced a significant degree of Oligocene-early Miocene reheating (~20 °C) despite being ~50 km west of the proposed west-dipping Lothidok Fault (Figure 3.8a). This would suggest that either a significant thickness of Lothidok Basin strata (Kalakol Basalt flows?) extended over the flexural margin into the Turkwell Trough or that the Lothidok Basin extended west beyond the location of the southern Lokichar Fault. This fault forms the western boundary of the Lokichar Basin to the south, which is infilled by an Eocene-lower Miocene fluvio-lacustrine sequence. The Oligocene-lower Miocene part of the succession has

a maximum thickness of ~1350 m, which is comparable to the amount of section deposited in the Lothidok basin during this time interval [Morley *et al.*, 1999b].

However, since much of the late Oligocene-middle Miocene Lothidok strata consists of basalt flows sourced from the igneous provinces situated to the northwest and northeast of the basin [Tiercelin *et al.*, 2004], it is conceivable that little or no accommodation space creation was required to facilitate the emplacement of these lavas (cross-section i'; Figure 3.8). The subsequent deposition of 580 m of overlying Lothidok Formation is, however, more difficult to explain without faulting. While the development of volcanic topography during the build-up of a thick (≥ 785 m) [Boschetto *et al.*, 1992] succession of Kalakol Basalt flows may have created limited amounts of accommodation space locally, it seems improbable that it would be sufficient to facilitate the subsequent deposition of a comparable thickness of sediment.

3.7.5 Middle Miocene basin reorganization

At some point during the middle to late Miocene, the northern Lokichar Fault developed immediately east of northern and southern sampling transects. The Lothidok Hills area experienced further subsidence in the hanging wall of the Lokichar fault, resulting in the formation of the North Lokichar Basin, a westward dipping half graben that contains a sequence of late Miocene-Pliocene growth strata (cross-section ii; Figure 3.8) [Boschetto *et al.*, 1992; Morley *et al.*, 1999b]. As a result, basement samples, now located in the western margin of the North Lokichar Basin, experienced rapid late Miocene-Pliocene cooling, as recorded by low-temperature thermochronology data and thermal history models. This is interpreted as representing denudation associated with flexural isostatic footwall uplift and late Miocene-Pliocene extension along the northern Lokichar Fault (Figure 3.8b).

Thermal history modelling indicates the Lokichar Fault footwall experienced between 30 and 50 °C of Late Miocene-Pliocene cooling. Assuming the elevated Turkana geothermal gradient [45 °C/km; Morley *et al.*, 1999b] was established prior to the late Miocene (Section 3.7.1), this degree of cooling is equivalent to between ~0.7 and 1.1 km of late Miocene-Pliocene denudation. These estimates are approximately equal to the amount of earlier subsidence, providing a possible explanation as to why very limited Lokichar Fault footwall topography has been developed.

If earlier Lothidok Basin deposition was governed by the development of a west-facing half-graben as previously suggested [Morley, 1999b; Morley *et al.*, 1999b], the middle Miocene (ca. 13 Ma) initiation of the east-facing North Lokichar Basin marks an important change in basin polarity. This period of basin reorganization and northward propagation of the Lokichar Fault coincided with the development of the extensive Napedet Hills volcanic centres between ~15 and 12 Ma [Morley, 1999b], leading Morley [1999a] to suggest a possible link between magma intrusion and strain migration. Regardless, the concurrent northwards migration of the Lokichar Fault and cessation of Lokichar Basin development in the late Miocene [Morley *et al.*, 1999b; Vétel *et al.*, 2004] signals a northward migration of strain in the Turkana Rift. This is consistent with rift propagation models that predict an early nucleation of deformation in the Turkana Depression, followed by a northward propagation of faulting into southern Ethiopia [Wolfenden *et al.*, 2004; Bonini *et al.*, 2005; Rooney *et al.*, 2007].

Following deposition of the Turkwel Beds, late Miocene-Pliocene extension occurred along east-dipping faults bounding the neighbouring Turkana and Kerio basins to the east [Morley *et al.*, 1999b]. The resulting uplift and erosion of the Lothidok and Napedet Hills exposed North Lokichar Basin infill, including the Turkwel Beds and sample 21. The retention of primarily pre-Neogene single grain AFT ages (> 80% of single grains; Figure 3.3a) suggests that sample 21 already resided at low temperatures, either below the PAZ or within its lower range, during that time. Only the youngest AFT age population (centred around 23 ± 3 Ma) may record this period of cooling. Furthermore, middle Miocene AHe ages (14.4 ± 0.9 to 17.7 ± 1.1 Ma; 4 grains) indicate that the maximum Neogene paleotemperature for sample 21 was well within the temperature range of the AHe PRZ (~80-30 °C).

3.8. Conclusions

In this study, low-temperature thermochronology data (ZHe, AFT and AHe) and thermal history modelling results are used to constrain the development of the footwall region of the North Lokichar Basin. The data provide important information impacting on both our understanding of pre-EARS tectonism in the Turkana Depression, and late Paleogene-Recent rift propagation within the region.

The earliest recorded history of the footwall margin indicates a period of erosion and exhumation during the latest Cretaceous-Early Palaeogene and further corroborates previous studies that have determined extensive cooling across the region of the Kenya rift during this period. This cooling is interpreted as being related to widespread erosion during Cretaceous rifting and transport of sediment into the central and SE Anza Graben. Pronounced exhumation of basement rocks throughout the North Lokichar area, uplift and erosion of the NW Anza Graben (Chalbi Desert region) together with the scarcity of concurrent Late Cretaceous-Early Paleogene subsidence recorded elsewhere in the region challenge the, often invoked, assumption that the Anza Rift once extended northwest through Turkana to connect with the Cretaceous rifts of South Sudan at least for the Late Cretaceous-Palaeogene period. Erosion may well have been episodic, but this episodicity cannot be demonstrated by the data. However, the presence of the eroded Muruanachok Grits (probably predating Lothidok Basin strata) suggests at least one period of subsidence during the Late Cretaceous-Palaeogene period where uplift has been modelled. Possibly mid-late Palaeogene exhumation marks the impact of a mantle thermal anomaly in the region that triggered the extensive latest Eocene-Oligocene volcanism, which represents the onset of the East African Rift tectonic activity in Northern Kenya and Southern Ethiopia.

Low-temperature thermochronology data record a period of reheating between ~30-10 Ma, interpreted as burial beneath ~0.2-0.8 km of overburden. A similarly thick (up to 1400 m) succession of Kalakol Basalts (28-18 Ma) and Lothidok Formation clastics (18-13 Ma) are preserved just east of the study area, where they form an east-thickening, wedge shaped package in the basal North Lokichar Basin. These are thought to have been deposited during the Oligocene-middle Miocene development of the Lothidok Basin, a west facing half-graben bounded by the inferred Lothidok normal fault. Thus, observed reheating is interpreted as representing burial of the Lothidok Basin flexural margin, where basement samples would have resided at that time. Due to the obscuring effect of significant surface volcanism in the critical Lothidok Hills area however, the theorised west-dipping Lothidok master fault has yet to be identified in the poor quality seismic reflection data. Therefore, an alternative scenario of reheating due to burial beneath late Oligocene-middle Miocene volcanic flows without the creation of accommodation space by normal faulting cannot be excluded.

Published interpretations of seismic reflection data indicate that the late Miocene (~9 Ma) is marked by an important period of basin reorganisation in southern Turkana, including the

northward migration of the Lokichar Fault into the Lothidok region. Subsequent late Miocene-Pliocene sedimentation in the North Lokichar Basin was controlled by the northern extension of the east-dipping Lokichar Fault. The scenario indicates that basement samples, previously located in what was believed to be the Lothidok Basin hinge margin, subsequently resided in the footwall of the Lokichar Fault. The thermochronology data reported here demonstrate that the footwall area experienced an ~10-3 Ma period of relatively rapid cooling of ~30°-50° due to isostatic uplift and erosion of ~0.7-1.1 km in response to late Miocene-Pliocene extension (~5 km) focussed on the North Lokichar Basin boundary fault. In the region of the southern sampling transect, thermochronology data indicates that the low-angle fault segment linking the North Lokichar Basin to the adjacent late Eocene-middle Miocene Lokichar Basin further south was active from about 10 Ma.

References

- Anderson, A. J., K. V. Hodges, and M. C. van Soest (2017), Empirical constraints on the effects of radiation damage on helium diffusion in zircon, *Geochim. Cosmochim. Acta*, 218, 308–322, doi:10.1016/j.gca.2017.09.006.
- Balestrieri, M. L., M. Bonini, G. Corti, F. Sani, and M. Philippon (2016), A refinement of the chronology of rift-related faulting in the Broadly Rifted Zone, southern Ethiopia, through apatite fission-track analysis, *Tectonophysics*, 671, 42–55, doi:10.1016/j.tecto.2016.01.012.
- Barbarand, J., A. Carter, I. Wood, and T. Hurford (2003), Compositional and structural control of fission-track annealing in apatite, *Chem. Geol.*, 198(1–2), 107–137, doi:10.1016/S0009-2541(02)00424-2.
- Bauer, F. U., U. A. Glasmacher, U. Ring, M. Karl, A. Schumann, and B. Nagudi (2013), Tracing the exhumation history of the Rwenzori Mountains, Albertine Rift, Uganda, using low-temperature thermochronology, *Tectonophysics*, 599, 8–28, doi:10.1016/j.tecto.2013.03.032.
- van der Beek, P., S. Cloetingh, and P. Andriessen (1994), Mechanisms of extensional basin formation and vertical motions at rift flanks: Constraints from tectonic modelling and fission-track thermochronology, *Earth Planet. Sci. Lett.*, 121, 417–433.
- van der Beek, P., E. Mbede, P. Andriessen, and D. Delvaux (1998), Denudation history of the Malawi and Rukwa Rift flanks (East African Rift system) from apatite fission track thermochronology, *J. African Earth Sci.*, 26(3), 363–385, doi:10.1016/S0899-5362(98)00021-9.
- BEICIP (1987), Geological Map of Kenya,
- Bellieni, G., E. Justin Visentin, B. Zanettin, E. M. Piccirillo, F. Radicati di Brozolo, and F. Rita (1981), Oligocene transitional tholeiitic magmatism in Northern Turkana (Kenya): Comparison with the Coeval Ethiopian volcanism, *Bull. Volcanol.*, 44(3), 411–427, doi:10.1007/BF02600573.
- Benoit, M. H., A. A. Nyblade, and M. E. Pasyanos (2006), Crustal thinning between the Ethiopian and East African plateaus from modeling Rayleigh wave dispersion, *Geophys. Res. Lett.*, doi:10.1029/2006GL025687.
- Beucher, R., R. W. Brown, S. Roper, F. Stuart, and C. Persano (2013), Natural age dispersion arising from the analysis of broken crystals: Part II. Practical application to apatite (U–Th)/He thermochronometry, *Geochim. Cosmochim. Acta*, 120, 395–416, doi:10.1016/j.gca.2013.05.042.
- Biswas, S., I. Coutand, D. Grujic, C. Hager, D. Stockli, and B. Grasemann (2007), Exhumation and uplift of the Shillong plateau and its influence on the eastern Himalayas: New constraints from apatite and zircon (U–Th–[Sm])/He and apatite fission track analyses, *Tectonics*, 26, doi:10.1029/2007TC002125.
- Bonini, M., G. Corti, F. Innocenti, P. Manetti, F. Mazzarini, T. Abebe, and Z. Pecskey (2005), Evolution of the Main Ethiopian Rift in the frame of Afar and Kenya rifts propagation, *Tectonics*, 24(1), n/a–n/a, doi:10.1029/2004TC001680.
- Boone, S. C., C. Seiler, A. J. Reid, B. Kohn, and A. Gleadow (2016), An Upper Cretaceous paleo-aquifer system in the Eromanga Basin of the central Gawler Craton, South Australia: evidence from apatite fission track thermochronology, *Aust. J. Earth Sci.*, 63(3), 315–331, doi:10.1080/08120099.2016.1199050.
- Boone, S. C., C. Seiler, B. P. Kohn, A. J. W. Gleadow, D. A. Foster, and L. Chung (2018), Influence of Rift Superposition on Lithospheric Response to East African Rift System Extension: Lapur Range, Turkana, Kenya, *Tectonics*, 37, 1–26,

doi:10.1002/2017TC004575.

- Boschetto, H. B. (1988), Geology of the Lothidok Range, Northern Kenya, The University of Utah.
- Boschetto, H. B., F. H. Brown, and I. McDougall (1992), Stratigraphy of the Lothidok Range, northern Kenya, and K/Ar ages of its Miocene primates, *J. Hum. Evol.*, 22(1), 47–71, doi:10.1016/0047-2484(92)90029-9.
- Bosworth, W. (1992), Mesozoic and early Tertiary rift tectonics in East Africa, *Tectonophysics*, 209(1–4), 115–137, doi:10.1016/0040-1951(92)90014-W.
- Bosworth, W., and C. K. Morley (1994), Structural and stratigraphic evolution of the Anza rift, Kenya, *Tectonophysics*, 236(1–4), 93–115, doi:10.1016/0040-1951(94)90171-6.
- Braun, J. (2005), Quantitative Constraints on the Rate of Landform Evolution Derived from Low-Temperature Thermochronology, *Rev. Mineral. Geochemistry*, 58, 351–374, doi:10.2138/rmg.2005.58.13.
- Brown, R. W., R. Beucher, S. Roper, C. Persano, F. Stuart, and P. Fitzgerald (2013), Natural age dispersion arising from the analysis of broken crystals. Part I: Theoretical basis and implications for the apatite (U-Th)/He thermochronometer, *Geochim. Cosmochim. Acta*, 122(120), 478–497, doi:10.1016/j.gca.2013.05.041.
- Browne, S. E., and J. D. Fairhead (1983), Gravity study of the Central African Rift system: A model of continental disruption 1. The Ngaoundere and Abu Gabra Rifts, *Tectonophysics*, 94(1–4), 187–203, doi:10.1016/0040-1951(83)90016-1.
- Brune, S., G. Corti, and G. Ranalli (2017), Controls of inherited lithospheric heterogeneity on rift linkage: Numerical and analog models of interaction between the Kenyan and Ethiopian rifts across the Turkana depression, *Tectonics*, 1–30, doi:10.1002/2017tc004739.
- Corti, G. (2009), Continental rift evolution: From rift initiation to incipient break-up in the Main Ethiopian Rift, East Africa, *Earth-Science Rev.*, 96(1–2), 1–53, doi:10.1016/j.earscirev.2009.06.005.
- Danišik, M., R. F. Sachsenhofer, V. A. Privalov, E. A. Panova, W. Frisch, and C. Spiegel (2008), Low-temperature thermal evolution of the Azov Massif (Ukrainian Shield–Ukraine) — Implications for interpreting (U–Th)/He and fission track ages from cratons, *Tectonophysics*, 456, 171–179, doi:10.1016/j.tecto.2008.04.022.
- Danišik, M., B. I. A. McInnes, C. L. Kirkland, B. J. McDonald, N. J. Evans, and T. Becker (2017), Seeing is believing: Visualization of He distribution in zircon and implications for thermal history reconstruction on single crystals, *Sci. Adv.*, 3, 1–9.
- Davidson, A. (1983), *The Omo River Project: reconnaissance geology and geochemistry of parts of Ilubabor, Kefa, Gemu Gofa and Sidamo, Ethiopia*.
- Dindi, E. W. (1994), Crustal structure of the Anza graben from gravity and magnetic investigations, *Tectonophysics*, 236(1–4), 359–371, doi:10.1016/0040-1951(94)90184-8.
- Donelick, R. A., P. B. O’Sullivan, and R. A. Ketcham (2005), Apatite Fission-Track Analysis, *Rev. Mineral. Geochemistry*, 58(1), 49–94, doi:10.2138/rmg.2005.58.3.
- Ebinger, C. J., and A. Ibrahim (1994), Multiple episodes of rifting in Central and East Africa: A re-evaluation of gravity data, *Geol. Rundschau*, 83(4), 689–702, doi:10.1007/BF00251068.
- Ebinger, C. J., and N. H. Sleep (1998), Cenozoic magnetism throughout east Africa resulting from impact of a single plume, *Nature*, 395, 788–791.
- Ebinger, C. J., T. Yemane, G. Woldegabriel, J. L. Aronson, and R. C. Walter (1993), Late Eocene-Recent volcanism and faulting in the southern main Ethiopian rift, *J. Geol. Soc. London.*, 150(1), 99–108, doi:10.1144/gsjgs.150.1.0099.
- Ebinger, C. J., T. Yemane, D. J. Harding, S. Tesfaye, S. Kelley, and D. C. Rex (2000), Rift deflection, migration, and propagation: Linkage of the Ethiopian and Eastern rifts,

- Africa, *GSA Bull.*, 112(2), 163–176.
- Ehlers, T. a. (2005), Crustal Thermal Processes and the Interpretation of Thermochronometer Data, *Rev. Mineral. Geochemistry*, 58(August), 315–350, doi:10.2138/rmg.2005.58.12.
- Ehlers, T. A., P. A. Armstrong, and D. S. Chapman (2001), Normal fault thermal regimes and the interpretation of low-temperature thermochronometers, *Phys. Earth Planet. Inter.*, 126, 179–194.
- Farley, K. A. (2000), Helium diffusion from apatite: General behavior as illustrated by Durango fluorapatite, *J. Geophys. Res.*, 105(B2), 2903–2914.
- Farley, K. A., and D. F. Stockli (2002), (U-Th)/He Dating of Phosphates: Apatite, Monazite, and Xenotime, *Rev. Mineral. Geochemistry*, 48(1), 559–577, doi:10.2138/rmg.2002.48.15.
- Farley, K. A., R. A. Wolf, and L. T. Silver (1996), The effects of long alpha-stopping distances on (U-Th)/He ages, *Geochim. Cosmochim. Acta*, 60(21), 4223–4229, doi:10.1016/S0016-7037(96)00193-7.
- Fishwick, S., and I. D. Bastow (2011), Towards a better understanding of African topography: A review of passive-source seismic studies of the African crust and upper mantle, *Geol. Soc. London, Spec. Publ.*, 357, 343–371, doi:10.1144/SP357.19.
- Fitzgerald, P. G. (1992), The Transantarctic Mountains of Southern Victoria Land: The Application of Apatite Fission Track Analysis to a Rift Shoulder Uplift, *Tectonics*, 11(3), 634–662.
- Fitzgerald, P. G., S. L. Baldwin, L. E. Webb, and P. B. O’Sullivan (2006), Interpretation of (U-Th)/He single grain ages from slowly cooled crustal terranes: A case study from the Transantarctic Mountains of southern Victoria Land, *Chem. Geol.*, 225(1–2), 91–120, doi:10.1016/j.chemgeo.2005.09.001.
- Fleischer, R., P. Price, and R. Walker (1975), *Nuclear Tracks in Solids: principles and applications*, University of California Press, Berkeley.
- Flowers, R. M., and S. A. Kelley (2011), Interpreting data dispersion and “inverted” dates in apatite (U–Th)/He and fission-track datasets: An example from the US midcontinent, *Geochim. Cosmochim. Acta*, 75(18), 5169–5186, doi:10.1016/j.gca.2011.06.016.
- Foster, D. A., and A. J. W. Gleadow (1992), The morphotectonic evolution of rift-margin mountains in central Kenya: Constraints from apatite fission-track thermochronology, *Earth Planet. Sci. Lett.*, 113(1–2), 157–171, doi:10.1016/0012-821X(92)90217-J.
- Foster, D. A., and A. J. W. Gleadow (1993), Episodic Denudation in East Africa: A Legacy of Intracontinental Tectonism, *Geophys. Res. Lett.*, 20(21), 2395–2398.
- Foster, D. A., and J. W. Gleadow (1996), Structural framework and denudation history of the flanks of the Kenya and Anza Rifts, East Africa, *Tectonics*, 15(2), 258–271, doi:10.1029/95TC02744.
- Furman, T. (2007), Geochemistry of East African Rift basalts: An overview, *J. African Earth Sci.*, 48, 147–160, doi:10.1016/j.jafrearsci.2006.06.009.
- Furman, T., K. M. Kaleta, J. G. Bryce, and B. B. Hanan (2006), Tertiary Mafic Lavas of Turkana, Kenya: Constraints on East African Plume Structure and the Occurrence of High- μ Volcanism in Africa, *J. Petrol.*, 47(6), 1221–1244, doi:10.1093/petrology/egl009.
- Galbraith, R. F. (1990), The Radial Plot: Graphical Assessment of Spread in Ages, *Int. J. Radiat. Appl. Instrumentation, Part D, Nucl. Tracks Radiat. Meas.*, 17(3), 207–214.
- Galbraith, R.F., 2005. Statistics for fission track analysis. Taylor & Francis Group, Boca Raton.
- Gallagher, K. (1995), Evolving temperature histories from apatite fission-track data, *Earth Planet. Sci. Lett.*, 136(3–4), 421–435.
- Gallagher, K. (2012), Transdimensional inverse thermal history modeling for quantitative

- thermochronology, *J. Geophys. Res. Solid Earth*, 117, doi:10.1029/2011JB008825.
- Gallagher, K., and R. Brown (1997), The onshore record of passive margin evolution, *J. Geol. Soc. London*, 154, 451–457, doi:10.1144/gsjgs.154.3.0451.
- Gallagher, K., C. J. Hawkesworth, and M. S. M. Mantovani (1994), The denudation history of the onshore continental margin of SE Brazil inferred from apatite fission track data, *J. Geophys. Res.*, 99(B9), 18,117–18,145.
- Gallagher, K., K. Charvin, S. Nielsen, M. Sambridge, and J. Stephenson (2009), Markov chain Monte Carlo (MCMC) sampling methods to determine optimal models, model resolution and model choice for Earth Science problems, *Mar. Pet. Geol.*, 26, 525–535, doi:10.1016/j.marpetgeo.2009.01.003.
- Gautheron, C., L. Tassan-Got, J. Barbarand, and M. Pagel (2009), Effect of alpha-damage annealing on apatite (U-Th)/He thermochronology, *Chem. Geol.*, 266, 157–170, doi:10.1016/j.chemgeo.2009.06.001.
- Gautheron, C., J. Barbarand, R. A. Ketcham, L. Tassan-Got, P. van der Beek, M. Pagel, R. Pinna-Jamme, F. Couffignal, and M. Fialin (2013), Chemical influence on α -recoil damage annealing in apatite: Implications for (U-Th)/He dating, *Chem. Geol.*, 351, 257–267, doi:10.1016/j.chemgeo.2013.05.027.
- George, R., N. Rogers, and S. Kelley (1998), Earliest magnetism in Ethiopia: Evidence for two mantle plumes in one flood basalt province, *Geology*, 26(10), 923–926.
- George, R. M., and N. W. Rogers (2002), Plume dynamics beneath the African plate inferred from the geochemistry of the Tertiary basalts of southern Ethiopia, *Contrib. to Mineral. Petrol.*, 144, 286–304, doi:10.1007/s00410-002-0396-z.
- Gleadow, A.J.W., Gleadow, S.J., Belton, D.X., Kohn, B.P., Krochmal, M.S., Brown, R.W., 2009. Coincidence mapping – a key strategy for the automatic counting of fission tracks in natural minerals. *Geol. Soc. London Spec. Publ. Thermochronological Methods From Palaeotemperature Constraints to Landsc. Evol. Model.* 324, 25–36. <https://doi.org/10.1144/SP324.2>
- Gleadow, A., M. Harrison, B. Kohn, R. Lugo-zazueta, and D. Phillips (2015), The Fish Canyon Tuff: A new look at an old low-temperature thermochronology standard, *Earth Planet. Sci. Lett.*, 424, 95–108, doi:10.1016/j.epsl.2015.05.003.
- Gleadow, A. J. W., and I. R. Duddy (1981), A Natural Long-Term Track Annealing Experiment for Apatite, *Nucl. Tracks*, 5(1–2), 169–174.
- GMRD-Sudan (1981), Geological Map of the Sudan,
- Green, P. F., I. R. Duddy, A. J. W. Gleadow, and J. F. Lovering (1989), Apatite Fission-Track Analysis as a Paleotemperature Indicator for Hydrocarbon Exploration, in *Thermal History of Sedimentary Basins, Methods and Case Histories*, edited by N. D. Naeser and T. H. McCulloh, pp. 181–196, Springer-Verlag, New York.
- Guenther, W. R., P. W. Reiners, R. A. Ketcham, L. Nasdala, and G. Giester (2013), Helium diffusion in natural zircon: radiation damage, anisotropy, and the interpretation of zircon (U-Th)/He thermochronology, *Am. J. Sci.*, 313(3), 145–198, doi:10.2475/03.2013.01.
- Guenther, W. R., P. W. Reiners, and Y. Tian (2014), Interpreting date-eU correlations in zircon (U-Th)/He datasets: A case study from the Longmen Shan, China, *Earth Planet. Sci. Lett.*, 403(OCTOBER), 328–339, doi:10.1016/j.epsl.2014.06.050.
- Guenther, W. R., P. W. Reiners, H. Drake, and M. Tillberg (2017), Zircon, titanite, and apatite (U-Th)/He ages and age-eU correlations from the Fennoscandian Shield, southern Sweden, *Tectonics*, 36, doi:10.1002/2017TC004525.
- Guiraud, R., and W. Bosworth (1997), Senonian basin inversion and rejuvenation of rifting in Africa and Arabia: synthesis and implications to plate-scale tectonics, *Tectonophysics*, 282, 39–82.
- Gunnell, Y., K. Gallagher, A. Carter, M. Widdowson, and A. J. Hurford (2003), Denudation

- history of the continental margin of western peninsular India since the early Mesozoic - reconciling apatite fission-track data with geomorphology, *Earth Planet. Sci. Lett.*, *215*, 187–201, doi:10.1016/S0012-821X(03)00380-7.
- Haileab, B., F. H. Brown, I. McDougall, and P. N. Gathogo (2004), Gombe Group basalts and initiation of Pliocene deposition in the Turkana depression, northern Kenya and southern Ethiopia, *Geol. Mag.*, *141*, 41–53, doi:10.1017/S001675680300815X.
- Hansen, S. E., A. A. Nyblade, and M. H. Benoit (2012), Mantle structure beneath Africa and Arabia from adaptively parameterized P-wave tomography: Implications for the origin of Cenozoic Afro-Arabian tectonism, *Earth Planet. Sci. Lett.*, *319–320*, 23–34, doi:10.1016/j.epsl.2011.12.023.
- Hasebe, N., Barbarand, J., Jarvis, K., Carter, A., Hurford, A.J., 2004. Apatite fission-track chronometry using laser ablation ICP-MS. *Chem. Geol.* *207*, 134–145.
- Hendrie, D. B., N. J. Kusznir, C. K. Morley, and C. J. Ebinger (1994), Cenozoic extension in northern Kenya: a quantitative model of rift basin development in the Turkana region, *Tectonophysics*, *236*(1–4), 409–438, doi:10.1016/0040-1951(94)90187-2.
- Hourigan, J.K., Reiners, P.W., Brandon, M.T., 2005. U-Th zonation-dependent alpha-ejection in (U-Th)/He chronometry. *Geochim. Cosmochim. Acta* *69*, 3349–3365. <https://doi.org/10.1016/j.gca.2005.01.024>
- Ibrahim, A. E., C. J. Ebinger, and J. D. Fairhead (1991), Interpretation of the Central African Rift system in Sudan based on new gravity and aeromagnetic data, *Eos (Washington. DC)*, *462*.
- Kasanzu, C. H. (2016), Apatite fission track and (U-Th)/He thermochronology from the Archean Tanzania craton: Contributions to cooling histories of Tanzanian basement rocks, *Geosci. Front.*, 1–9, doi:10.1016/j.gsf.2016.09.007.
- Kasanzu, C. H., B. Linol, M. J. de Wit, R. Brown, C. Persano, and F. M. Stuart (2016), From source to sink in central Gondwana: Exhumation of the Precambrian basement rocks of Tanzania and sediment accumulation in the adjacent Congo basin, *Tectonics*, *35*(9), 2034–2051, doi:10.1002/2016TC004147.
- Kazmin, V. (1972), Geological Map of Ethiopia,
- Keranen, K., and S. L. Klemperer (2008), Discontinuous and diachronous evolution of the Main Ethiopian Rift: Implications for development of continental rifts, *Earth Planet. Sci. Lett.*, *265*, 96–111, doi:10.1016/j.epsl.2007.09.038.
- Ketcham, R. A. (2005), Forward and Inverse Modeling of Low- Temperature Thermochronometry Data, *Rev. Mineral. Geochemistry*, *58*, 275–314, doi:10.2138/rmg.2005.58.11.
- Ketcham, R. A., A. Carter, R. A. Donelick, J. Barbarand, and A. J. Hurford (2007a), Improved measurement of fission-track annealing in apatite using c-axis projection, *Am. Mineral.*, *92*(5–6), 789–798, doi:10.2138/am.2007.2280.
- Ketcham, R. A., A. Carter, R. A. Donelick, J. Barbarand, and A. J. Hurford (2007b), Improved modeling of fission-track annealing in apatite, *Am. Mineral.*, *92*, 799–810, doi:10.2138/am.2007.2281.
- Ketcham, R. A., W. R. Guenther, and P. W. Reiners (2013), Geometric analysis of radiation damage connectivity in zircon, and its implications for helium diffusion, *Am. Mineral.*, *98*(2–3), 350–360, doi:10.2138/am.2013.4249.
- Lachenbruch, A., and J. Sass (1977), Heat flow in the United States and the thermal regime of the crust, edited by J. G. Heacock, *Am. Geophys. Union, Geophys. Monogr. Ser.*, *20*, 626–675.
- Laslett, G. M., P. F. Green, I. R. Duddy, and A. J. W. Gleadow (1987), Thermal annealing of fission tracks in apatite 2 . A quantitative analysis, *Chem. Geol. (Isotope Geosci. Sect.)*, *65*, 1–13, doi:10.1016/0168-9622(87)90057-1.

- Lehto, T. et al. (2014), Geological Map of Uganda, Scale 1:1 000 000,
- Mackintosh, V., B. Kohn, A. Gleadow, and Y. Tian (2017), Phanerozoic Morphotectonic Evolution of the Zimbabwe Craton: Unexpected Outcomes From a Multiple Low-Temperature Thermochronology Study, *Tectonics*, 36, 1–24, doi:10.1002/2017TC004703.
- Mariita, N. O., and G. R. Keller (2007), An integrated geophysical study of the northern Kenya rift, *J. African Earth Sci.*, 48(2–3), 80–94, doi:10.1016/j.jafrearsci.2006.05.008.
- Mbongo Djimbi, D., C. Gautheron, J. Roques, L. Tassan-Got, C. Gerin, and E. Simoni (2015), Impact of apatite chemical composition on (U-Th)/He thermochronometry: An atomistic point of view, *Geochim. Cosmochim. Acta*, 167, doi:10.1016/j.gca.2015.06.017.
- McDougall, I., and F. H. Brown (2009), Timing of volcanism and evolution of the northern Kenya Rift, *Geol. Mag.*, 146(1), 34–37, doi:10.1017/S0016756808005347.
- McDowell, F.W., McIntosh, W.C., Farley, K.A., 2005. A precise ⁴⁰Ar-³⁹Ar reference age for the Durango apatite (U-Th)/He and fission track dating standard. *Chem. Geol.* 214, 249–263.
- Meesters, A. G. C. A., and T. J. Dunai (2002), Solving the production – diffusion equation for finite diffusion domains of various shapes: Part II. Application to cases with α -ejection and nonhomogenous distribution of the source, *Chem. Geol.*, 186, 347–363, doi:10.1016/S0009-2541(01)00423-5.
- Mitchell, S. G., and P. W. Reiners (2003), Influence of wildfires on apatite and zircon (U-Th)/He ages, *Geology*, 31(12), 1025–1028, doi:10.1130/G19758.1.
- Morley, C. K. (1994), Interaction of deep and shallow processes in the evolution of the Kenya rift, *Tectonophysics*, 236, 81–91.
- Morley, C. K. (1999a), Basin Evolution Trends in East Africa, in *Geoscience of Rift Systems-Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 131–150.
- Morley, C. K. (1999b), Boundary Fault Angle, with Particular Reference to the Lokichar Fault, Turkana Region, Kenya, in *Geoscience of Rift Systems - Evolution of East Africa: AAPG Studies in Geology No. 44*, vol. 44, edited by C. K. Morley, pp. 115–129.
- Morley, C. K. (2002), Tectonic settings of continental extensional provinces and their impact on sedimentation and hydrocarbon prospectivity, *SEPM Spec. Publ.*, 73, 22–56.
- Morley, C. K., W. a. Wescott, D. M. Stone, R. M. Harper, S. T. Wigger, and F. M. Karanja (1992), Tectonic evolution of the northern Kenyan Rift, *J. Geol. Soc. London.*, 149(3), 333–348, doi:10.1144/gsjgs.149.3.0333.
- Morley, C. K., W. Bosworth, R. A. Day, R. Lauck, R. Boshier, D. M. Stone, S. T. Wigger, W. A. Wescott, D. Haun, and N. Bassett (1999a), Geology and Geophysics of the Anza Graben, in *Geoscience of Rift Systems-Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 67–90.
- Morley, C. K., D. M. Stone, R. M. Harper, and S. T. Wigger (1999b), Geology and Geophysics of the Western Turkana Basins , Kenya, in *Geoscience of Rift Systems—Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 19–54.
- Mortimer, E., Kirstein, L.A., Stuart, F.M., Strecker, M.R., 2016. Spatio-temporal trends in normal-fault segmentation recorded by low-temperature thermochronology: Livingstone fault scarp, Malawi Rift, East African Rift System. *Earth Planet. Sci. Lett.* 455, 62–72. <https://doi.org/10.1016/j.epsl.2016.08.040>
- Moucha, R., and A. M. Forte (2011), Changes in African topography driven by mantle convection, *Nat. Geosci.*, 4(10), 707–712, doi:10.1038/ngeo1235.
- NASA (2000), Shuttle Radar Topography Mission (SRTM) 1 Arc-Second Global,

- Nasdala, L., P. W. Reiners, J. I. Garver, A. K. Kennedy, R. A. Stern, E. Balan, and R. Wirth (2004), Incomplete retention of radiation damage in zircon from Sri Lanka, *Am. Mineral.*, 89(1), 219–231, doi:10.2138/am-2004-0126.
- Noble, W. P., D. A. Foster, and A. J. W. Gleadow (1997), The post-Pan-African thermal and extensional history of crystalline basement rocks in eastern Tanzania, *Tectonophysics*, 275(4), 331–350, doi:10.1016/S0040-1951(97)00026-7.
- Nyblade, A. A. (1990), Terrestrial heat flow in east and southern Africa, *J. Geophys. Res. Solid Earth*, 95(B11), 17371–17384.
- Nyblade, A. A. (2011), The upper-mantle low-velocity anomaly beneath Ethiopia, Kenya, and Tanzania: Constraints on the origin of the African superswell in eastern Africa and plate versus plume models of mantle dynamics, *Geol. Soc. Am. Spec. Pap.*, 478, 1–14, doi:10.1130/2011.2478(03).
- Nyblade, A. A., and R. A. Brazier (2002), Precambrian lithospheric controls on the development of the east African Rift System, *Geology*, 7613, 755–758, doi:10.1130/0091-7613(2002)030<0755.
- Olen, S. M., T. A. Ehlers, and M. S. Densmore (2012), Limits to reconstructing paleotopography from thermochronometer data, *J. Geophys. Res.*, 117, 1–20, doi:10.1029/2011JF001985.
- Orme, D. A., W. R. Guenther, A. K. Laskowski, and P. W. Reiners (2016), Long-term tectonothermal history of Laramide basement from zircon???He age-eU correlations, *Earth Planet. Sci. Lett.*, 453, doi:10.1016/j.epsl.2016.07.046.
- Paton, C., Hellstrom, J., Paul, B., Woodhead, J., Hergt, J., 2011. Iolite: Freeware for the visualisation and processing of mass spectrometric data. *J. Anal. At. Spectrom.* 26, 2508–2518. <https://doi.org/10.1039/c1ja10172b>
- Patterson, B., A. K. Behrensmeyer, and W. D. Dill (1970), Geology and Fauna of a New Pliocene Locality in North-western Kenya, *Nature*, 226, 918–921.
- Pérez-Gussinyé, M., M. Metois, M. Fernández, J. Vergés, J. Fulla, and A. R. Lowry (2009), Effective elastic thickness of Africa and its relationship to other proxies for lithospheric structure and surface tectonics, *Earth Planet. Sci. Lett.*, 287, 152–167, doi:10.1016/j.epsl.2009.08.004.
- Philippon, M., G. Corti, F. Sani, M. Bonini, M. Balestrieri, P. Molin, E. Willingshofer, D. Sokoutis, and S. Cloetingh (2014), Evolution, distribution, and characteristics of rifting in southern Ethiopia, *Tectonics*, 33, 1–24, doi:10.1002/2013TC003430.
- Philpotts, A., Ague, J., 2009. Principles of igneous and metamorphic petrology. Cambridge University Press.
- Pik, R. (2011), Geodynamics: East Africa on the rise, *Nat. Geosci.*, 4(10), 660–661, doi:10.1038/ngeo1274.
- Pik, R., B. Marty, J. Carignan, and J. Lavé (2003), Stability of the Upper Nile drainage network (Ethiopia) deduced from (U-Th)/He thermochronometry: implications for uplift and erosion of the Afar plume dome, *Earth P*, 215, 73–88, doi:10.1016/S0012-821X(03)00457-6.
- Pik, R., B. Marty, J. Carignan, G. Yirgu, and T. Ayalew (2008), Timing of East African Rift development in southern Ethiopia : Implication for mantle plume activity and evolution of topography, *Geology*, 36(2), 167–170, doi:10.1130/G24233A.1.
- Powell, J., D. Schneider, D. Stockli, and K. Fallas (2016), Zircon (U-Th)/He thermochronology of Neoproterozoic strata from the Mackenzie Mountains, Canada: Implications for the Phanerozoic exhumation and deformation history of the northern Canadian Cordillera, *Tectonics*, 35, 1–27, doi:10.1002/2015TC003989.Received.
- Reeves, C. V., J. P. Teasdale, and E. S. Mahanjane (2016), Insight into the Eastern Margin of Africa from a new tectonic model of the Indian Ocean, *Geol. Soc. London Spec. Publ.*,

- 431, doi:10.1144/SP431.12.
- Reeves, C. V., F. M. Karanja, and I. N. MacLeod (1987), Geophysical evidence for a failed Jurassic rift and triple junction in Kenya, *Earth Planet. Sci. Lett.*, 81(2–3), 299–311, doi:10.1016/0012-821X(87)90166-X.
- Reiners, P. W., 2005. Zircon (U-Th)/He Thermochronometry. *Rev. Mineral. Geochemistry* 58, 151–179. <https://doi.org/10.2138/rmg.2005.58.6>
- Reiners, P. W., and M. T. Brandon (2006), Using Thermochronology to Understand Orogenic Erosion, *Annu. Rev. Earth Planet. Sci.*, 34, 419–466, doi:10.1146/annurev.earth.34.031405.125202.
- Reiners, P. W., and K. A. Farley (2001), Influence of crystal size on apatite (U-Th)/ He thermochronology: an example from the Bighorn Mountains, Wyoming, *Earth Planet. Sci. Lett.*, 188, 413–420.
- Reiners, P. W., K. A. Farley, and H. J. Hickes (2002), He diffusion and (U-Th)/He thermochronometry of zircon: Initial results from Fish Canyon Tuff and Gold Butte, *Tectonophysics*, 349(1–4), 297–308, doi:10.1016/S0040-1951(02)00058-6.
- Reiners, P. W., T. A. Ehlers, and P. K. Zeitler (2005), Past, present, and future of thermochronology, *Revi*, 58, 1–18, doi:10.2138/rmg.2005.58.1.
- Reiners, P. W., S. N. Thomson, D. McPhillips, R. A. Donelick, and J. J. Roering (2007), Wildfire thermochronology and the fate and transport of apatite in hillslope and fluvial environments, *J. Geophys. Res. Earth Surf.*, 112(4), doi:10.1029/2007JF000759.
- Renaut, R. W., C. K. Morley, and B. Jones (2002), Fossil hot-spring travertine in the Turkana Basin, northern Kenya: Structure, facies, and genesis, *SEPM Spec. Publ.*, 73, 123–141.
- Rogers, N. W. (2006), Basaltic magmatism and the geodynamics of the East African Rift System, *Geol. Soc. London, Spec. Publ.*, 259, 77–93, doi:10.1144/GSL.SP.2006.259.01.08.
- Rooney, T., T. Furman, I. Bastow, D. Ayalew, and G. Yirgu (2007), Lithospheric modification during crustal extension in the Main Ethiopian Rift, *J. Geophys. Res.*, 112, 1–21, doi:10.1029/2006JB004916.
- Schull, T. J. (1988), Rift Basins of Interior Sudan: Petroleum Exploration and Discovery, , 72(10), 1128–1142.
- Seiler, C., A. J. W. Gleadow, J. M. Fletcher, and B. P. Kohn (2009), Thermal evolution of a sheared continental margin: Insights from the Ballenas transform in Baja California, Mexico, *Earth Planet. Sci. Lett.*, 285(1–2), 61–74, doi:10.1016/j.epsl.2009.05.043.
- Seiler, C., Kohn, B., Gleadow, A., 2014. Apatite fission track analysis by LA-ICP-MS: An evaluation of the absolute dating approach, in: International Conference on Thermochronology.
- Sembroni, A., C. Faccenna, T. W. Becker, P. Molin, and B. Abebe (2016), Long-term, deep-mantle support of the Ethiopia-Yemen Plateau, *Tectonics*, 35, 469–488, doi:10.1002/2015TC004000.Received.
- Shuster, D. L., R. M. Flowers, and K. A. Farley (2006), The influence of natural radiation damage on helium diffusion kinetics in apatite, *Earth Planet. Sci. Lett.*, 249(3–4), 148–161, doi:10.1016/j.epsl.2006.07.028.
- Sippel, J., C. Meeßen, M. Cacace, J. Mechie, S. Fishwick, C. Heine, M. Scheck-Wenderoth, and M. R. Strecker (2017), The Kenya rift revisited: insights into lithospheric strength through data-driven 3-D gravity and thermal modelling, *Solid Earth*, 8, 45–81, doi:10.5194/se-8-45-2017.
- Smith, M., and P. Mosley (1993), Crustal Heterogeneity and Basement Influence on the Development of the Kenya Rift, East Africa, *Tectonics*, 12(2), 591–606.
- Spear, F.S., 1993. Metamorphic phase equilibria and pressure-temperature-time paths. Mineralogical Society of America, Washington, DC.

- Spiegel, C., B. P. Kohn, D. X. Belton, and A. J. W. Gleadow (2007), Morphotectonic evolution of the central Kenya rift flanks: Implications for late Cenozoic environmental change in East Africa, *Geology*, 35(5), 427–430, doi:10.1130/G23108A.1.
- Stockli, D. F. (2005), Application of Low-Temperature Thermochronometry to Extensional Tectonic Settings, *Rev. Mineral. Geochemistry*, 58, 411–448, doi:10.2138/rmg.2005.58.16.
- Tagami, T., Shimada, C., 1996. Natural long-term annealing of the zircon fission track system around a granitic pluton. *J. Geophys. Res. Solid Earth* 101, 8245–8255.
- Talbot, M. R., C. K. Morley, J.-J. Tiercelin, A. Le Hérissé, J.-L. Potdevin, and B. Le Gall (2004), Hydrocarbon potential of the Meso-Cenozoic Turkana Depression, northern Kenya. II. Source rocks: quality, maturation, depositional environments and structural control, *Mar. Pet. Geol.*, 21, 63–78, doi:10.1016/j.marpetgeo.2003.11.008.
- Tiercelin, J.-J., J.-L. Potdevin, C. K. Morley, M. R. Talbot, H. Bellon, A. Rio, B. Le Gall, and W. Vétel (2004), Hydrocarbon potential of the Meso-Cenozoic Turkana Depression, northern Kenya. I. Reservoirs: depositional environments, diagenetic characteristics, and source rock–reservoir relationships, *Mar. Pet. Geol.*, 21, 41–62, doi:10.1016/j.marpetgeo.2003.11.007.
- Tiercelin, J.-J., T. Nalpas, P. Thuo, and J.-L. Potdevin (2012a), Hydrocarbon Prospectivity in Mesozoic and Early–Middle Cenozoic Rift Basins of Central and Northern Kenya, Eastern Africa, *AAPG Mem.*, 100, 179–208, doi:10.1306/13351553M1001742.
- Tiercelin, J. J. et al. (2012b), Stratigraphy, sedimentology and diagenetic evolution of the Lapur Sandstone in northern Kenya: Implications for oil exploration of the Meso-Cenozoic Turkana depression, *J. African Earth Sci.*, 71–72, 43–79, doi:10.1016/j.jafrearsci.2012.06.007.
- Torres Acosta, V., A. Bande, E. R. Sobel, M. Parra, T. F. Schildgen, F. Stuart, and M. R. Strecker (2015), Cenozoic extension in the Kenya Rift from low-temperature thermochronology: Links to diachronous spatiotemporal evolution of rifting in East Africa, *Tectonics*, 34(12), 2367–2386, doi:10.1002/2015TC003949.
- Vermeesch, P. (2009), RadialPlotter: A Java application for fission track, luminescence and other radial plots, *Radiat. Meas.*, 44(4), 409–410.
- Vétel, W., and B. Le Gall (2006), Dynamics of prolonged continental extension in magmatic rifts: the Turkana Rift case study (North Kenya), *Geol. Soc. London Spec. Publ.*, 259, 209–233.
- Vétel, W., B. Le Gall, and T. C. Johnson (2004), Recent tectonics in the Turkana Rift (North Kenya): an integrated approach from drainage network, satellite imagery and reflection seismic analyses, *Basin Res.*, 16, 165–181, doi:10.1111/j.1365-2117.2004.00227.x.
- Vétel, W., B. Le Gall, and J. J. Walsh (2005), Geometry and growth of an inner rift fault pattern: the Kino Sogo Fault Belt, Turkana Rift (North Kenya), *J. Struct. Geol.*, 27, 2204–2222, doi:10.1016/j.jsg.2005.07.003.
- Wagner, G., and P. Van den Haute (1992), *Fission-Track Dating*, Kluwer Academic Publishers, Dordrecht.
- Wagner, M., R. Altherr, and P. Van Den Haute (1992), Apatite fission-track analysis of Kenyan basement rocks: constraints on the thermotectonic evolution of the Kenya dome. A reconnaissance study, *Tectonophysics*, 204(1–2), 93–110, doi:10.1016/0040-1951(92)90272-8.
- Walsh, J., and R. G. Dodson (1969), Geology of Northern Turkana, Degree Sheets 1, 2, 9, and 10,
- Wescott, W. A., C. K. Morley, and F. M. Karanja (1993), Geology of the “Turkana Grits” in the Lariu range and Mt. Porr areas, southern Lake Turkana, Northwestern Kenya, *J. African Earth Sci.*, 16(4), 425–435, doi:10.1016/0899-5362(93)90101-U.

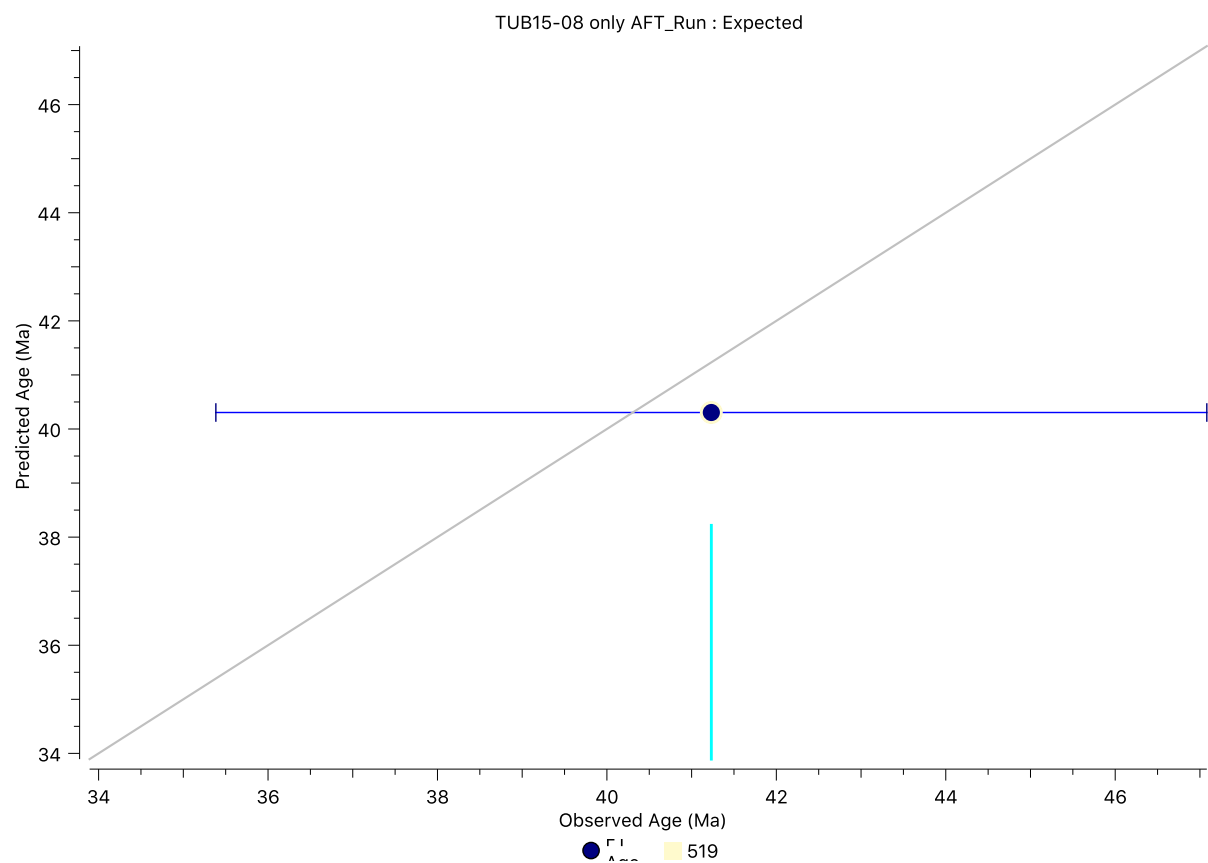
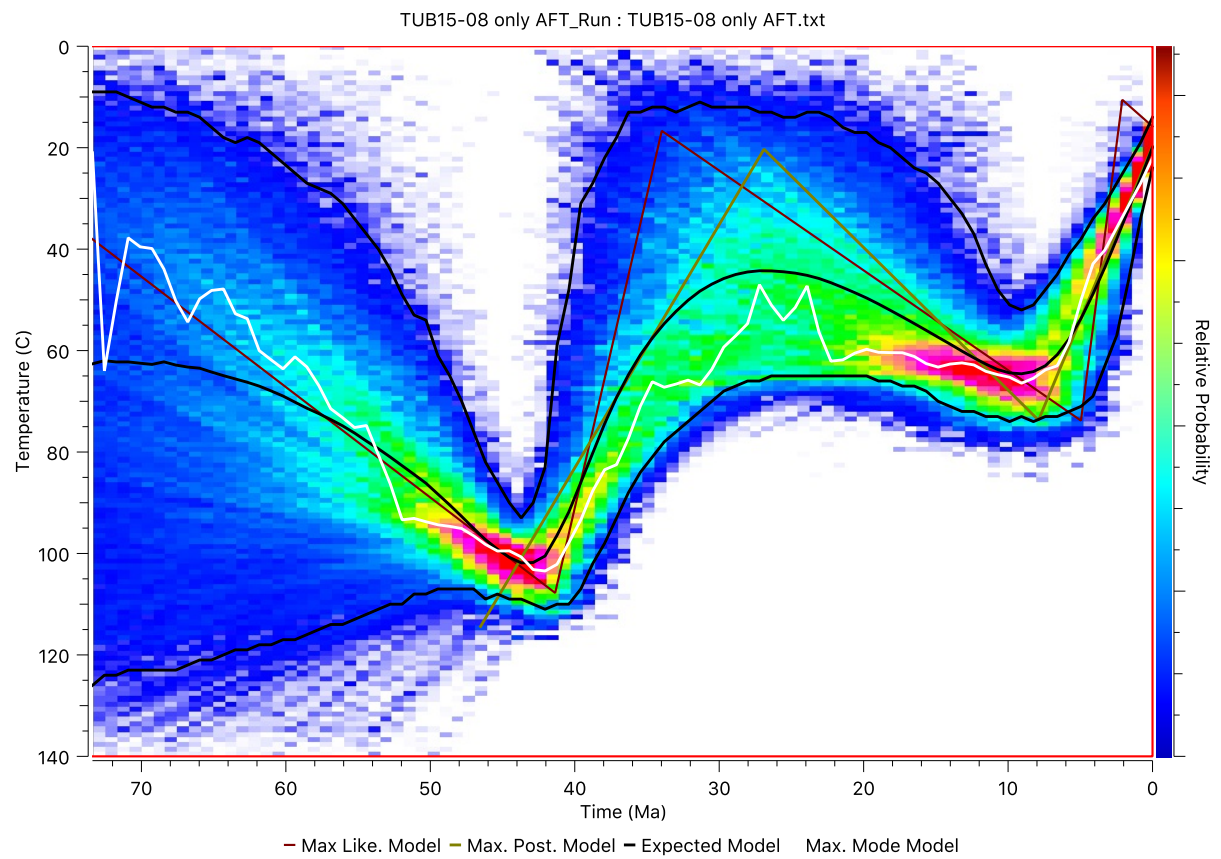
- Wescott, W. A., S. T. Wigger, D. M. Stone, and C. K. Morley (1999), Geology and Geophysics of the Lotikipi Plain, *Geosci. Rift Syst. East Africa*, (44), 55–66.
- Wheildon, J., P. Morgan, K. H. Williamson, T. R. Evans, and C. A. Swanberg (1994), Heat flow in the Kenya rift zone, *Tectonophysics*, 236(1–4), 131–149, doi:10.1016/0040-1951(94)90173-2.
- Wichura, H., R. Bousquet, R. Oberhänsli, M. R. Strecker, and M. H. Trauth (2010), Evidence for middle Miocene uplift of the East African Plateau, *Geology*, 38(6), 543–546, doi:10.1130/G31022.1.
- Wichura, H., R. Bousquet, R. Oberhänsli, M. R. Strecker, and M. H. Trauth (2011), The Mid-Miocene East African Plateau: a pre-rift topographic model inferred from the emplacement of the phonolitic Yatta lava flow, Kenya, *Geol. Soc. London, Spec. Publ.*, 357(1), 285–300, doi:10.1144/sp357.15.
- Wildman, M., R. Brown, R. Beucher, C. Persano, F. Stuart, K. Gallagher, J. Schwanethal, and A. Carter (2016), The chronology and tectonic style of landscape evolution along the elevated Atlantic continental margin of South Africa resolved by joint apatite fission track and (U-Th-Sm)/He thermochronology, *Tectonics*, 35(3), 511–545, doi:10.1002/2015TC004042.
- Williamson, P. G., and R. J. G. Savage (1986), Early rift sedimentation in the Turkana basin, northern Kenya, *Geol. Soc. London, Spec. Publ.*, 25(1), 267–283, doi:10.1144/GSL.SP.1986.025.01.22.
- Wolf, R. A., K. A. Farley, and D. M. Kass (1998), Modeling of the temperature sensitivity of the apatite (U – Th)/He thermochronometer, *Chem. Geol.*, 148, 105–114.
- Wolfe, M. R., and D. F. Stockli (2010), Zircon (U-Th)/He thermochronometry in the KTB drill hole, Germany, and its implications for bulk He diffusion kinetics in zircon, *Earth Planet. Sci. Lett.*, 295(1–2), 69–82, doi:10.1016/j.epsl.2010.03.025.
- Wolfenden, E., C. Ebinger, G. Yirgu, A. Deino, and D. Ayalew (2004), Evolution of the northern Main Ethiopian rift: birth of a triple junction, *Earth Planet. Sci. Lett.*, 224(1–2), 213–228, doi:10.1016/j.epsl.2004.04.022.
- Zanettin, B., E. Justin Visentin, G. Bellieni, E. M. Piccirillo, and R. Francesca (1983), The volcanism of the North Turkana Basin, Kenya: age, succession and structural evolution, *Bull. des Centres Rech. Explor. Elf-Aquitaine*, 7, 249–255.

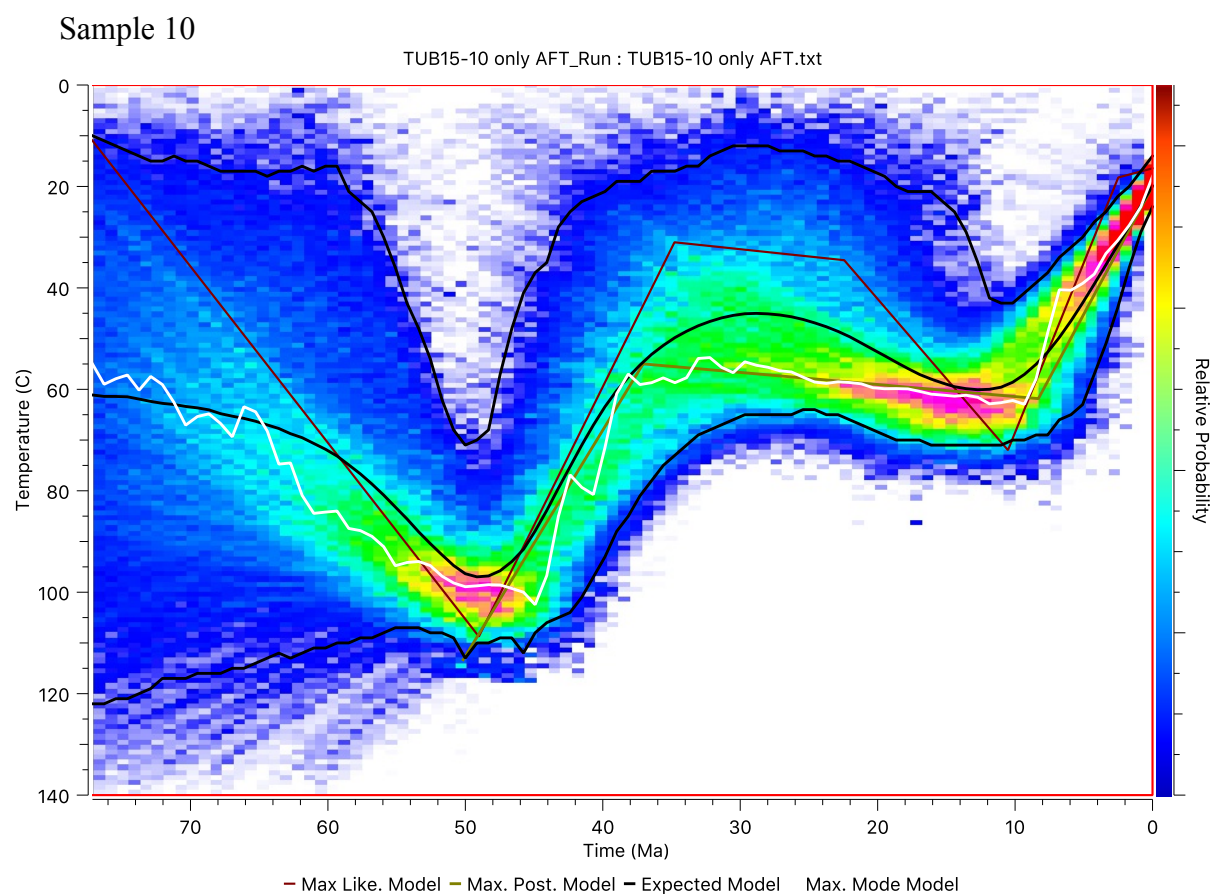
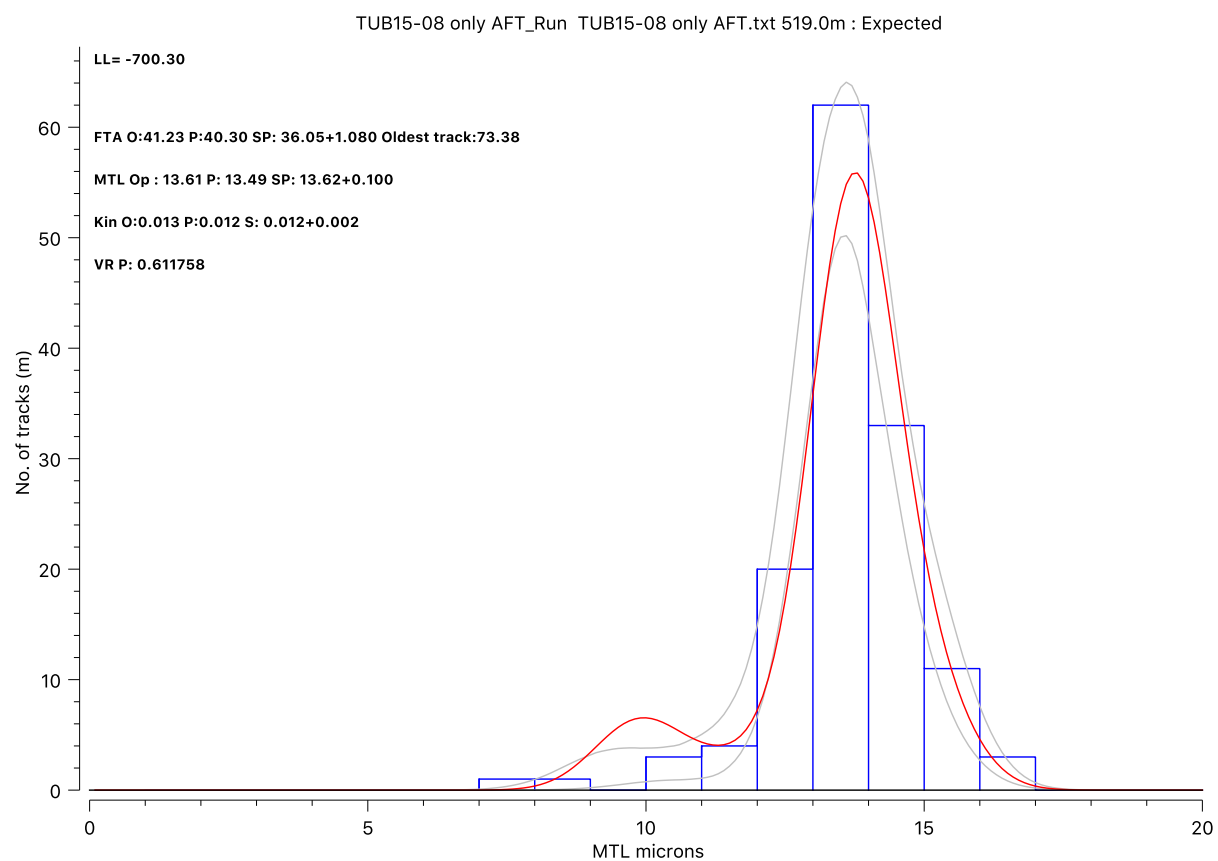
Supplementary Data A: Thermal history models using only AFT data

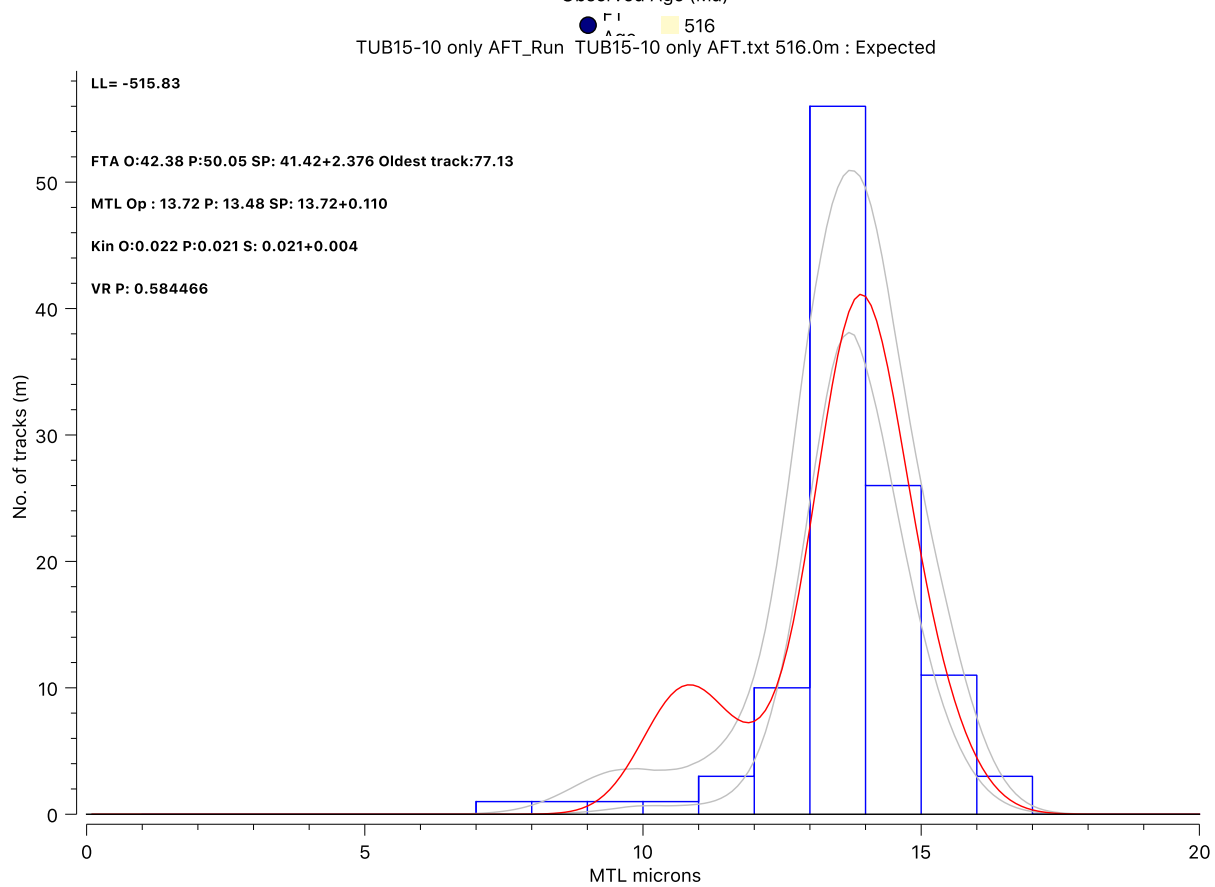
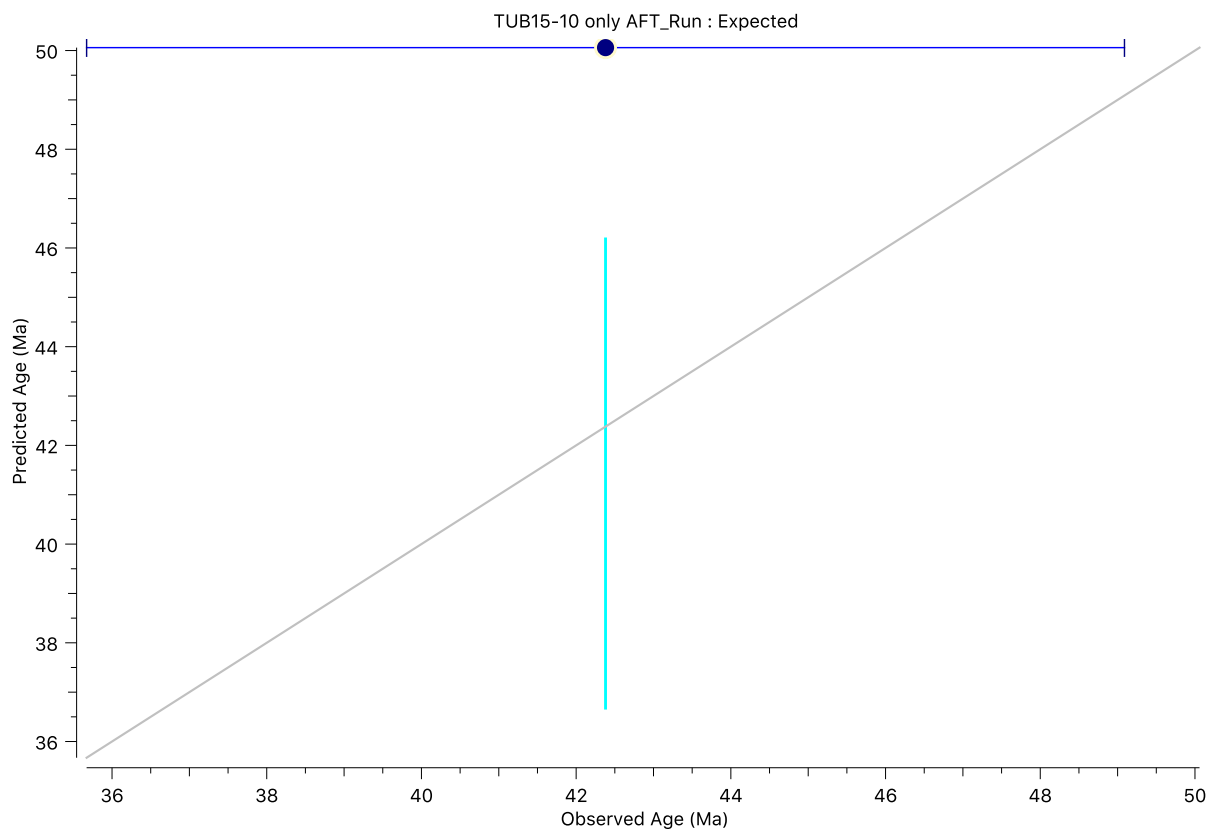
Thermal history models displayed below were generated in the software *QTQt* using only AFT data, following the procedure outlined in section 3.6. The Bayesian transdimensional Markov chain Monte Carlo (MCMC) employed by *QTQt* requires the user to define a time-temperature range space in which the MCMC samples [Gallagher *et al.*, 2009; Gallagher, 2012]. For thermal history modelling of AFT data only, a temperature range of 0-140 °C and a time range of 0 Ma to twice the age of the AFT age was used. Individual thermal history models were run for 150,000 iterations, with the first 50,000 discarded as burn-in [Gallagher *et al.*, 2009]. To ensure satisfactory sampling, the acceptance rates of the proposed model parameters and the posterior sampling chain were checked after each modelling run [Gallagher *et al.*, 2012]. The resulting models displayed below are the expected time-temperature histories (weighted mean of the posterior distribution), deemed to be most representative sample thermal histories [see Gallagher, 2012 and Wildman *et al.*, 2016 for discussion].

Only samples with a statistically significant number of confined track length measurements (> 100) were modelled. A sole time-temperature constraint was applied, requiring samples to be at near-surface temperatures (20 ± 10 °C) by the present-day. Plots of observed versus predicted AFT ages and confined track distributions are displayed below each thermal history model, allowing for the data fit of the models to be assessed. Thermal history models generated for samples 23 and 30 using only AFT are displayed in Figure 3.7.

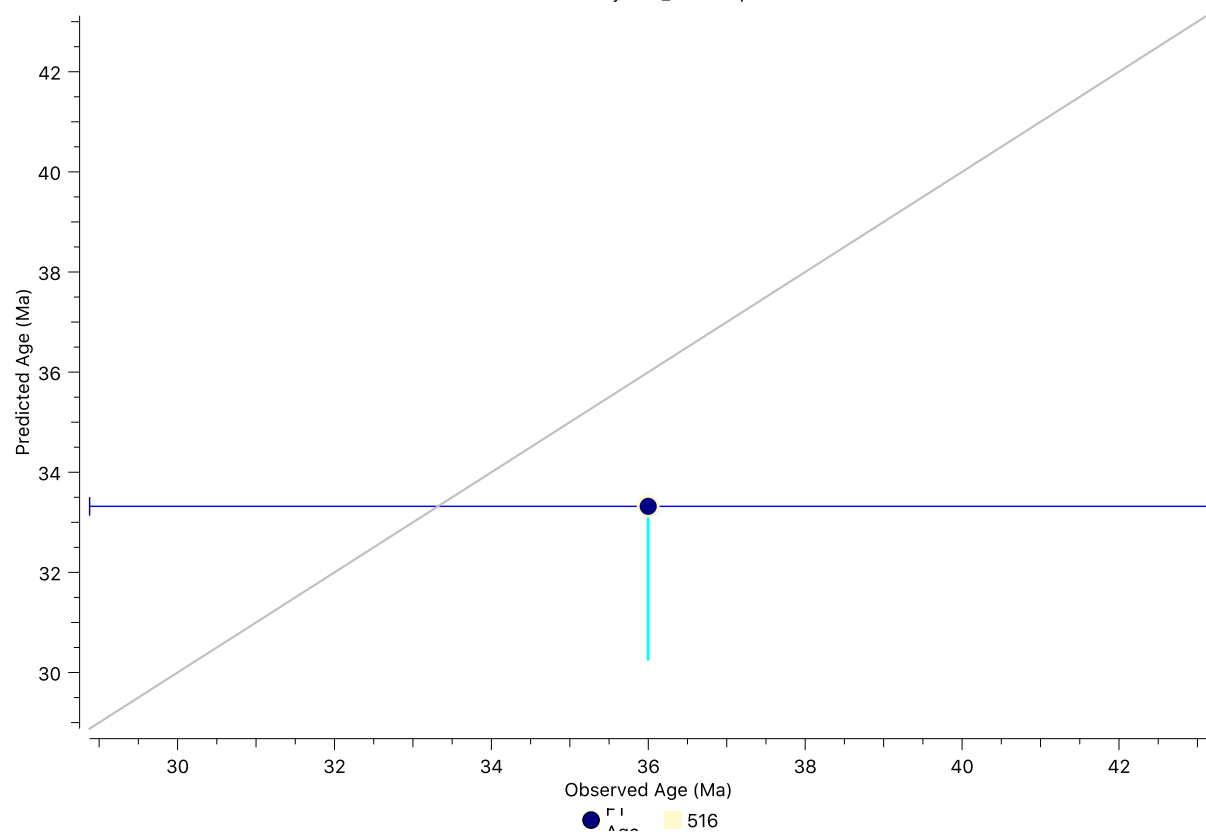
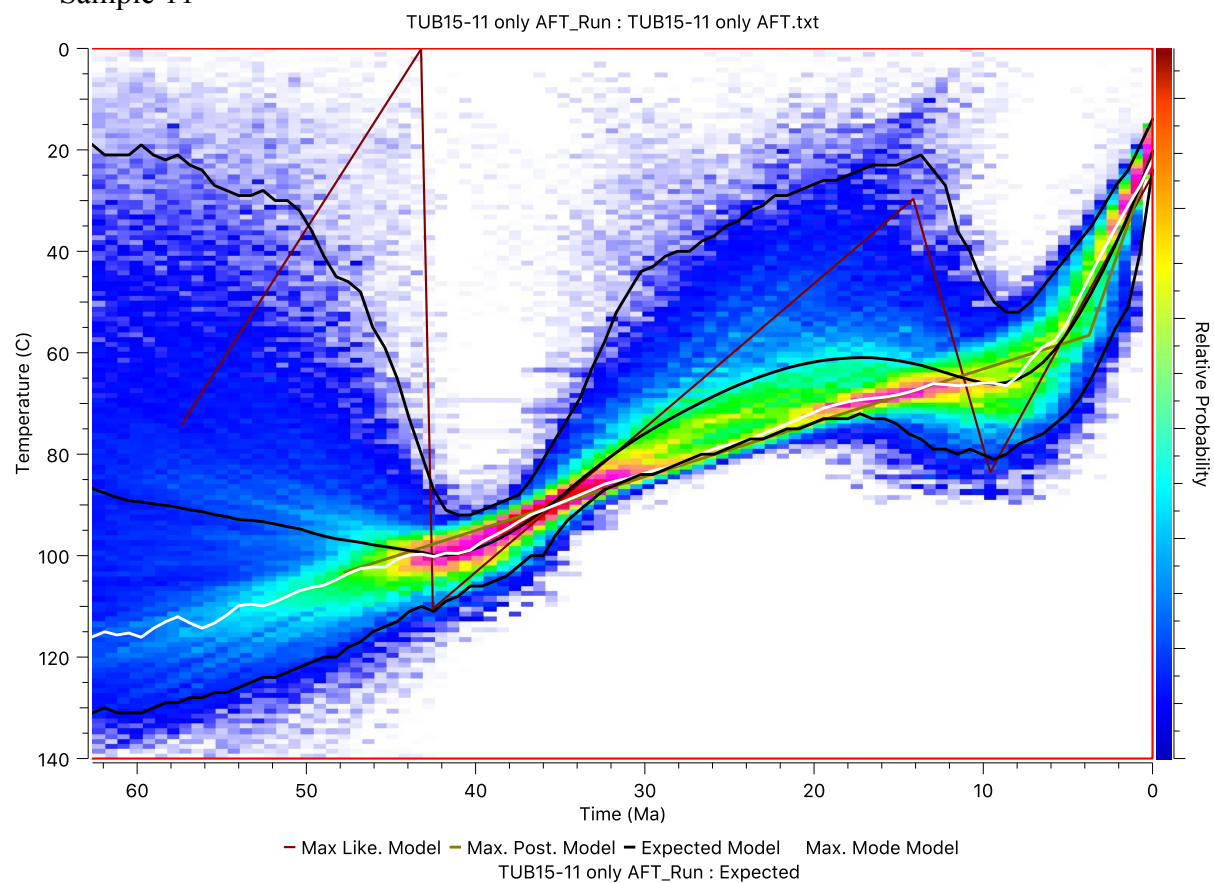
Sample 8

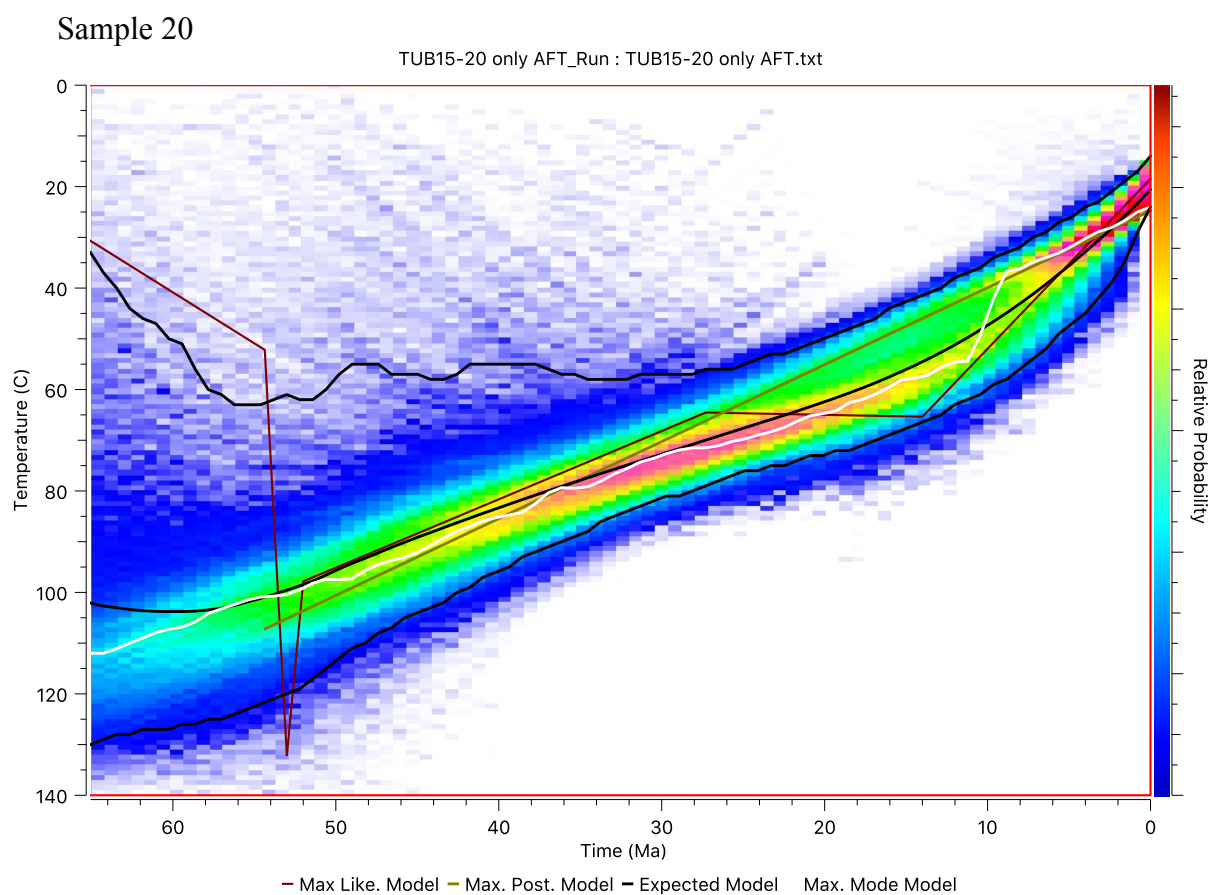
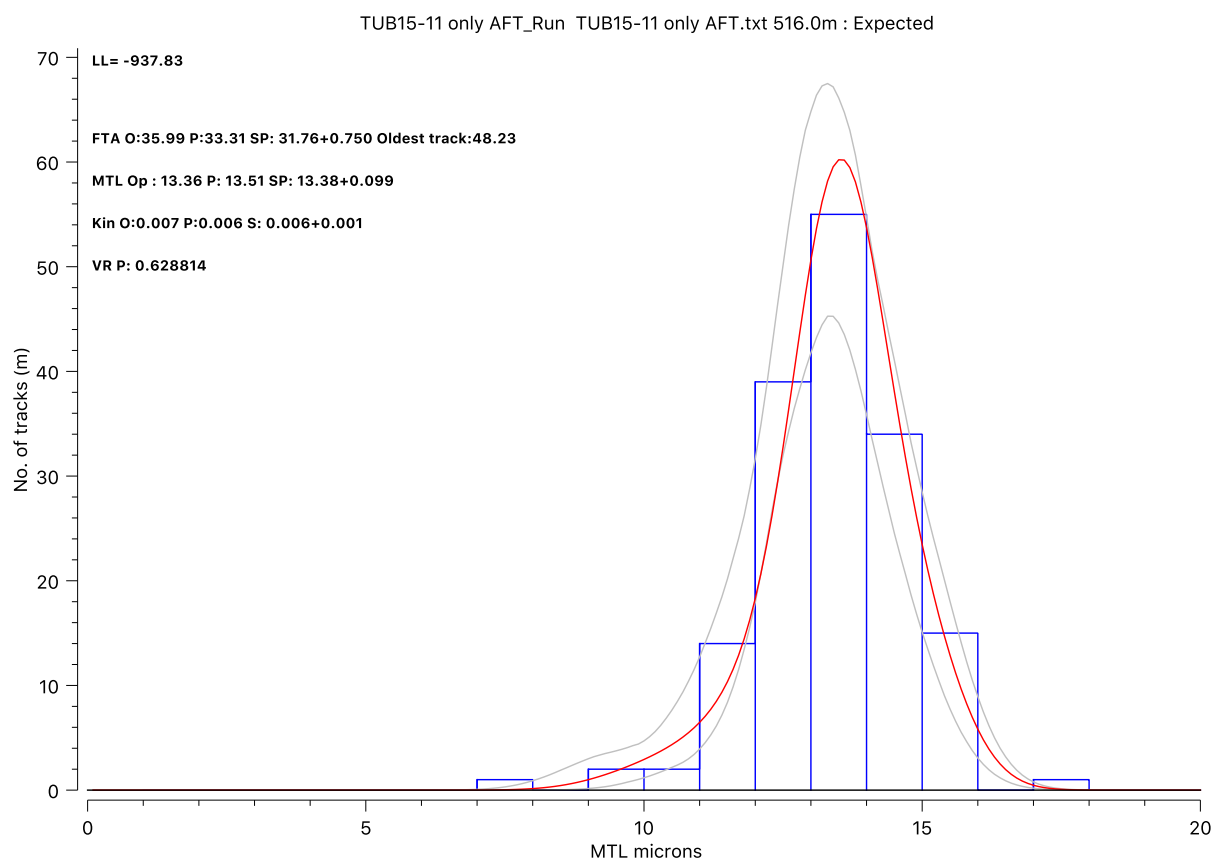


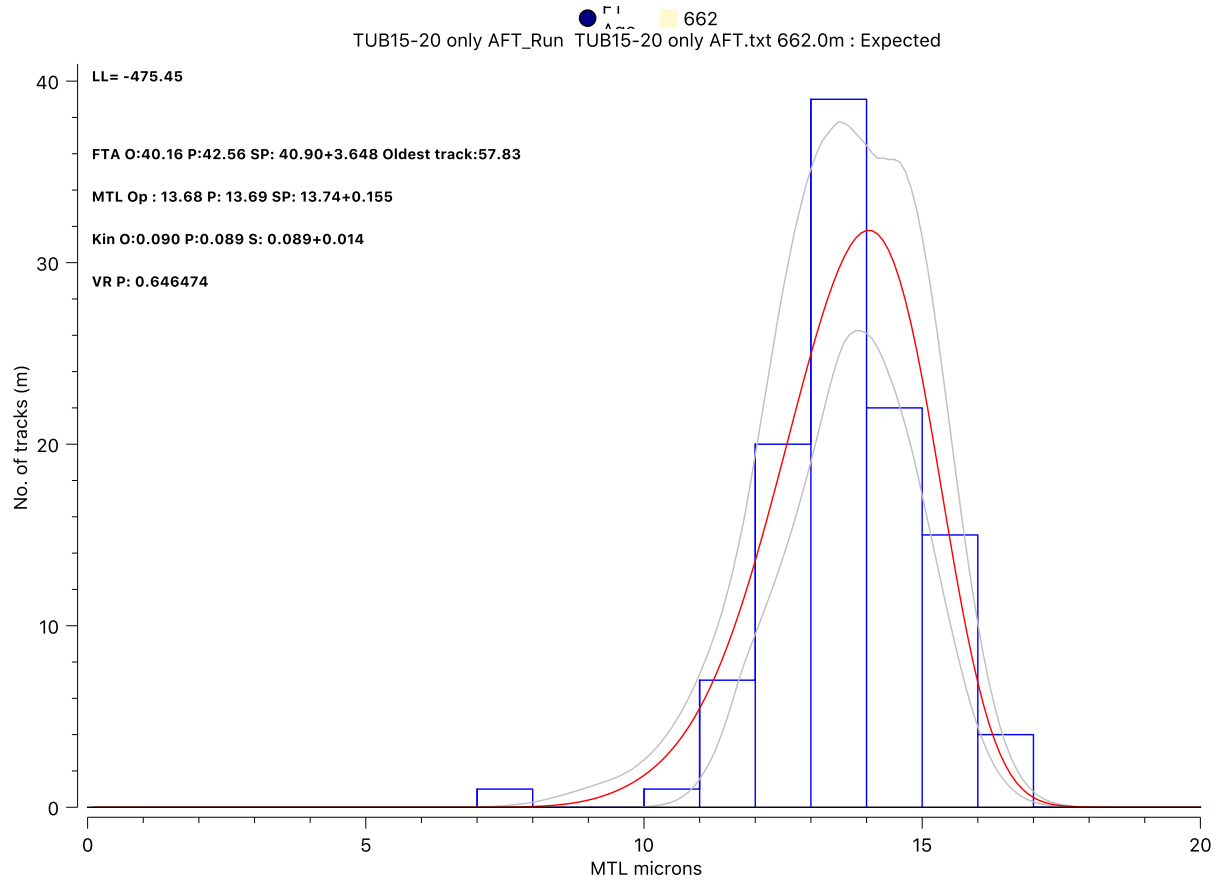
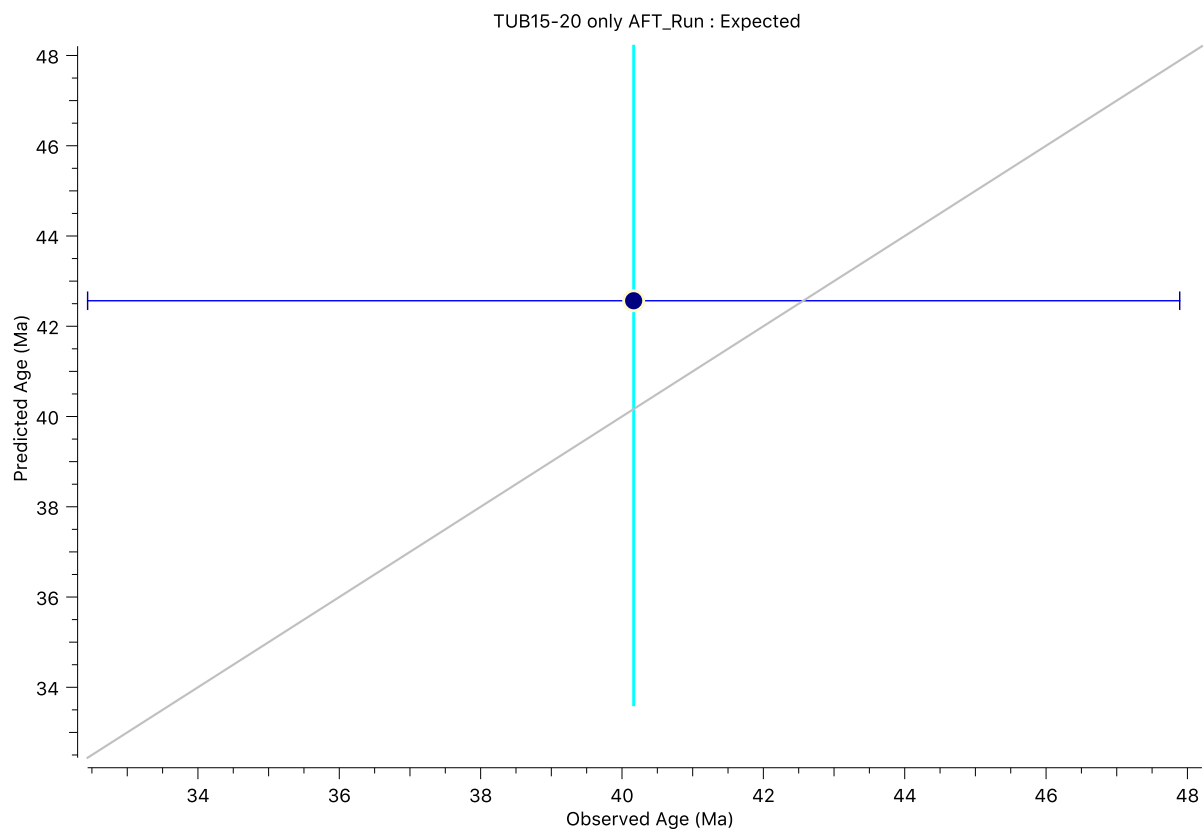




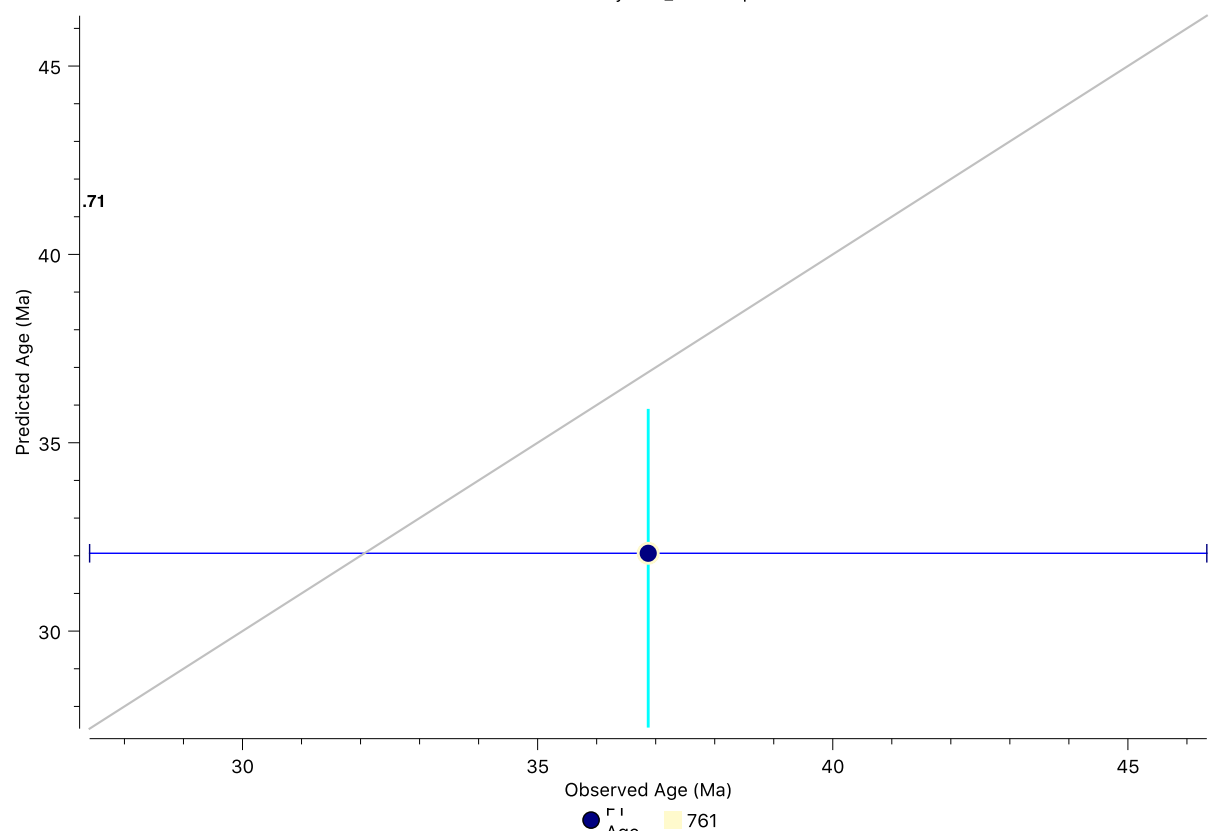
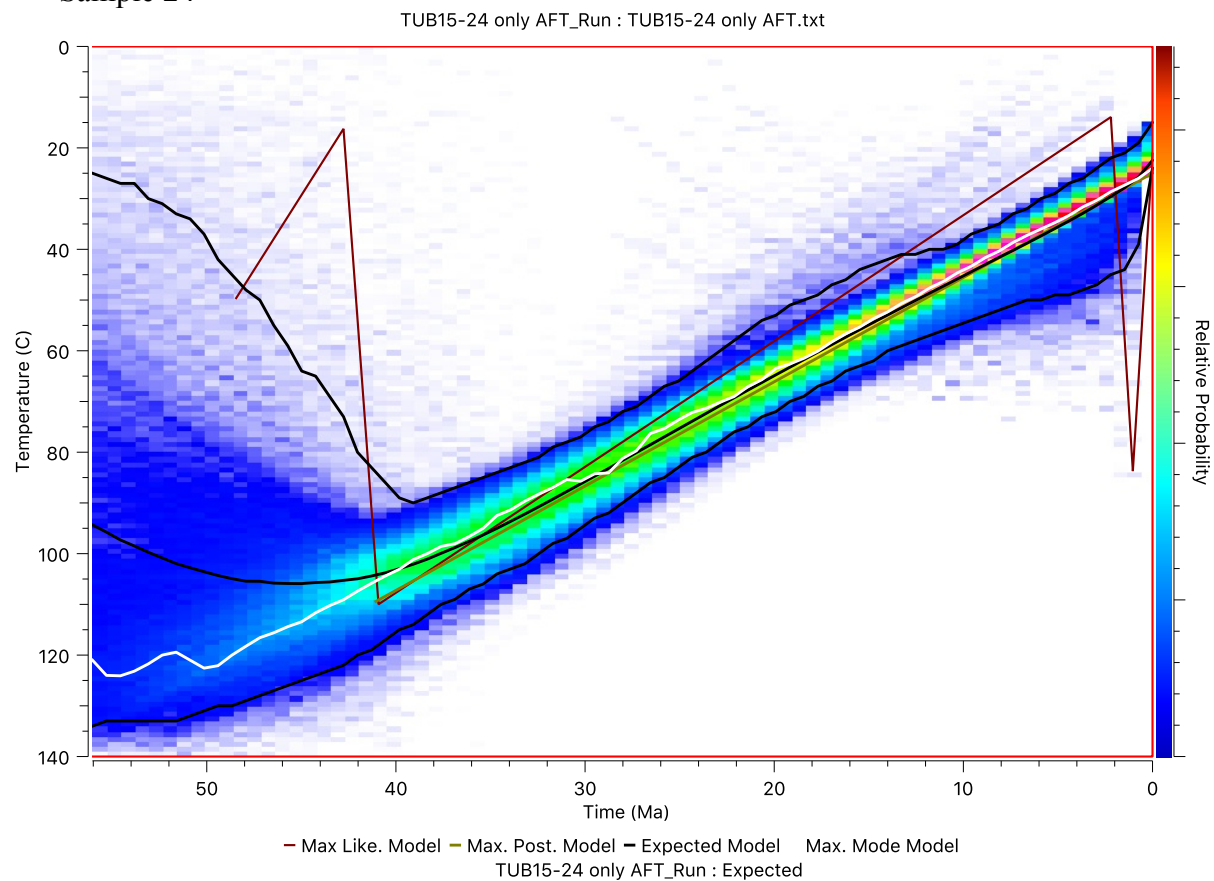
Sample 11

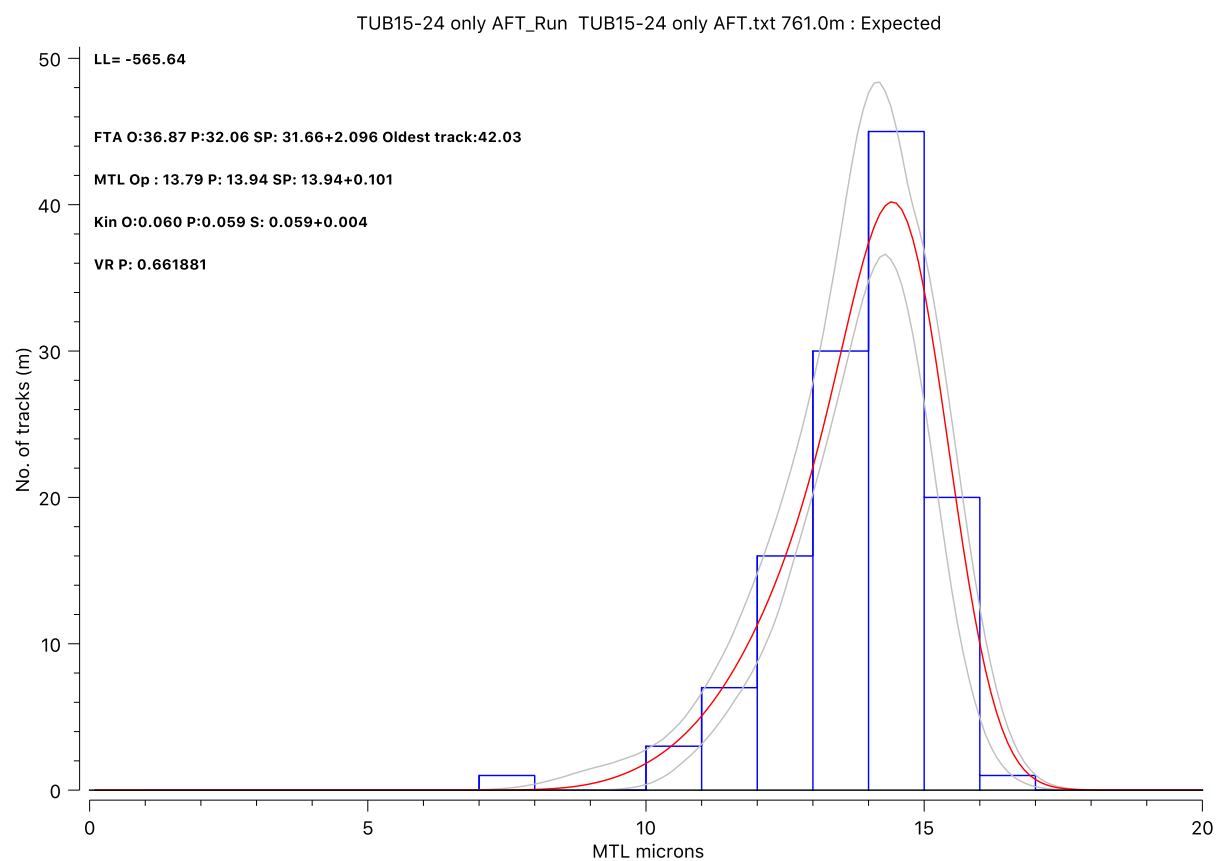






Sample 24





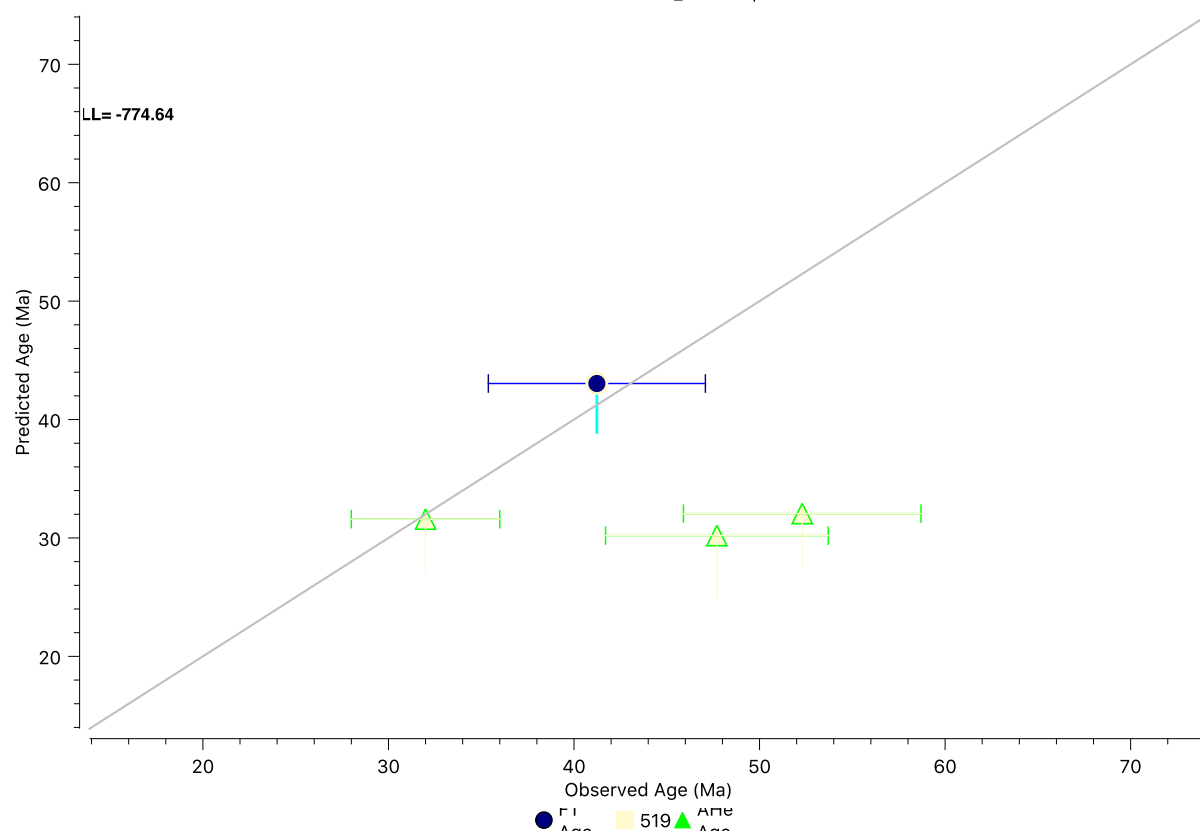
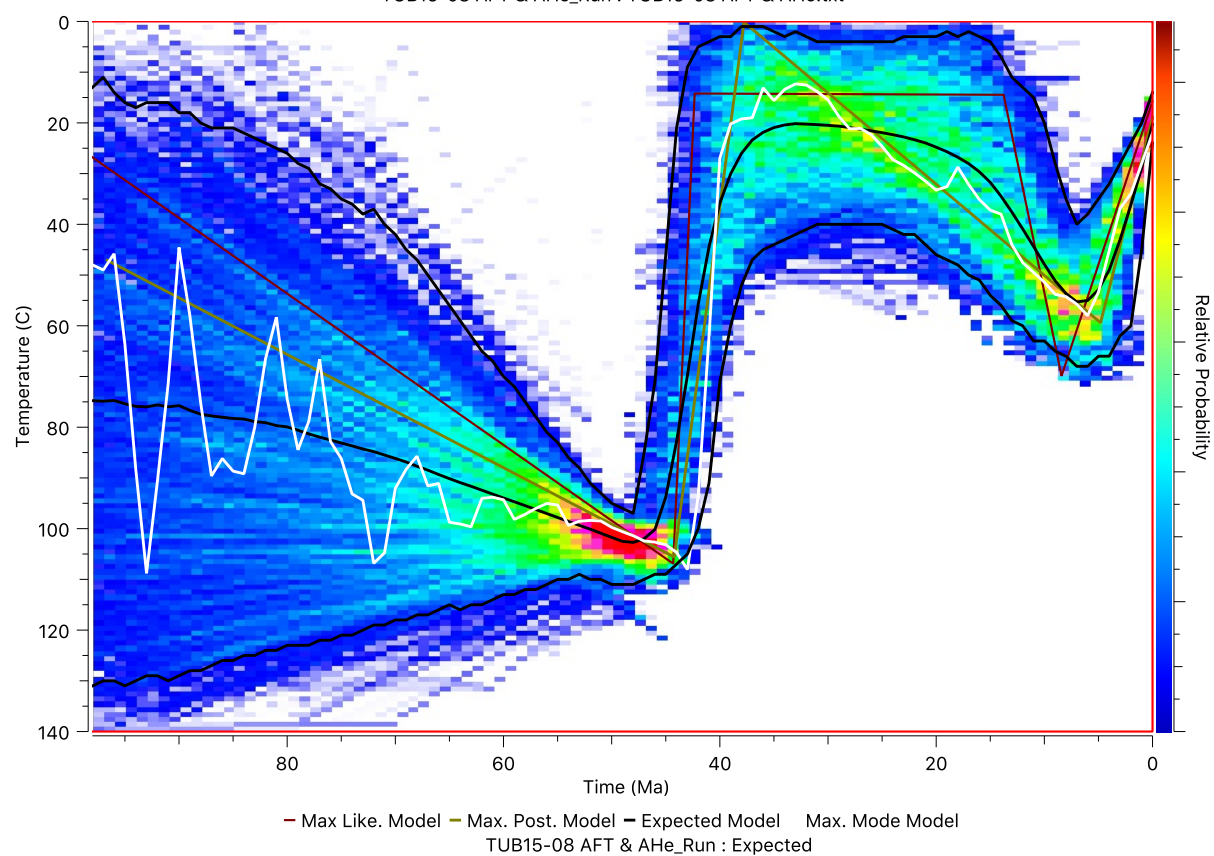
Supplementary Data B: Thermal history models using AFT and AHe data

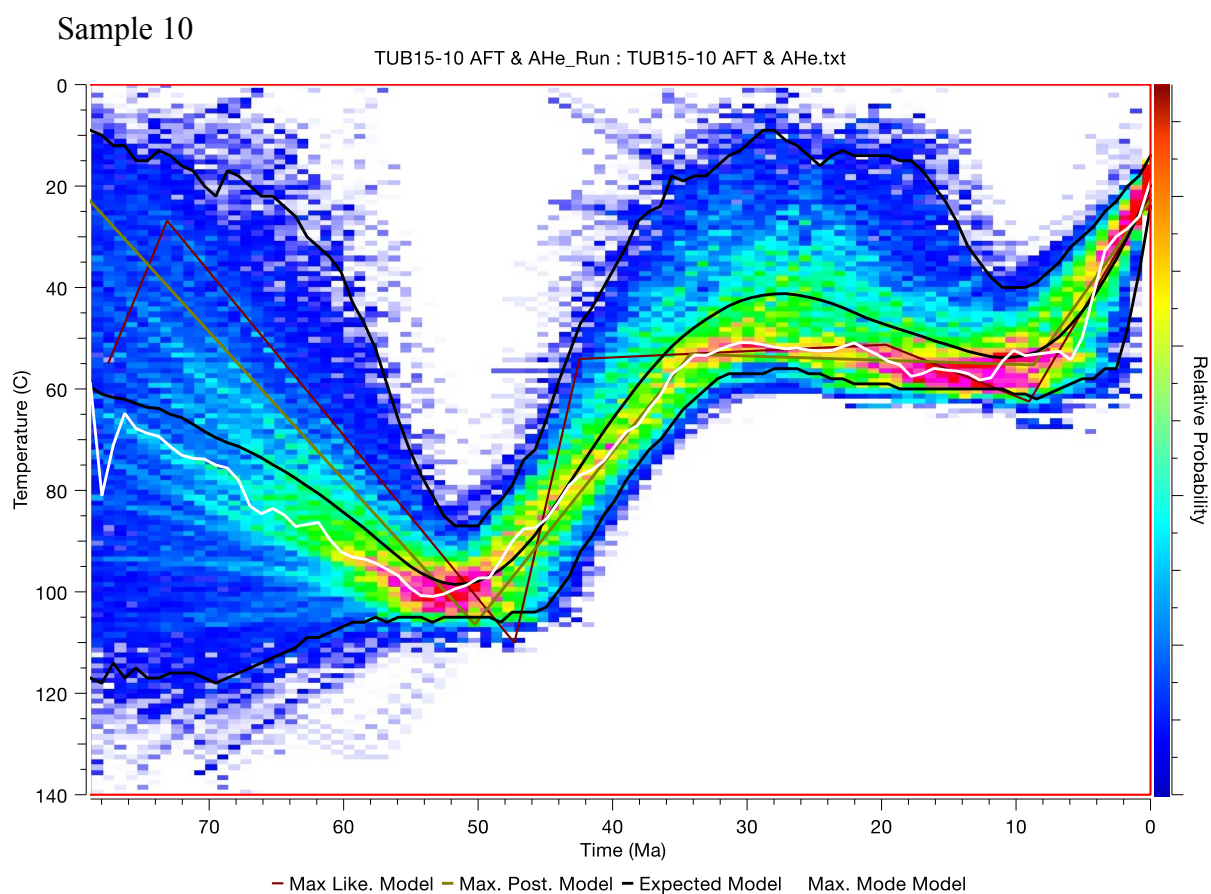
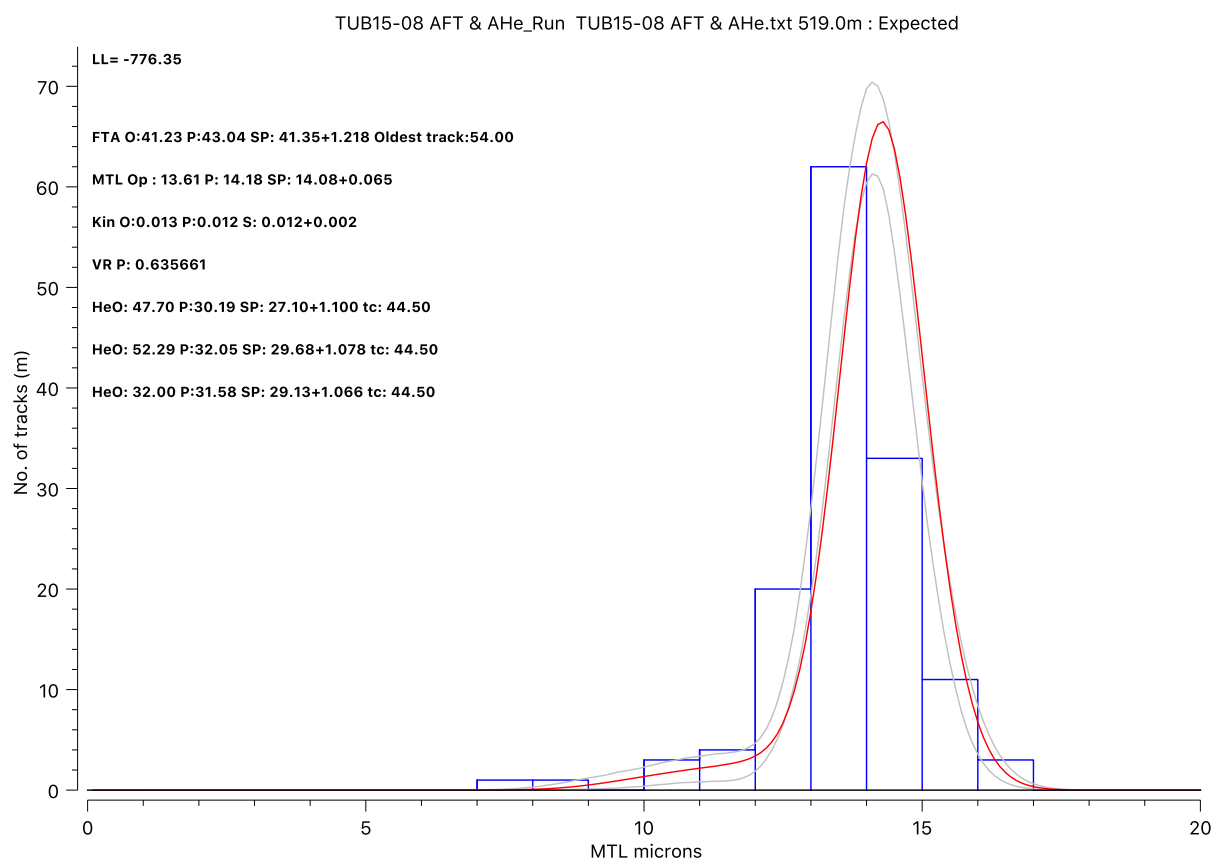
Thermal history models displayed below were generated in the software *QTQt* using AFT and AHe data following the procedure outlined in section 3.6. The Bayesian transdimensional Markov chain Monte Carlo (MCMC) employed by *QTQt* requires the user to define a time-temperature range space in which the MCMC samples [Gallagher *et al.*, 2009; Gallagher, 2012]. For thermal history modelling of AFT and AHe data, a temperature range of 0-140 °C and a time range of 0 Ma to twice the age of the AFT age was used. Individual thermal history models were run for 150,000 iterations, with the first 50,000 discarded as burn-in [Gallagher *et al.*, 2009]. To ensure satisfactory sampling, the acceptance rates of the proposed model parameters and the posterior sampling chain were checked after each modelling run [Gallagher *et al.*, 2012]. The resulting models displayed below are the expected time-temperature histories (weighted mean of the posterior distribution), deemed to be most representative sample thermal histories [see Gallagher, 2012 and Wildman *et al.*, 2016 for discussion].

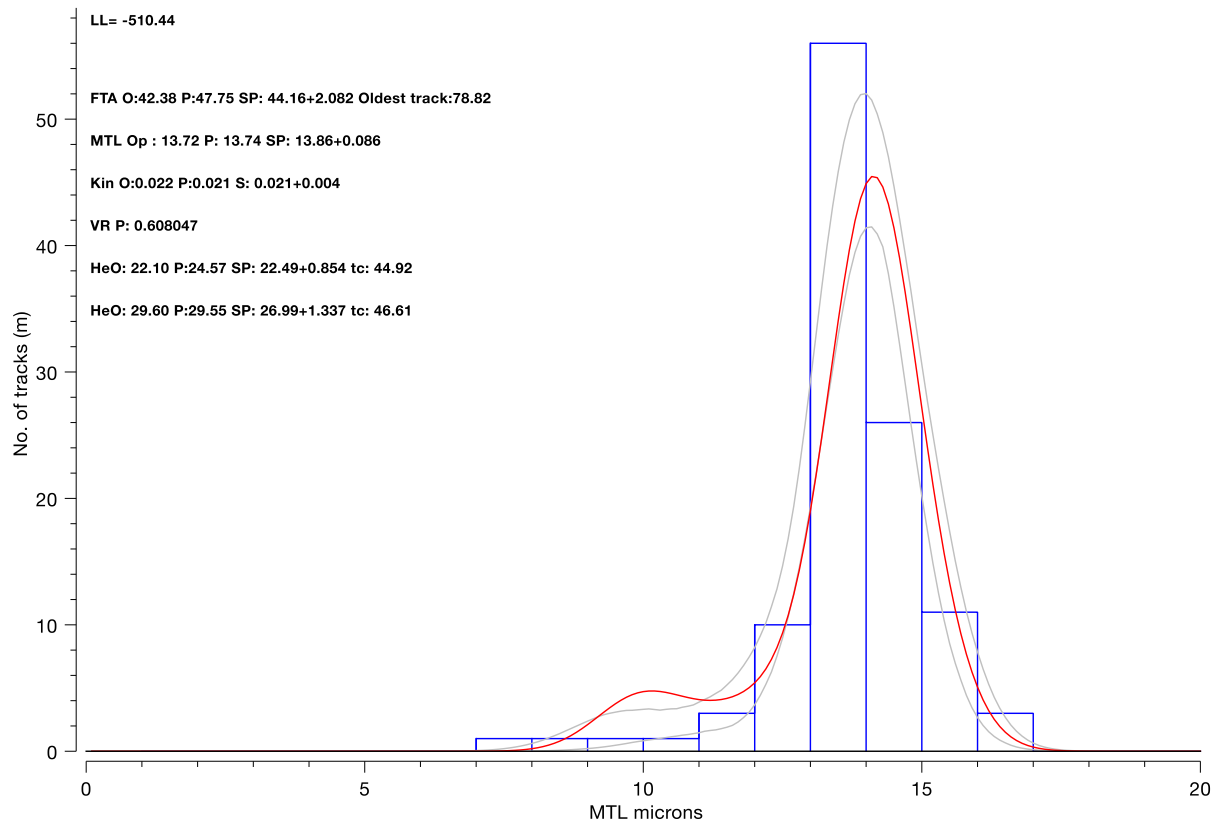
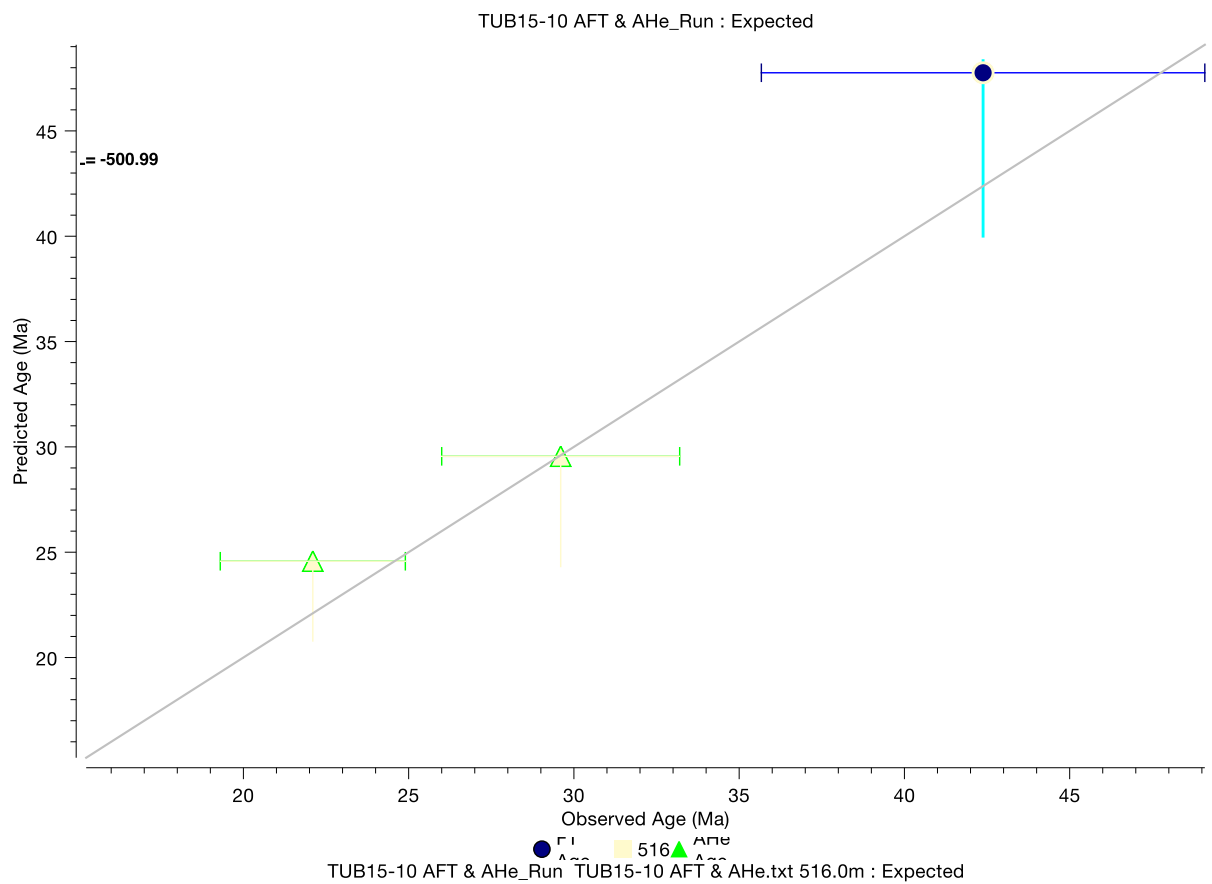
Only samples with a statistically significant number of confined track length measurements (> 100) were modelled. 0-termination AHe grains were omitted during modelling due to the current implementation of the Brown *et al.* [2013] fragmentation model's inability to process 0-termination grains. A sole time-temperature constraint was applied, requiring samples to be at near-surface temperatures (20 ± 10 °C) by the present-day. Plots of observed versus predicted thermochron ages and confined track distributions are displayed below each thermal history model, allowing for the data fit of the models to be assessed. Samples 11 and 30 did not produce suitable AHe data for thermal history modelling. Thermal history models generated for samples 20 and 24 using AFT and AHe data are displayed in Figure 3.7.

Sample 8

TUB15-08 AFT & AHe_Run : TUB15-08 AFT & AHe.txt







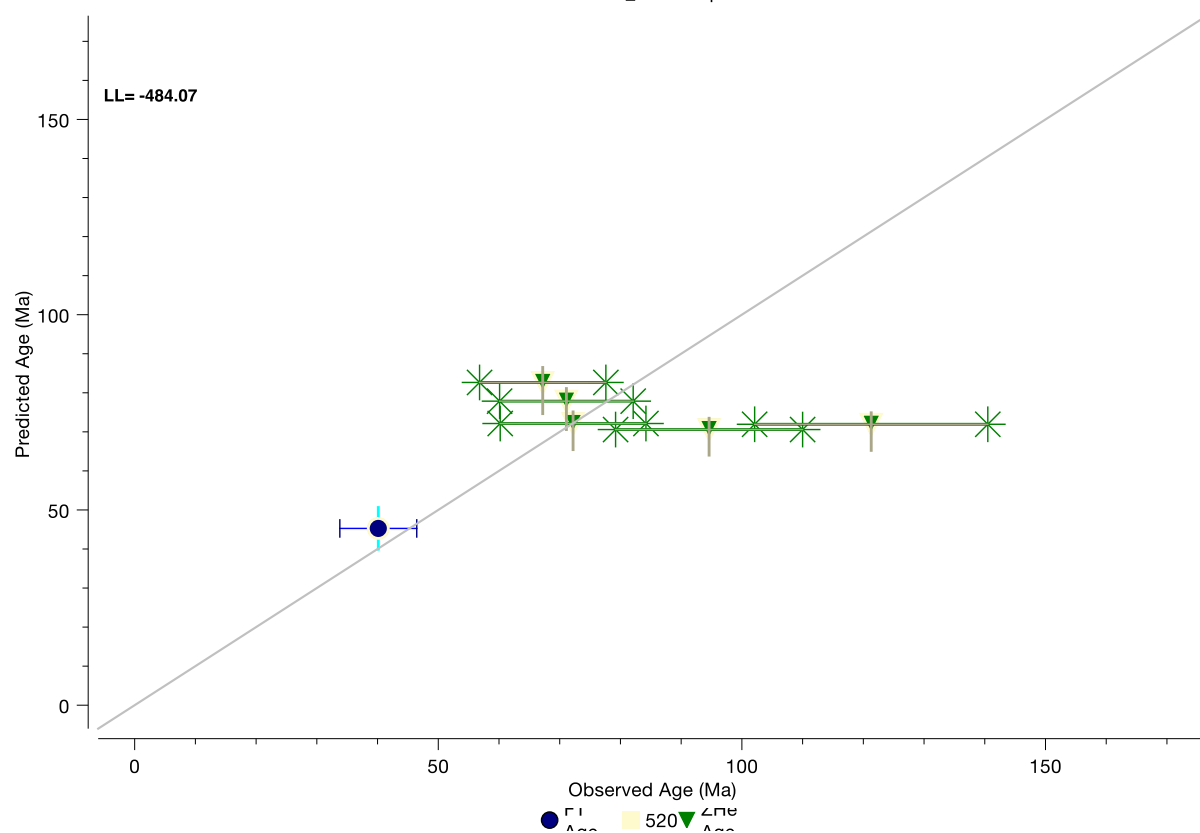
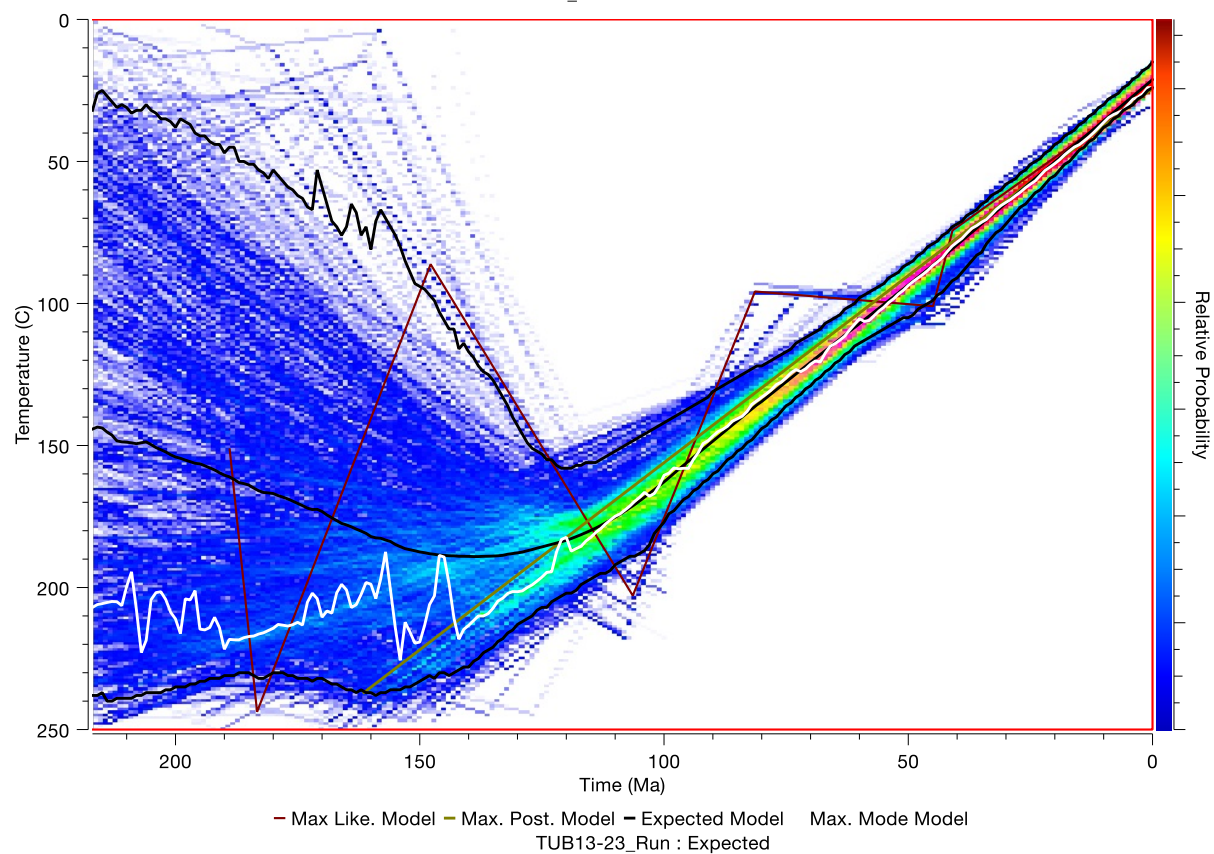
Supplementary Data C: Thermal history models using ZHe, AFT and AHe data deemed non-robust

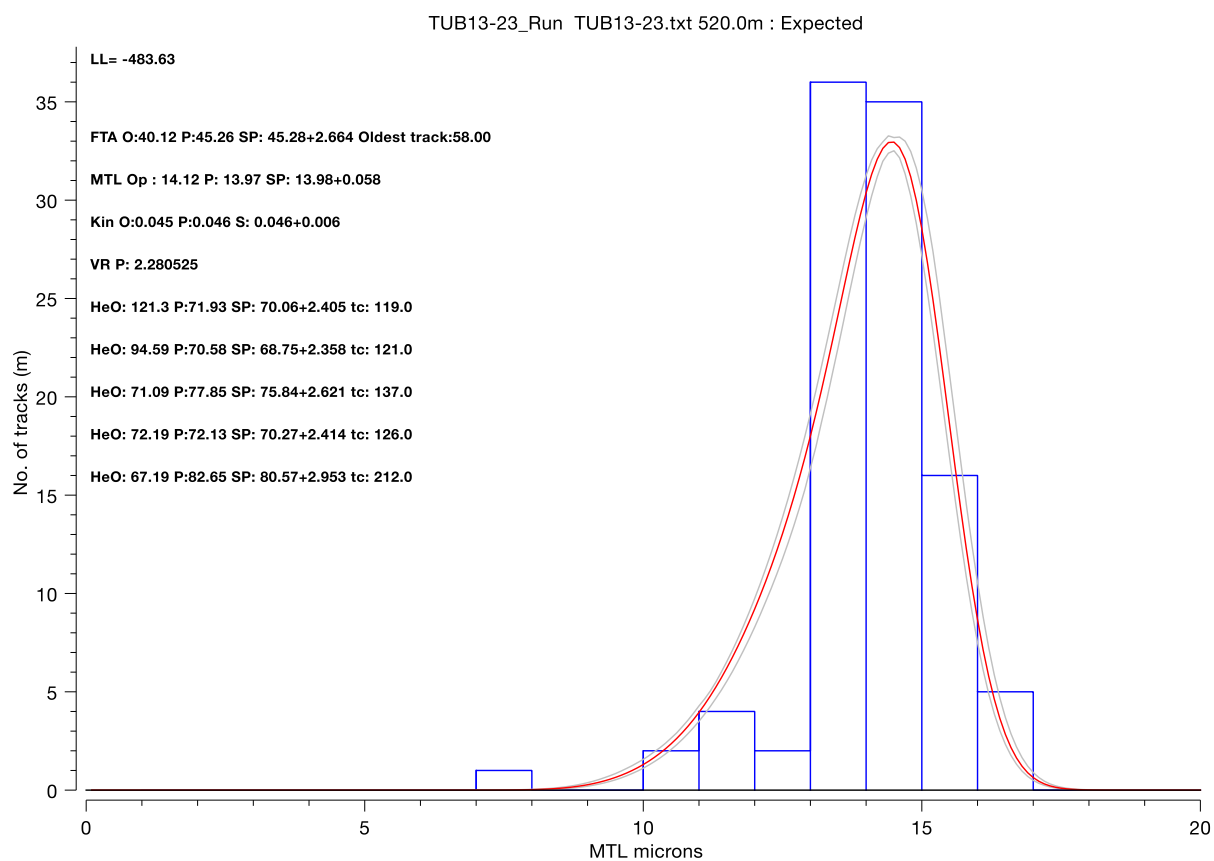
Thermal history models displayed below were generated in the software *QTQt* using ZHe, AFT and AHe data, following the procedure outlined in section 3.6. The Bayesian transdimensional Markov chain Monte Carlo (MCMC) employed by *QTQt* requires the user to define a time-temperature range space in which the MCMC samples [Gallagher *et al.*, 2009; Gallagher, 2012]. When performing thermal history modelling of ZHe, AFT and AHe data (when available), a temperature range of 0-250 °C and a time range of 0 Ma to twice the age of the oldest thermochronometer age was used. Individual thermal history models were run for 150,000 iterations, with the first 50,000 discarded as burn-in [Gallagher *et al.*, 2009]. To ensure satisfactory sampling, the acceptance rates of the proposed model parameters and the posterior sampling chain were checked after each modelling run [Gallagher *et al.*, 2012]. The resulting models displayed below are the expected time-temperature histories (weighted mean of the posterior distribution), deemed to be most representative sample thermal histories [see Gallagher, 2012 and Wildman *et al.*, 2016 for discussion].

Only samples with a statistically significant number of confined track length measurements (> 100) were modelled. 0-termination AHe grains were omitted during modelling due to the current implementation of the Brown *et al.* [2013] fragmentation model's inability to process 0-termination grains. A sole time-temperature constraint was applied, requiring samples to be at near-surface temperatures (20 ± 10 °C) by the present-day. Plots of observed versus predicted thermochron ages and confined track distributions are displayed below each thermal history model, allowing for the data fit of the models to be assessed. Thermal history models generated for samples 23 and 24 using ZHe, AFT and AHe data were deemed non-robust due to the Guenthner *et al.* [2013] He diffusion model's inability to reproduce the spread in ZHe ages and their inconsistency with robust thermal histories generated from neighbouring samples.

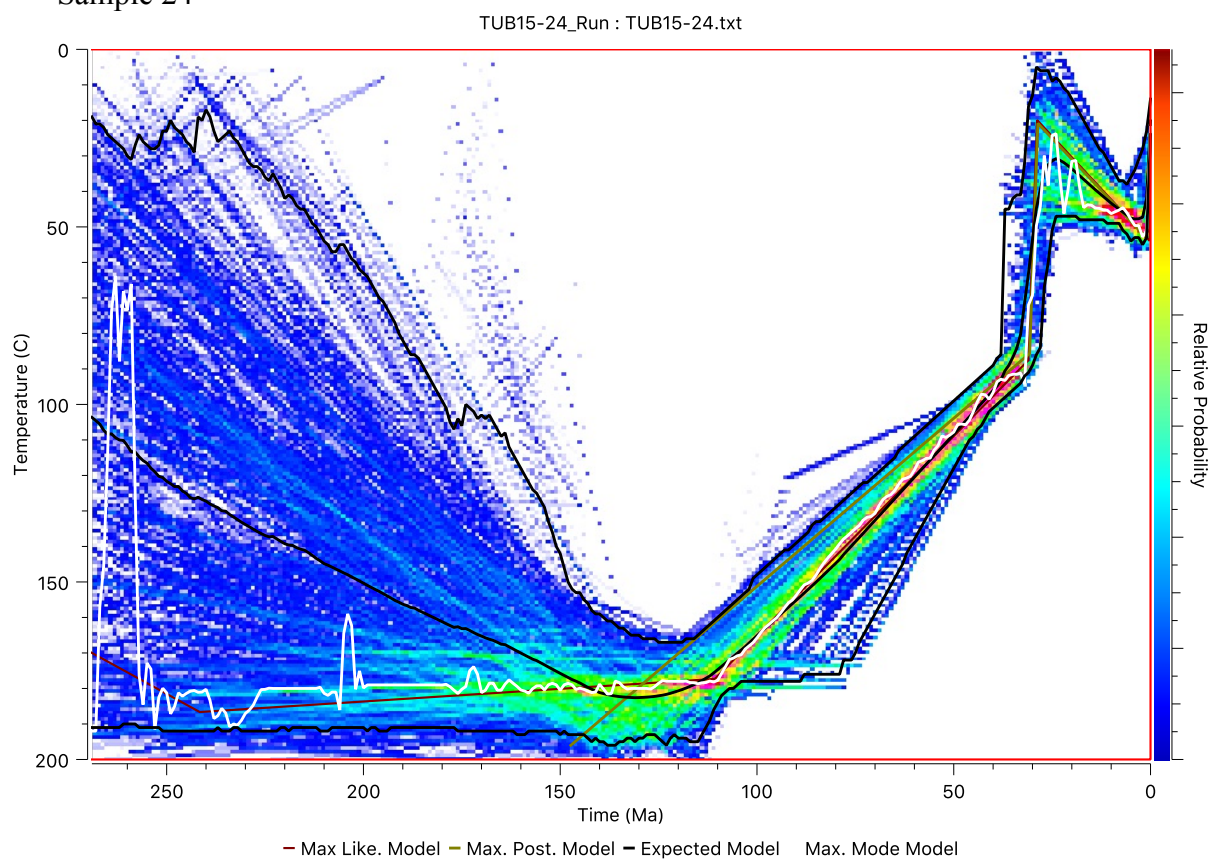
Sample 23

TUB13-23_Run : TUB13-23.txt





Sample 24



The role of inherited crustal fabrics in the structural evolution of the Broadly Rifted Zone, Ethiopian Rift: A multi low-temperature thermochronometer study

Abstract

The Broadly Rifted Zone (BRZ) of southern Ethiopia is a long-lived and structurally complex segment of the East African Rift System (EARS). Here, deformation is accommodated across a diffuse (~300 km) region of basins and ranges in stark contrast to the narrow (<100 km) rift valley morphology exhibited by the Main Ethiopian Rift (MER) to the north. Thus, constraining the structural evolution and inception of rifting in the BRZ is critical for understanding the evolution of the greater EARS. However, due to a scarcity of geological and structural studies and lack of published subsurface data, the tectonic evolution of the BRZ remains poorly understood. In this study, we present new apatite (U-Th-Sm)/He track low-temperature thermochronology data from Beto and Galana basin boundary fault systems to constrain and compare the tectono-thermal evolution of the western and eastern BRZ, respectively. In addition, existing apatite fission track data from the region are augmented by new confined track measurements, allowing for more rigorous thermal history modelling. The resulting time-temperature reconstructions suggest that EARS-related extension began concurrently across the width of the BRZ in the early Miocene, at least 7 Myr prior to the formation of the late Miocene MER further north. Moreover, the increased time-temperature resolution provided by combined AFT and AHe thermochronology also reveals spatiotemporal variations in basin margin denudation patterns, thus providing insights into master fault system growth and displacement trends. The Beto Basin boundary fault system originated initially as isolated normal fault segments, which accumulated substantial displacement prior to linking and forming a through-going fault array between 14-6 Ma. By contrast, the master fault system bounding the Galana Basin approached a substantial portion of its full length prior to the accumulation of significant displacement. Our findings in the BRZ highlight the importance of lithospheric rheological heterogeneities in the evolution of intracontinental rifts. The anomalous tectonic evolution of the BRZ compared to the greater Ethiopian Rift, both in its earlier onset and its wider distribution of deformation, is likely a product of inherently weaker lithosphere, thinned by earlier Cretaceous-early Paleogene Anza-South Sudan rifting. Furthermore, the longitudinal contrast in the structural evolution of the BRZ reflects differences in the regional distribution of pre-existing basement fabrics. The trend of the highly-segmented Beto Basin master fault reflects the local foliation pattern in the gneissic basement, while the long, linear fault array bounding the Galana western margin mirrors the ~N-S trending, deep-seated lithospheric heterogeneities inherited from the Neoproterozoic East African Orogeny. This strongly suggests that the disparate structural evolution of the BRZ was controlled by pre-existing lithospheric heterogeneities.

4.1. Introduction

Continental rifts are elongate regions of extended lithosphere bounded by normal fault systems. These fault arrays accommodate km-scale displacement, resulting in the formation of sedimentary basins. Normal faults grow by elongation, accumulation of slip and linkage with other fault segments [Cartwright *et al.*, 1995; Fossen and Rotevatn, 2016; Kim and Sanderson, 2005; Walsh *et al.*, 2002]. However, no consensus currently exists as to how fault length and displacement respectively increase over time and whether fault segment linkage occurs prior to or during the accumulation of significant slip. Field observations of ancient faults reveal that these systems can develop via a spectrum of growth mechanisms, between two end-member growth models equally applicable in nature [Cartwright *et al.*, 1995; Gawthorpe *et al.*, 1997; Gupta *et al.*, 1998; Morley, 1999c; Sharp *et al.*, 2000; Kim and Sanderson, 2005; Fossen and Rotevatn, 2016]. One is the segment growth and linkage model [Anders and Schlische, 1994; Trudgill and Cartwright, 1994], where initially isolated fault segments elongate and accumulate displacement via simple tip propagation before eventually linking to form a through-going fault system. Whereas, a second is the constant length model [Walsh *et al.*, 2002, 2003], where a fault array approaches or attains its full length prior to accumulating significant displacement.

Field studies and numerical and analogue models have shown that the way in which normal fault systems grow reflects lithospheric mechanical, thermal and kinematic properties and rift boundary conditions [e.g. Bürgmann *et al.*, 1994; Steen and Andresen, 1999; Wilkins and Gross, 2002; Childs *et al.*, 2009; Jackson and Rotevatn, 2013]. Thus, investigations into the evolution of boundary fault systems in extensional tectonic settings can provide valuable information regarding paleo-plate boundary conditions and the rheological structure of the deformed lithosphere during rift evolution. Moreover, the migration of activity along and between these basin bounding fault systems provides insight into rift-scale evolutionary trends and strain localisation [Gawthorpe *et al.*, 2003].

Numerous studies have shown that the growth of normal fault arrays are reflected in the evolution of extensional basins [Schlische, 1991; Gawthorpe *et al.*, 1994; Schlische and Anders, 1996; Gupta *et al.*, 1998], governing the spatially and temporally variable creation of

hanging wall accommodation space and footwall uplift [*Jackson and Mckenzie*, 1983; *Walsh and Watterson*, 1988; *Cowie et al.*, 2000; *Dawers and Underhill*, 2000]. However, as the process of displacement profile readjustment conceals the relative timing of strain accommodation, the temporal evolution of fault systems is difficult to decipher from structural analysis of their final configuration alone [*Commins et al.*, 2005]. Accordingly, alternative methods, such as sequence stratigraphy [e.g. *Gawthorpe et al.*, 1997; *Davies et al.*, 2000; *Dawers and Underhill*, 2000] and seismic techniques [e.g. *Morley*, 1999; *Morley and Wescott*, 1999; *Mcleod et al.*, 2000], are often used to help constrain spatio-temporal displacement accumulation trends. However, in instances where little control on basin stratigraphy exists and sub-surface data are unavailable, the predictable pattern of footwall uplift and denudation trends in response to lateral fault growth and displacement accumulation can be used instead [*Harbor*, 1997; *Densmore et al.*, 2004, 2007].

Such is the case for the Broadly Rifted Zone (BRZ) of southern Ethiopia (Figure 4.1), a long-lived and structurally complex part of the greater East African Rift System (EARS). While the EARS is primarily comprised of a narrow (~100 km) corridor of graben and half-graben rift basins middle-late Miocene in age, active deformation in the BRZ is diffused over a much broader region (as wide as ~300km) in a series of basins and ranges that extends into the Turkana Depression of northern Kenya [*Baker et al.*, 1972; *Moore Jr and Davidson*, 1978]. Moreover, field mapping, structural geology and thermochronological studies suggest that EARS extensional deformation began earlier in the Turkana Depression (late Paleogene) and BRZ (early Miocene) than in the Main Ethiopian Rift (MER) and Kenyan Rift to the north and south [*Boone et al.*, in review; *Woldegabriel and Aronson*, 1987; *Morley et al.*, 1992, 1999b, *Ebinger et al.*, 1993, 2000; *Pik et al.*, 2008]. Unlike the MER and Kenyan Rift, where volcano-tectonic activity is well documented, the structural evolution of the BRZ by comparison is poorly constrained due to a relative scarcity of regional geological and structural observations and total lack of published subsurface data.

Low-temperature thermochronometers, such as apatite fission track (AFT) analysis, provide alternative, yet powerful, tools to constrain periods of rift basin development by determining the timing of denudation and cooling associated with rift-related footwall uplift [e.g. *Fitzgerald*, 1992; *Stockli*, 2005b; *Storti et al.*, 2008]. While the well-understood behaviour of the AFT system makes it a robust method for constraining the timing and degree of cooling related to extensional tectonism [*Kohn et al.*, 2005], in certain cases its temperature

sensitivity ($\sim 60\text{-}120\text{ }^{\circ}\text{C}$) can be too high to adequately record small-amplitude variations in footwall cooling histories associated with normal fault slip [Brown *et al.*, 2017]. Such was the case for previous thermochronology studies in the BRZ [Philippon *et al.*, 2014; Balestrieri *et al.*, 2016], where AFT data provided minimum age constraints for the formation of the Beto Basin of the western BRZ (between 14-10 Ma) and the Galana Basin ($\sim 10\text{-}8$ Ma), bounded by the Amaro Horst further east. However, the onset of rift-related denudation for many samples, such as those from the southern Amaro Horst, was poorly constrained due to Neogene exhumation rates being insufficient to be recorded by the temperature sensitivity of the AFT system. The lower temperature sensitivity apatite (U-Th-Sm)/He (AHe) system ($\sim 30\text{-}90\text{ }^{\circ}\text{C}$) provides an alternative thermochronometer better suited to be able to record low-magnitude tectono-thermal processes in the upper $\sim 1\text{-}2$ km of the crust that might otherwise be poorly resolved by AFT analyses alone [Stockli *et al.*, 2000; Balestrieri *et al.*, 2005].

AHe thermochronology has been widely employed to constrain the timing of normal fault activity in extensional settings [e.g. Colgan *et al.*, 2006; Fosdick and Colgan, 2008; Gorynski *et al.*, 2013; Brown *et al.*, 2017]. Moreover, some authors have utilized its sensitivity to relatively low-amplitude variations in cooling histories as a tool to constrain the mechanisms by which normal fault systems evolve. For example, Curry *et al.* [2016] presented a framework to interpret along-strike spatiotemporal variations in the onset of normal fault footwall denudational cooling to test conceptual fault growth models. While, variations in AHe cooling ages in the Malawi Rift were used by Mortimer and colleagues [2016] to constrain the development, interaction and linkage of paleo-fault segments.

Here, we present combined AFT and AHe analyses and thermal history modelling results from the Beto and Galana Basins (Amaro Horst region) from the western and eastern BRZ, respectively. These data reveal along-strike footwall cooling history trends for the fault systems bounding these basins, providing new constraints for the propagation of rifting in the BRZ. Moreover, the resulting high-resolution time-temperature reconstructions reveal along-strike differences in the onset of cooling as well as variations in relative cooling rates over time, associated with fault array growth and strain migration patterns during the development of these grabens. The mechanical, thermal and kinematic parameters, which controlled the structural evolution of the BRZ are also explored.

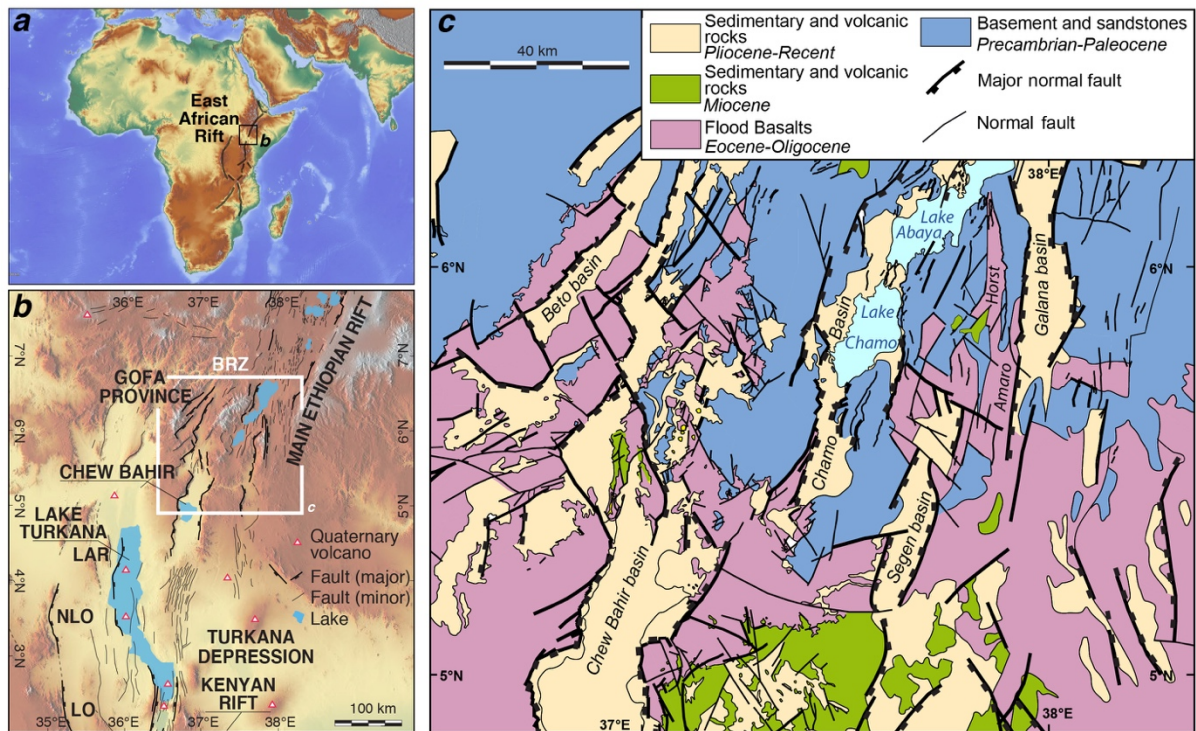


Figure 4.1. Overview maps of the study area illustrating its geographical context, complex structural architecture and geology. **a:** Topographic map of Africa showing location of the East African Rift System (EARS). **b:** Topographic map of the Broadly Rifted Zone (BRZ) and Turkana Depression, showing their spatial relationship with the narrow Main Ethiopian Rift (MER) and Kenyan Rift. **c:** Simplified geological map of the BRZ with the Beto Basin in the west and Amaro Horst in the east. LAR = Lapur Range; LO = Lokichar Basin; NLO = North Lokichar Basin. Geological map after Davidson [1983], Mengesha et al. [1996], Ebinger et al. [2000a], Philippon et al. [2014] and Balestrieri et al. [2016].

4.2. Geological Setting

Since attaining its modern terrane configuration during the late Proterozoic-early Paleozoic Pan-African Orogeny [Kröner and Stern, 2005], East Africa has been the centre of a series of continental break-up events [Kreuser, 1995]. During Cretaceous-early Paleogene time, the kinematically linked, NW-SE trending Anza and South Sudan rifts developed in association with the opening of the Atlantic and Indian Oceans and rifting of Madagascar from Africa's eastern margin [Reeves et al., 2016]. The Anza Rift is expressed by an ~600 km long zone of extended crust stretching from the Kenyan coast to the eastern shores of Lake Turkana, in northern Kenya (Figure 4.1) [Reeves et al., 1987; Guiraud and Maurin, 1992; Morley et al., 1999a]. To the west, a series of parallel Cretaceous-Paleogene basins, known collectively as the South Sudan Rifts, narrow as they approach the South Sudanese-Kenyan border [Browne and Fairhead, 1983]. Despite a lack of data in eastern South Sudan and the obscuring effect

of subsequent late Paleogene-Recent tectonics and volcanism in northern Kenya [Guiraud and Bosworth, 1997], the compatible geometries and coincident development of the Anza and South Sudan rifts suggest a possible link between the two systems west of Lake Turkana [Schull, 1988; Ibrahim *et al.*, 1991].

4.2.1 The late Paleogene-Recent East African Rift System (EARS)

The late Paleogene-Recent EARS is widely considered the archetypal magma-assisted continental rift, having developed diachronously above one or multiple Paleogene mantle plume(s) [e.g. Ebinger and Sleep, 1998; George *et al.*, 1998; George and Rogers, 2002; Rogers, 2006; Civiero *et al.*, 2016]. However, the continuity, location and depth of the asthenospheric thermal upwelling remain controversial [Ebinger and Sleep, 1998; Weeraratne *et al.*, 2003; Simmons *et al.*, 2007; Fishwick and Bastow, 2011]. Beneath Ethiopia, a mantle slow-velocity anomaly has been detected rising from depths exceeding 650 km to as shallow as 75 km beneath the MER, where it impinges beneath the crust ~25 km west of rift axis [Bastow *et al.*, 2005, 2011; Benoit *et al.*, 2006; Fishwick, 2010].

The ~N-S trending EARS extends more than 3,500 km southwards from the Red Sea (Figure 4.1a) and is composed of two branches, the eastern branch forming the Ethiopian and Kenyan Rifts and the arcuate western branch comprised of a series of elongate lakes running from Uganda to Mozambique. These largely follow a trend of Precambrian mobile belts formed during the Pan-African Orogeny, such as the Neoproterozoic Mozambique Belt running from Ethiopia south to Mozambique [Morley *et al.*, 1999c; Kröner and Stern, 2005].

The Ethiopian Rift is a key sector of the EARS, defining the incipient boundary between the Nubian and Somalian plates [Chorowicz, 2005]. The Ethiopian Rift developed within the Mozambique Belt, which contains numerous N-NNE trending sutures interpreted to mark an important crustal boundary between juvenile crust to the west from reworked Paleoproterozoic-Archean crust to the east [Berhe, 1990; Stern *et al.*, 1990; Stern, 2002]. Analyses of the modern day stress field suggest a current ~E-W (~N100°E) extension direction between the Nubian and Somalian plates [Bosworth *et al.*, 1992; Foster and Jackson, 1998; Fernandes *et al.*, 2004; Bendick *et al.*, 2006; Keir *et al.*, 2006]. Although, it is disputed whether these kinematic boundary conditions have remained constant during the entirety of rift evolution [Corti, 2009], with some authors arguing for a poly-phase history involving a change in plate motion direction at some time during the interval 6.6-2.6 Ma

[*Bonini et al.*, 1997, 2005; *Boccaletti et al.*, 1998; *Wolfenden et al.*, 2004]. More recent studies however, suggest that the ~E-W extension direction in the Ethiopian Rift has remained constant since at least 11 Ma [*Royer et al.*, 2006].

Along its ~ 1000 km length, the Ethiopian Rift encompasses multiple rift segments at varying stages of continental breakup, each with its own distinct structural style [*Ebinger*, 2005]. The northern end is defined by the Afar depression, a wide region characterised by embryonic oceanic spreading centres ($\geq 100\%$ extension) [*Barberi et al.*, 1972; *Hayward and Ebinger*, 1996]. Basaltic volcanism began in northern Ethiopia as early as 32-30 Ma, however the main phase of rift basin development began much later ca. 20-15 Ma [*Barberi et al.*, 1972; *Davidson and Rex*, 1980; *Hayward and Ebinger*, 1996]. Southwards, the Ethiopian rift narrows to form an ~100 km wide rift valley separating the uplifted western Ethiopian and eastern Somalian plateaux referred to as the MER (Figure 4.1b) [*Mohr*, 1983]. This narrow corridor comprises a series of long, widely spaced and discontinuous boundary faults with large vertical offsets [> 1 km; *Boccaletti et al.*, 1998] that define a series of middle to late Miocene-Recent grabens and half-graben basins [*Bonini et al.*, 2005; *Corti*, 2009; *Abebe et al.*, 2010]. Here, the crust thickens southwards along the rift axis from 24-26 km below the southern Afar region to 38-40 km in the central MER [*Dugda et al.*, 2005; *Maguire et al.*, 2006; *Mickus et al.*, 2007].

South of Lake Abaya (Figure 4.1c), the zone of rifting broadens again as it transitions into the BRZ, encompassing an ~300 km wide region of basins and ranges [*Baker et al.*, 1972; *Moore and Davidson*, 1978; *Ebinger et al.*, 2000]. This area marks the transition from the high elevations of the Ethiopian Dome (averaging 1500 m asl) to the topographically subdued Turkana Depression of northern Kenya (averaging ~600 m asl). In addition, the BRZ is characterized by an increase in physiographic complexity [*Ebinger et al.*, 2000]. To the west, the BRZ is composed of the Gofa Basin and Range, or Gofa Province, composed of a series of ~NE-SW trending blocks of Precambrian basement and Tertiary volcanic rocks tilting to the northwest [*Moore Jr and Davidson*, 1978]. These are bounded by highly segmented, SE-dipping normal faults [*Davidson*, 1983; *Philippon et al.*, 2014]. While further east in the Amaro Horst region, the BRZ is made up of more ~N-S trending basins delimited by few, widely spaced normal faults [*Ebinger et al.*, 1993]. In general, the subsurface rift architecture of the BRZ is poorly constrained due to the lack of detailed geophysical data. Nonetheless, low-temperature thermochronology analyses [*Philippon et al.*, 2014; *Balestrieri et al.*, 2016]

and geochronological constraints [Woldegabriel *et al.*, 1991; Ebinger *et al.*, 1993, 2000a; Bonini *et al.*, 2005] suggest a middle Miocene initiation of major rifting within the BRZ. However, Pik *et al.* [2008] proposed an even earlier onset of rifting during early Miocene time in the southern BRZ based on AHe analyses from the Chew Bahir Basin.

4.2.2 Beto Basin

The Gofa Province is comprised of a series of ~NE-SW trending, west-dipping half-grabens that host alluvial and lacustrine sediments. In general, these basins widen towards the south, resulting in slightly triangular shapes in map view (Figure 4.1c) [Ebinger *et al.*, 2000].

One of the more prominent half-grabens in the Gofa Province is the Beto Basin (Figure 4.2). Its north-western margin is bounded by a ~N45°E striking master fault that extends more than 40km, dips steeply towards the southeast, has a maximum throw of more than 3 km [Davidson, 1983] and its fault plane exhibits clear dip-slip kinematic indicators [Philippon *et al.*, 2014]. Paleostress inversion of data from the basin-bounding master fault indicates an average extensional direction of N105°E [Philippon *et al.*, 2014]. The Beto Basin southwest margin is bounded by an ~N150°E striking transverse fault.

4.2.2.1 Stratigraphy

Faulting in the Beto Basin has exposed Precambrian basement (gneisses, amphibolites, and granulites) believed to have been deformed during the late Pan-African Orogeny [Levitte *et al.*, 1974; Davidson, 1983; Yemane and Yohunie, 1987]. Basement rocks are unconformably overlain by the 45-35 Ma Amaro-Gamo basalts and Amaro tuffs [Ebinger *et al.*, 1993], which reach a thickness > 1100 m in the Shengema Mountains to the northwest of the basin. In the footwall of the basin bounding fault, this trap series dips between ~20-40° [Philippon *et al.*, 2014].

4.2.3 Amaro Horst region

At approximately latitude 6° N as the MER transitions into the eastern BRZ, the rift valley bifurcates south of Lake Abaya into two near-parallel grabens, the western Chamo (or Ganjuli) Basin and eastern Galana Basin (Figure 4.1c). These basins are separated by the ~N-S trending Amaro Horst, a narrow (~20 km) fault-bounded block of Precambrian basement.

The ~40 km wide Chamo Basin is bounded to the east by a series of large step faults, whose variable orientations (NW-SE to NE-SE) give rise to a distinct zig-zag pattern in map view (Figures 4.1c & 4.2). The Chamo Basin is thought to have a symmetric geometry, with syn-rift deposits estimated to reach thicknesses of up to 2 km [Ebinger *et al.*, 1993].

By contrast, the ~30 km wide Galana Basin has an overall asymmetric geometry, bounded on its western margin by a series of ~N0°-25°E trending, east-dipping step faults that rises to the 3200m high Amaro Horst (Figure 3) [Philippon *et al.*, 2014]. Here, the basin fill may reach a maximum thickness of more than 4 km [Ebinger *et al.*, 1993]. To the east, the Galana Basin is bounded by a series of W-dipping, ~N-S trending normal faults with small throws and WNW-dipping monoclines [Ebinger *et al.*, 1993]. Paleostress inversion of structural data from this region are consistent with an ~N94°E extension direction [Philippon *et al.*, 2014].

4.2.3.1 Stratigraphy

In the Amaro Horst region, Precambrian basement rocks are unconformably overlain by discontinuous, thin (<1 m) patches of laterized gritstone and conglomerate [Levitte *et al.*, 1974]. The few outcrops of these highly eroded sedimentary rocks are too small to be shown in Figure 4.2. The age of these red beds is unknown, however a late Mesozoic age has been suggested by Levitte *et al.* [1974] by correlation with well-dated laterites from central Ethiopia [Mohr, 1962] and lithologically similar deposits from northern Turkana [Arambourg and Wolff, 1969]. Where present, the Amaro-Gamo basalts and Amaro tuffs (45-35 Ma) unconformably overlie Precambrian basement and the red bed regolith [Ebinger *et al.*, 1993]. These, in turn, are unconformably overlain by the 18-11 Ma Getra-Kele basalts, which are absent in the Gofa Province further west [Woldegabriel *et al.*, 1991; Ebinger *et al.*, 1993]. A series of ~14 Ma trachytic-syenitic volcanic plugs, named the Mymo trachytes, intrude basement and overlying Amaro-Gamo basalts along the fault systems bounding the Chamo and Galana Basins [Ebinger *et al.*, 1993].

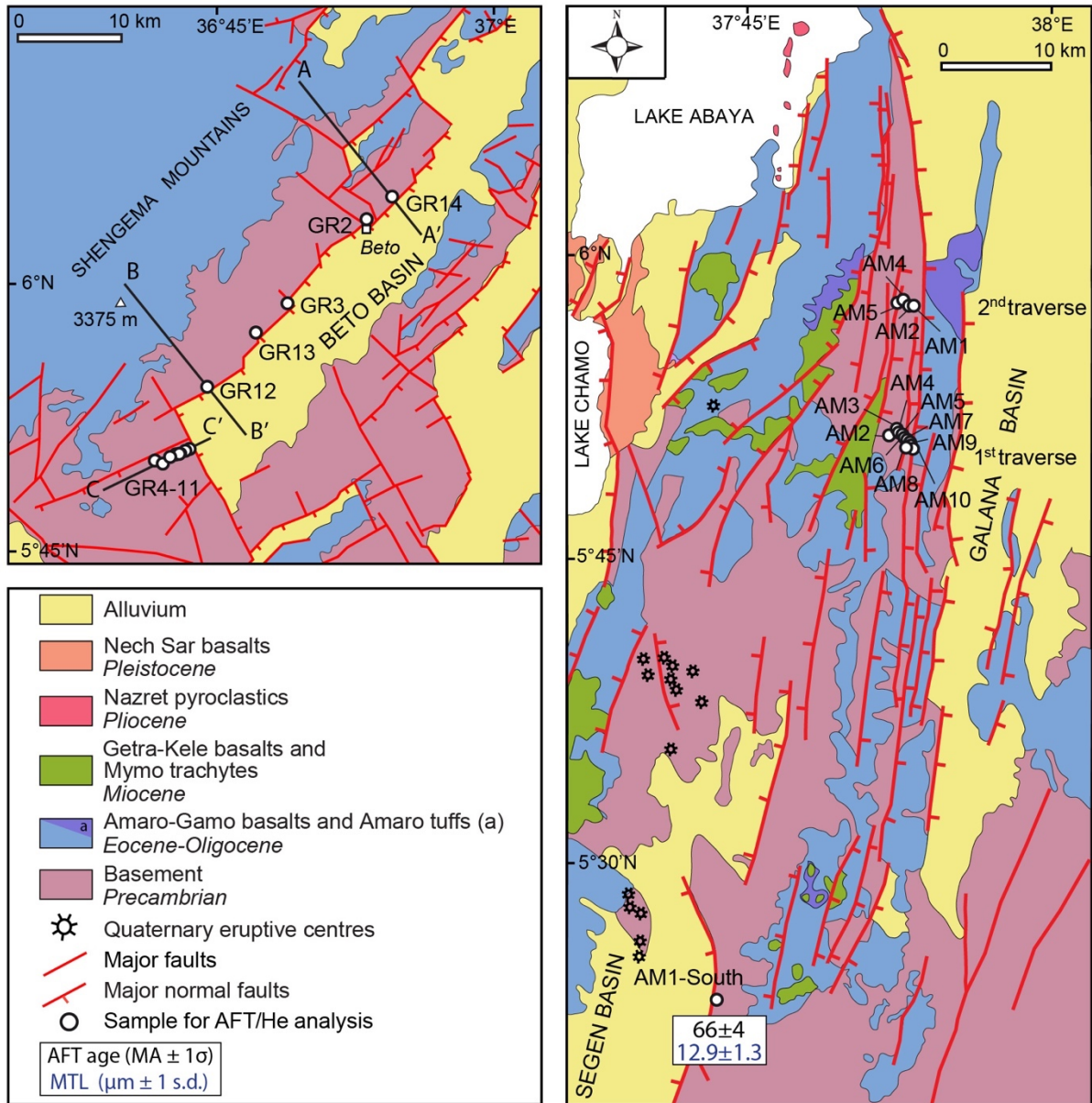


Figure 4.2. Geological maps of the Beto Basin and Amaro Horst regions with sample localities shown. The AFT age [Ma $\pm$ 1 σ] [Balestrieri et al., 2016] and mean track length (MTL; this study) [μ m $\pm$ 1 s.d.] data for sample AM1-South are shown. Remaining AFT data from this and previous studies are shown in Figure 4.3. Geological maps after Balestrieri et al. [2016], Davidson [1983], Ebinger et al. [1993], Levitte et al. [1974] and Woldegabriel et al. [1997].

4.3. Methodology

4.3.1 Sample collection and analyses

Thirteen Precambrian granitoid and gneissic basement samples from the Beto Basin, Gofa Province, and fifteen basement rocks from the Amaro Horst, between the Galana Basin to the east and Chamo and Segen Basins to the west, were collected to constrain the low-temperature thermal history of these areas of the Broadly Rifted Zone (BRZ) of southern Ethiopia (Tables 4.1-3 and Figures 4.2 and 4.3).

Field sampling in the Gofa Province was conducted along the northwest Beto Basin margin, with five samples collected along the base of the SE-dipping master fault scarp. An additional eight samples were collected in a vertical profile in the footwall of the transverse fault defining the southwest basin margin.

Sampling of the Amaro Horst consisted of two vertical transects up its eastern flanks, along the western margin of the Galana Basin. The first traverse consisted of nine samples between elevations of 1800 and 2420 m, which crossed a prominent east-dipping normal fault of unknown displacement between samples AM6-*1sttr* (2080 m) and AM7-*1sttr* (2000 m) [Balestrieri *et al.*, 2016]. For ease of discussion, samples collected west, or above, this observed fault (AM2-*1sttr* to AM6-*1sttr*) are referred to herein as upper 1st traverse samples. Accordingly, samples collected east, or below, the observed fault (AM7-*1sttr* to AM10-*1sttr*) are referred to as lower 1st traverse samples. Four additional samples were collected as part of the 2nd traverse between elevations of 1685 and 2004 m, 11 km to the north. Two additional basement samples were collected on the western side of the Amaro Horst. AM1-*West* was collected from the western base of the Amaro Horst where it defines the eastern margin of Chamo Basin, at the same approximate latitude as the 1st traverse. AM1-*South* was sampled ~50 km SSW of the 1st traverse at the base of western Amaro Horst, along the eastern margin of the Segen Basin.

4.3.2 Apatite fission track (AFT) and (U-Th-Sm)/He (AHe) methodologies

The AFT methodology is based on damage trails in apatite, called fission tracks, which form as a consequence of spontaneous fission of ²³⁸U [Fleischer *et al.*, 1975; Wagner and Van den Haute, 1992]. Fission tracks are temperature sensitive between ~60 and ~110-120 °C,

referred to as the partial annealing zone (PAZ), over geological timescales [Gleadow and Duddy, 1981; Laslett *et al.*, 1987]. The compositional effect on fission track annealing kinetics can be assessed by a proxy measurement of the diameter of etched spontaneous track openings parallel to the apatite crystallographic c-axis (Dpar) [Sullivan and Parrish, 1995; Barbarand *et al.*, 2003]. By combining fission track age and mean track length (MTL) data the thermal history of a sample within the PAZ can be determined [Gallagher, 1995; Ketcham, 2005].

AHe thermochronology is based upon the production of ^4He during the α -decay of ^{238}U , ^{235}U , ^{232}Th and ^{147}Sm [Farley and Stockli, 2002]. Volume diffusion of ^4He is a function of time, temperature, crystal size and accumulated radiation damage [Farley, 2000; Reiners and Farley, 2001; Shuster *et al.*, 2006]. In apatite, ^4He diffusion accelerates significantly above $\sim 30^\circ\text{C}$ and nearly instantaneously above $\sim 90^\circ\text{C}$ [Reiners *et al.*, 2005; Reiners and Brandon, 2006]. Within these temperatures, referred to as the apatite partial retention zone (AHe PRZ), ^4He is partially retained.

Combined AFT and AHe thermochronology provides a powerful tool for constraining the thermal history of apatite-bearing rocks between ~ 30 - 120°C , approximately corresponding to the upper 2-5 km of the crust. Inverse modelling of AFT and AHe results (Section 6) allows for the identification of the most plausible time-temperature evolution recorded by those data [Gallagher, 1995]. The resulting thermal history reconstructions in turn provide important insights into the denudational and tectonic histories of geological terrains [Kohn *et al.*, 2005].

A comprehensive review of the AHe methodology and technical procedure used in this study can be found in Gleadow *et al.* [2015]. A detailed summary of the AFT and AHe analytical protocols are given in the footnotes of Tables 4.1 (AFT), 4.2 (Beto Basin AHe) and 4.3 (Amaro Horst AHe). The thermal history modelling approach applied to these data, along with a discussion of the time-temperature reconstructions, are presented in Section 4.6.

Samples AM2-1sttr, AM1-West and AM4-2ndtr yielded insufficient apatite grains for AHe experimentation, while AHe analyses was not attempted on samples GR5 to 10. For samples that did yield suitable grains for AHe analyses, an alpha ejection correction (F_T) was applied to calculate a corrected age (Tables 4.2 & 4.3), as suggested by Farley *et al.* [1996].

However, it has been shown that for samples with protracted thermal histories in the PRZ, as is the case for slowly cooled basement terranes, volume diffusion is the dominant mechanism governing He loss. Thus, the application of F_T in these cases may result in inaccurately old ages [Meesters and Dunai, 2002; Danišik *et al.*, 2008]. Therefore, all AHe data discussed in the text and displayed in figures are uncorrected ages. The effects of alpha ejection will still be accounted for during inverse thermal history modelling, discussed in Section 4.6.

AHe data were complemented by published AFT analyses [Philippon *et al.*, 2014; Balestrieri *et al.*, 2016]. However, for the Beto Basin NW margin and Amaro Horst 2nd traverse samples, confined track length measurements were very low or, in many cases non-existent, preventing statistically robust thermal history modelling to be performed. Therefore, new confined track measurements were carried out on these samples after being irradiated by ^{252}Cf for 60 minutes at the University of Melbourne, increasing the number of visible confined tracks and improving MTL statistics.

4.4. Results

Newly acquired confined track data are presented in Table 4.1, and Figures 4.2 and 4.3. AHe results from the Beto Basin and Amaro Horst are presented in Tables 4.2 and 4.3, respectively. All ages are reported to one decimal place in the tables but rounded to whole numbers in the text, except for ages with 1σ errors < 1 Ma. Single grain AHe ages are reported with a conservative 6.2% (1σ) analytical error. Results are discussed by methodology in order of decreasing temperature sensitivities. Within these subsections, data are grouped by regions of shared thermal histories (Beto Basin and Amaro Horst), as determined by Philippon *et al.* [2014] and Balestrieri *et al.* [2016].

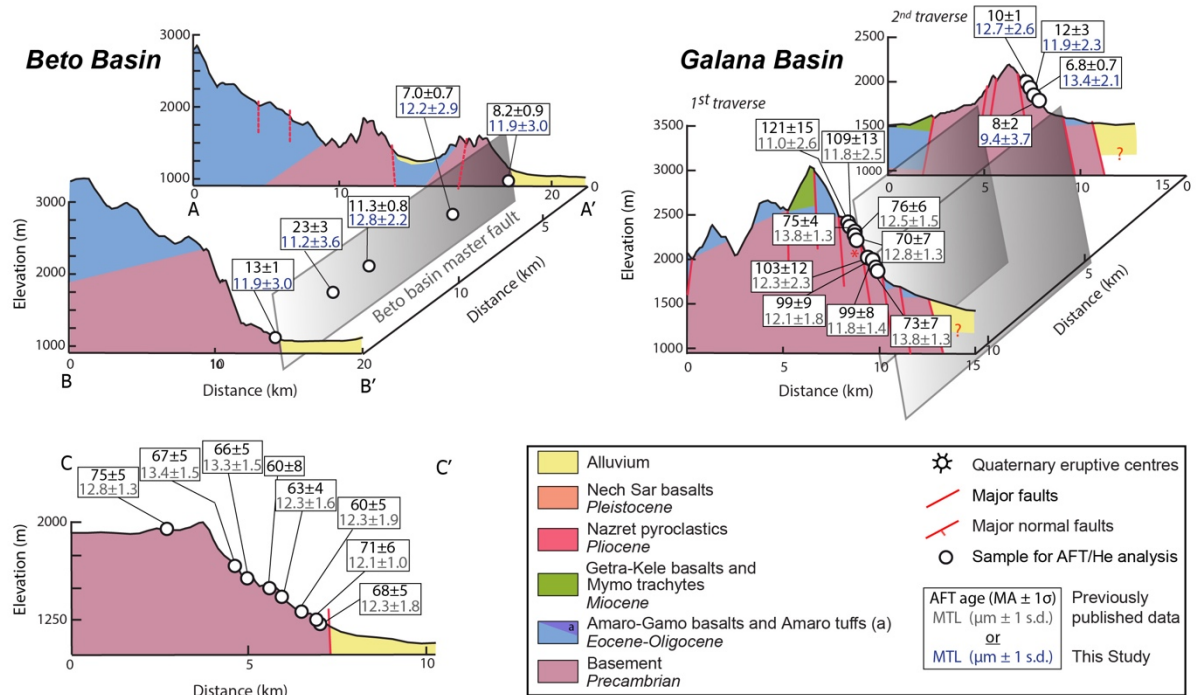


Figure 4.3. Pseudo 3-D geological cross-sections of Beto and Galana Basins, orthogonal to basin-bounding normal faults, illustrating sample positions, and previously published AFT ages [Ma ± 1σ] and mean track lengths (MTL; in grey) [μm ± 1 s.d.] [Philippon et al., 2014; Balestrieri et al., 2016]. Augmented MTL data presented in this study are shown in blue. An ~N-S transect of the southern Beto Basin margins is also shown (location shown in Figure 4.2). Fault planes discussed in text are schematically illustrated in grey. Note, the western fault illustrated in the Amaro Horst 3D cross-section does not extend as far north as the 2nd traverse.

4.4.1 Apatite fission track analyses

4.4.1.1 Beto Basin

Precambrian basement from the southwest margin of the Beto Basin (AFT ages between 76 ± 5 Ma to 59 ± 8 Ma) yield intermediate to long MTL (12.1 to 13.4 μm) and small to moderate standard deviations (0.95 to 1.90 μm) that show no correlation with Dpar (2.3-3.4 μm) (Table 4.1). By contrast, samples from the Beto Basin-bounding fault scarp produced MTL that range between 11.2 and 12.8 μm, with large standard deviations (2.17 to 3.57 μm). Dpar measurements (2.1-2.5 μm) exhibit a positive correlation to MTL and slightly negative correlation to AFT age.

4.4.1.2 Amaro Horst

MTL measurements from the Amaro Horst 1st traverse are intermediate to long (11.0 to 13.8 μm) with intermediate standard deviations (1.26 to 2.63 μm). These data exhibit a negative correlation to elevation (Figure 4.3) that, like AFT age-elevation relationships, are consistent

with higher samples being less thermally rejuvenated with respect to the AFT system. This pattern is slightly disturbed by a “step” in MTL and AFT ages corresponding to where *Balestrieri et al.* [2016] observed a normal fault with significant, but unknown offset. Only AM1-South yields a MTL of 12.9 μm with a standard deviation of 1.30 μm , consistent with other 1st traverse samples.

By contrast, samples from the 2nd traverse of the Amaro Host yield newly obtained MTL that are longer (12.7-13.4 μm) than their 1st traverse counterparts, but similarly moderate standard deviations (2.07-2.57 μm). These data are consistent with their much younger AFT ages (12 ± 3 and 6.8 ± 0.7 Ma), suggesting a greater degree and possibly more recent thermal resetting. Accordingly, the number of confined track length measurements for 2nd traverse samples are low, even after ^{252}Cf irradiation. Despite only five confined track measurements being obtained, sample AM4-2nd*tr* yields a MTL of 11.9 μm with a standard deviation of 2.28 μm that is in agreement with AM2-2nd*tr* and AM5-2nd*tr*. Conversely, AM1-2nd*tr* (6 confined track measurements) yields a markedly shortened MTL (9.4 μm) with very large standard deviation (3.70 μm), which is at odds with other samples in the dataset. Thus, AM1-2nd*tr* is treated as an outlier and omitted from further discussion.

Table 4.1. AFT Summary Table

Sample No.	No. of grains	N_d	ρ_d [10 ⁵ cm ⁻²]	N_s	ρ_s [10 ⁵ cm ⁻²]	N_i	ρ_i [10 ⁵ cm ⁻²]	²³⁸ U [μg/g]	$P(\chi^2)$ [%]	Dpar [μm]	St.Dev. [μm]	Central age [Ma ± 1σ]	N_{length}	Mean track length [μm ± se]	St.Dev. [μm]
Beto Basin															
Northwestern Margin (SW-NE)															
GR12	25	3148	8.03	199	1.59	2250	18.08	27.6	4.5	2.1	0.30	12.6 ± 1.3	66	11.9 ± 0.4	3.02
GR13	28	5612	7.16	86	0.40	503	2.48	4.2	8.8	2.1	0.20	22.8 ± 3.3	54	11.2 ± 0.5	3.57
GR3*	25	5612	7.16	403	2.14	4647	24.70	39.3	21.4	2.5	0.20	11.3 ± 0.8	98	12.8 ± 0.2	2.17
GR2*	25	5612	7.16	111	0.51	2069	9.58	16.1	89.7	2.4	0.20	7.0 ± 0.7	77	12.2 ± 0.3	2.90
GR14	26	3119	7.00	112	1.12	2012	20.08	31.7	72.6	2.2	0.40	8.2 ± 0.9	60	11.9 ± 0.4	2.97
Amaro Horst															
2nd traverse (West-East)															
AM5-2 nd tr	26	5612	7.16	66	0.32	842	4.17	7.6	99.5	2.2	0.40	10.2 ± 1.4	76	12.7 ± 0.3	2.57
AM4-2 nd tr	26	5612	7.16	108	1.82	935	15.85	31.4	< 1	2.1	0.40	12.3 ± 2.7	5	11.9 ± 1.0	2.28
AM2-2 nd tr	23	5612	7.16	109	0.71	2113	13.80	23.8	90.0	2.2	0.60	6.8 ± 0.7	153	13.4 ± 0.2	2.07
AM1-2 nd tr	19	5612	7.16	18	0.19	288	3.07	5.6	52.0	2.1	0.22	8.2 ± 2.0	6	9.4 ± 1.5	3.70

Notes: Confined track data (in bold) were obtained by Samuel C Boone. These analyses were performed on image sets captured by TrackWorks using a 3.2MP AVT Oscar F-320C camera mounted on a Zeiss Axiolmager microscope with a 1000x total magnification and a 100x dry objective (calibration = 0.07/0.07μm/pixel) after etching with 5M HNO₃ for 20 sec at 20°C. Confined track lengths (TINTs) were measured as true 3D lengths using

FastTracks after irradiation by ²⁵²Cf and are corrected for a refractive index of 1.634 for apatite. AFT age data are from Philippon et al. (2014) (samples with asterisk) and Balestrieri et al. (2016).

N_d = number of tracks counted on external detector; ρ_d = standard track density; N_s = number of spontaneous tracks counted; ρ_s = spontaneous track density; N_i = number of induced tracks counted; ρ_i = induced track density;

$P(\chi^2) = (\chi^2)$ probability (Galbraith, 1981); Dpar = mean etch pit diameter parallel to c-axis; N_{length} = number of lengths measured; se = standard error

Table 4.2. Beto Basin Apatite (U-Th-Sm)/He Data Summary

Sample No.	Longitude/Latitude	Elevation [m]	Lab #	⁴ He [ncc]	Mass [mg]	U [ppm]	Th [ppm]	Sm [ppm]	Th/U ratio	[eU] (ppm) ^a	Grain length [μm]	Grain half-width [μm]	R _s ^b [μm]	F _r ^c	Uncorrected Age [Ma ± 1σ]	Corrected Age [Ma ± 1σ]	^d Crystal morphology
Southwestern Margin (SW-NE)																	
GR11	36.69022/5.84000	1949	15382	0.350	0.00772	22.3	31.1	254.5	1.39	29.6	249.3	67.4	79.6	0.79	12.4 ± 0.8	15.7 ± 1.0	1T
			15383	0.959	0.00687	33.9	38.7	193.7	1.14	43.0	210.8	62.6	72.4	0.76	26.5 ± 1.6	34.9 ± 2.2	2T
			15384	0.854	0.01121	33.0	22.4	140.7	0.68	38.3	286.5	62.4	76.8	0.80	16.3 ± 1.0	20.3 ± 1.3	0T
			15805	0.298	0.01084	12.7	22.3	172.6	1.75	17.9	123.1	93.6	79.8	0.80	12.4 ± 0.8	15.5 ± 1.0	0T
			15806	0.399	0.01533	14.7	20.9	188.8	1.42	19.6	263.8	83.6	95.2	0.82	10.7 ± 0.7	13.1 ± 0.8	2T
GR4	36.71787/5.85493	1238	15379	0.124	0.01498	2.0	0.6	81.7	0.28	2.1	362.7	70.5	88.5	0.81	30.7 ± 1.9	37.7 ± 2.3	2T
			15380	0.173	0.01220	8.3	1.8	55.3	0.21	8.7	322.7	67.4	83.7	0.80	13.2 ± 0.8	16.4 ± 1.0	2T
			15381	0.041	0.00867	2.6	0.3	69.7	0.10	2.7	223.0	68.4	78.5	0.80	14.0 ± 0.9	17.7 ± 1.1	2T
			15807	0.229	0.00698	5.4	15.9	104.5	2.95	9.1	204.4	64.1	73.2	0.76	29.1 ± 1.8	38.6 ± 2.4	2T
			15808	0.115	0.01172	3.8	0.2	89.4	0.05	3.8	255.9	67.5	80.1	0.82	20.5 ± 1.3	24.9 ± 1.5	0T
Northwestern Margin (SW-NE)																	
GR12	36.736917/5.90465	1088	15385	0.312	0.01054	21.8	4.0	178.3	0.18	22.7	236.7	73.2	83.8	0.81	10.6 ± 0.7	13.2 ± 0.8	2T
			15386	0.125	0.00695	17.9	1.3	276.3	0.07	18.2	211.5	71.4	80.1	0.81	8.0 ± 0.5	9.8 ± 0.6	1T
			15387	0.027	0.02170	0.6	1.9	12.3	3.48	1.0	282.3	87.4	100.1	0.84	9.9 ± 0.6	11.8 ± 0.7	0T
			15803	0.141	0.01019	19.9	2.5	312.9	0.12	20.5	236.2	82.2	91.5	0.83	5.5 ± 0.3	6.6 ± 0.4	1T
			15804	0.190	0.00789	12.8	6.0	441.0	0.47	14.2	226.3	73.1	82.9	0.81	13.4 ± 0.8	16.6 ± 1.0	1T
GR13	36.77597/5.94070	1104	15388	0.019	0.00989	2.1	2.4	250.2	1.12	2.7	207.2	68.9	77.6	0.80	5.3 ± 0.3	6.7 ± 0.4	0T
			15389	0.228	0.00470	3.1	2.4	589.0	0.77	3.7	176.3	65.2	71.4	0.77	89.7 ± 5.6	115.8 ± 7.2	1T
			15390	0.294	0.01692	2.5	2.9	267.2	1.17	3.2	238.0	84.1	93.2	0.83	40.2 ± 2.5	48.2 ± 3.0	0T
			15809	0.305	0.03015	1.4	0.6	460.4	0.43	1.5	338.2	94.2	110.5	0.87	40.3 ± 2.5	46.6 ± 2.9	0T
			15810	0.079	0.01461	1.2	0.8	399.4	0.70	1.4	240.1	77.8	88.1	0.83	23.3 ± 1.4	28.1 ± 1.7	0T
GR3	36.81173/5.98045	1214	15394	0.369	0.00915	32.0	63.7	130.1	1.99	47.0	257.0	72.7	85.0	0.80	7.0 ± 0.4	8.7 ± 0.5	1T
			15395	0.360	0.01266	24.3	49.0	103.1	2.02	35.8	243.6	71.9	83.3	0.81	6.5 ± 0.4	8.0 ± 0.5	0T
			15396	0.173	0.00990	15.2	37.2	91.6	2.44	23.9	204.2	69.4	77.7	0.79	6.0 ± 0.4	7.5 ± 0.5	0T
GR2	36.88225/6.05857	1115	15397	0.029	0.02486	2.6	2.2	7.3	0.85	3.1	256.8	98.1	106.5	0.86	3.1 ± 0.2	3.6 ± 0.2	0T
			15398	0.036	0.01448	3.6	2.5	5.4	0.70	4.2	332.1	78.9	95.7	0.83	5.0 ± 0.3	6.0 ± 0.4	1T
			15399	0.030	0.02050	3.8	2.6	6.9	0.67	4.4	403.2	84.3	104.6	0.85	2.7 ± 0.2	3.2 ± 0.2	1T
			15801	0.113	0.02543	5.8	3.9	12.2	0.66	6.7	300.6	91.7	105.4	0.86	5.4 ± 0.3	6.3 ± 0.4	0T
			15802	0.099	0.01766	4.1	15.0	11.8	3.69	7.6	256.1	82.8	93.9	0.83	6.1 ± 0.4	7.3 ± 0.5	0T
GR14	36.90665/6.07739	1156	15391	0.184	0.02094	10.1	8.1	33.0	0.80	12.0	190.0	104.7	101.3	0.85	6.0 ± 0.4	7.1 ± 0.4	0T
			15392	0.294	0.01563	26.7	16.4	84.2	0.61	30.6	174.8	94.3	91.9	0.84	5.0 ± 0.3	6.0 ± 0.4	0T
			15393	0.278	0.01075	37.9	25.8	47.6	0.68	44.0	148.3	84.9	81.0	0.81	4.8 ± 0.3	5.9 ± 0.4	0T
			15811	0.136	0.01205	8.5	12.0	37.0	1.41	11.3	199.6	77.5	83.7	0.81	8.2 ± 0.5	10.1 ± 0.6	0T
Durango reference material																	
Durango			15421	6.555	-	-	-	-	21.12	-	-	-	-	1.00	30.2 ± 1.9	30.2 ± 1.9	-
			15434	18.648	-	-	-	-	24.76	-	-	-	-	1.00	30.1 ± 1.9	30.1 ± 1.9	-
			15472	22.720	-	-	-	-	24.38	-	-	-	-	1.00	30.1 ± 1.9	30.1 ± 1.9	-
			15910	31.234	-	-	-	-	19.99	-	-	-	-	1.00	32.3 ± 2.0	32.3 ± 2.0	-
			15952	31.993	-	-	-	-	23.23	-	-	-	-	1.00	29.8 ± 1.8	29.8 ± 1.8	-

Notes: Clear, euhedral, non-fractured apatites were separated manually by hand-picking under an Olympus SZX12 binocular microscope. Apatites, in acid-treated platinum capsules, were outgassed under vacuum at ~900°C for 5 minutes, using a semiconductor Coherent Quattro FAP 820 nm diode laser with fibre-optic coupling to the sample chamber. He volume was determined by isotope dilution using a pure ³He spike, calibrated against an independent ⁴He standard and measured using a Balzers quadrupole (QMS 200 – Prisma) mass spectrometer. A hot blank was run after each gas extraction to verify complete outgassing of the apatite grains. The second re-extract contributed less than 0.5% of the total measured ⁴He for all samples. ²³⁸U, ²³²Th and ¹⁴⁷Sm content was obtained by total dissolution of outgassed apatites, still in platinum capsules, in HNO₃ and analysed using an Agilent 7700x ICP-MS. Durango apatite [McDowell *et al.*, 2005] and Fish Canyon Tuff zircon [Gleadow *et al.*, 2015] were also run as ‘unknowns’ with each batch of apatite samples and served as a check on sample accuracy. (U-Th-Sm)/He ages were calculated and corrected for α-emission following the approach of Farley *et al.*, [1996]. Analytical uncertainties, including the α-ejection correction, an estimated 5 μm uncertainty in grain dimension, gas analysis (<1%) and ICP-MS analytical uncertainties, are conservatively assessed at ~6.2%. Accuracy and precision of U, Th and Sm content range up to 2%, but are typically <1%.

^aeU = effective uranium

^bR_s = equivalent spherical radius

^cF_r is the α-ejection correction after Farley *et al.*, [1996].

^dGrain morphology - 0T = no crystal terminations, 1T = one crystal termination and 2T = 2 crystal terminations.

Table 4.3. Amaro Horst Apatite (U-Th-Sm)/He Data Summary

Sample No.	Longitude/Latitude	Elevation [m]	Lab #	⁴ He [ncc]	Mass [mg]	U [ppm]	Th [ppm]	Sm [ppm]	Th/U ratio	[eU] (ppm) ^a	Grain length [μm]	Grain half-width [μm]	R _e ^b [μm]	F _r ^c	Uncorrected Age [Ma ± 1σ]	Corrected Age [Ma ± 1σ]	^d Crystal morphology
1st traverse (West-East)																	
Above secondary fault																	
AM3-1 st tr	37.87570/5.84692	2400	15409	0.083	0.01306	2.4	3.1	108.0	1.29	3.1	331.8	74.6	91.4	0.82	15.8 ± 1.0	19.3 ± 1.2	1T
			15410	0.358	0.00967	24.0	16.8	166.9	0.70	27.9	232.4	80.7	89.9	0.82	10.8 ± 0.7	13.2 ± 0.8	1T
			15411	1.288	0.02759	7.9	10.6	54.9	1.34	10.4	302.4	95.3	108.7	0.86	36.6 ± 2.3	42.8 ± 2.7	0T
			15748	0.295	0.01723	3.5	8.1	29.6	2.33	5.4	332.2	79.0	95.7	0.81	25.8 ± 1.6	31.7 ± 2.0	2T
			15749	0.091	0.01061	1.0	5.1	23.7	5.27	2.2	300.0	71.1	86.2	0.80	31.7 ± 2.0	39.7 ± 2.5	1T
AM4-1 st tr	37.88127/5.84423	2330	15412	5.726	0.01313	70.8	17.8	141.2	0.25	75.0	265.1	70.2	83.3	0.82	47.5 ± 2.9	57.6 ± 3.6	0T
			15413	1.924	0.01006	35.6	21.1	212.3	0.59	40.6	231.9	65.7	76.8	0.80	38.4 ± 2.4	47.8 ± 3.0	0T
			15414	3.953	0.00795	71.7	22.5	181.2	0.31	77.0	195.5	63.6	72.0	0.80	52.7 ± 3.3	66.3 ± 4.1	0T
AM5-1 st tr	37.87928/5.84167	2200	15415	0.336	0.01417	7.1	0.4	116.9	0.06	7.2	224.9	79.2	87.8	0.84	26.4 ± 1.6	31.5 ± 2.0	0T
AM6-1 st tr	37.88270/5.84025	2080	15418	0.340	0.01116	3.9	2.5	180.4	0.63	4.5	185.5	77.3	81.9	0.82	53.0 ± 3.3	64.9 ± 4.0	0T
			15419	0.174	0.00833	13.2	2.4	267.7	0.18	13.8	240.1	72.3	83.3	0.82	12.2 ± 0.8	15.0 ± 0.9	1T
			15420	2.076	0.01356	9.7	1.6	280.4	0.16	10.1	254.8	91.8	101.2	0.85	119.1 ± 7.4	140.3 ± 8.7	1T
			15799	0.150	0.00889	8.2	3.6	286.5	0.43	9.0	246.2	73.7	85.1	0.81	14.7 ± 0.9	18.1 ± 1.1	1T
			15800	0.029	0.00622	1.2	0.7	196.5	0.62	1.4	159.2	62.4	67.2	0.78	24.4 ± 1.5	31.5 ± 2.0	0T
Below secondary fault																	
AM7-1 st tr	37.88445/5.84010	2000	15422	0.111	0.01530	0.7	0.6	82.3	0.86	0.8	200.4	87.1	91.1	0.83	61.8 ± 3.8	74.3 ± 4.6	0T
			15423	0.025	0.01267	2.8	0.3	103.8	0.11	2.9	200.3	79.3	85.2	0.83	5.5 ± 0.3	6.6 ± 0.4	0T
AM8-1 st tr	37.88587/5.83923	1950	15425	0.469	0.01699	12.1	1.5	113.5	0.12	12.5	358.6	68.6	86.4	0.83	18.0 ± 1.1	21.6 ± 1.3	0T
			15426	2.280	0.01219	10.9	7.8	85.3	0.71	12.7	271.7	66.8	80.4	0.81	118.7 ± 7.4	146.3 ± 9.1	0T
			15427	0.377	0.00966	10.7	6.0	119.5	0.56	12.1	224.7	65.4	76.0	0.80	26.0 ± 1.6	32.5 ± 2.0	0T
			15797	0.178	0.00430	5.4	4.9	99.2	0.89	6.6	202.4	50.5	60.7	0.72	50.8 ± 3.1	70.9 ± 4.4	2T
			15798	0.125	0.00937	13.8	4.6	172.7	0.33	14.9	239.1	77.5	87.8	0.82	7.3 ± 0.5	8.8 ± 0.5	1T
AM9-1 st tr	37.88597/5.83885	1920	15428	0.036	0.00983	4.0	0.7	49.3	0.18	4.2	206.0	75.7	83.1	0.81	7.2 ± 0.4	9.0 ± 0.6	2T
			15429	0.093	0.00800	2.9	1.5	97.9	0.50	3.3	165.5	69.4	73.3	0.80	28.1 ± 1.7	35.4 ± 2.2	0T
			15795	0.073	0.01420	6.3	1.3	153.7	0.20	6.6	189.7	86.3	89.0	0.84	6.2 ± 0.4	7.4 ± 0.5	0T
			15796	0.366	0.02621	3.3	1.0	94.9	0.31	3.5	217.0	109.6	109.2	0.87	31.3 ± 1.9	36.2 ± 2.2	0T
AM10-1 st tr	37.88743/5.83827	1850	15431	0.028	0.01732	1.4	2.1	18.0	1.52	1.9	277.3	99.4	109.7	0.85	6.9 ± 0.4	8.1 ± 0.5	1T
			15432	1.241	0.01258	19.4	2.0	186.8	0.10	19.9	241.4	72.0	83.2	0.83	40.3 ± 2.5	48.7 ± 3.0	0T
			15433	0.780	0.01317	16.6	1.5	181.7	0.09	17.0	217.2	77.7	85.8	0.83	28.2 ± 1.8	33.9 ± 2.1	0T
			15793	0.910	0.00753	13.4	2.2	180.6	0.16	13.9	263.2	58.7	72.0	0.77	70.0 ± 4.3	90.5 ± 5.6	2T
			15794	2.472	0.01165	51.5	10.0	248.7	0.19	53.9	251.9	67.8	80.1	0.82	32.1 ± 2.0	39.2 ± 2.4	0T
Southern Amaro Horst (Segen Basin)																	
AM1-South	37.72333/5.37972	1000	15401	0.718	0.00978	8.5	0.8	160.7	0.09	8.7	254.4	76.0	87.8	0.83	67.6 ± 4.2	81.8 ± 5.1	1T
			15402	0.034	0.00890	5.1	2.1	91.8	0.41	5.6	189.5	68.3	75.3	0.80	5.4 ± 0.3	6.8 ± 0.4	0T
			15746	1.133	0.00688	19.7	10.2	81.1	0.52	22.1	234.8	65.8	77.1	0.79	60.6 ± 3.8	76.4 ± 4.7	1T
			15747	1.007	0.00775	16.3	10.5	68.4	0.65	18.8	205.5	61.2	70.8	0.79	56.4 ± 3.5	71.8 ± 4.5	0T
2nd traverse (West-East)																	
AM5-2 nd tr	37.88800/5.94877	2004	15438	0.061	0.01341	10.2	1.5	104.8	0.15	10.6	221.7	77.6	86.2	0.83	3.5 ± 0.2	4.2 ± 0.3	0T
			15439	1.127	0.01481	27.9	3.7	265.4	0.13	28.8	239.8	78.4	88.6	0.84	21.5 ± 1.3	25.7 ± 1.6	0T
			15739	0.099	0.00811	16.6	7.3	135.5	0.44	18.3	210.6	78.5	85.8	0.82	5.4 ± 0.3	6.7 ± 0.4	1T
AM2-2 nd tr	37.89165/5.94570	1756	15435	0.040	0.00759	11.8	12.6	347.0	1.06	14.8	223.2	58.2	69.2	0.78	2.8 ± 0.2	3.6 ± 0.2	0T
			15436	0.091	0.00710	35.2	4.3	412.4	0.12	36.2	225.0	61.6	72.5	0.78	2.9 ± 0.2	3.7 ± 0.2	2T
			15437	0.052	0.00468	19.6	17.1	374.3	0.87	23.6	201.1	59.0	68.4	0.76	3.8 ± 0.2	5.0 ± 0.3	1T
			15741	0.142	0.00853	23.5	8.9	332.7	0.38	25.6	247.8	64.4	76.6	0.78	5.2 ± 0.3	6.7 ± 0.4	2T
			15742	0.115	0.01553	13.0	8.3	316.1	0.64	15.0	253.6	85.8	96.2	0.82	3.9 ± 0.2	4.8 ± 0.3	2T
AM1-2 nd tr	37.89472/5.94618	1685	15442	0.047	0.01896	3.4	0.6	363.4	0.19	3.5	301.8	98.3	111.2	0.86	5.0 ± 0.3	5.8 ± 0.4	1T
			15443	0.107	0.00871	2.0	4.5	136.2	2.26	3.1	249.0	72.2	84.0	0.80	31.1 ± 1.9	38.9 ± 2.4	1T
			15743	0.064	0.01637	4.2	6.6	85.0	1.59	5.8	264.6	78.4	90.8	0.83	5.5 ± 0.3	6.7 ± 0.4	0T
			15744	0.029	0.01646	2.7	2.2	168.8	0.79	3.2	326.1	77.9	94.3	0.82	4.3 ± 0.3	5.2 ± 0.3	2T
Durango reference material																	
Durango			15421	6.555	-	-	-	-	21.12	-	-	-	-	1.00	30.2 ± 1.9	30.2 ± 1.9	-
			15434	18.648	-	-	-	-	24.76	-	-	-	-	1.00	30.1 ± 1.9	30.1 ± 1.9	-
			15472	22.720	-	-	-	-	24.38	-	-	-	-	1.00	30.1 ± 1.9	30.1 ± 1.9	-
			15910	31.234	-	-	-	-	19.99	-	-	-	-	1.00	32.3 ± 2.0	32.3 ± 2.0	-
			15952	31.993	-	-	-	-	23.23	-	-	-	-	1.00	29.8 ± 1.8	29.8 ± 1.8	-

Notes: Analytical protocol as for Table 4.2.

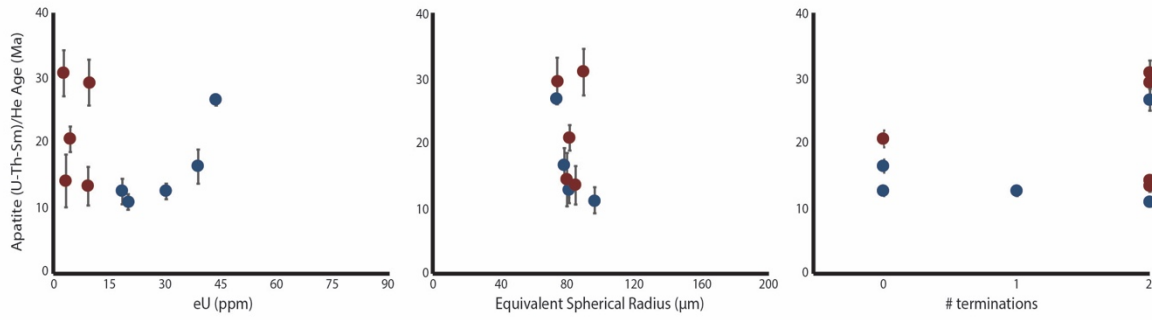
^a eU = effective uranium

^b R_e = equivalent spherical radius

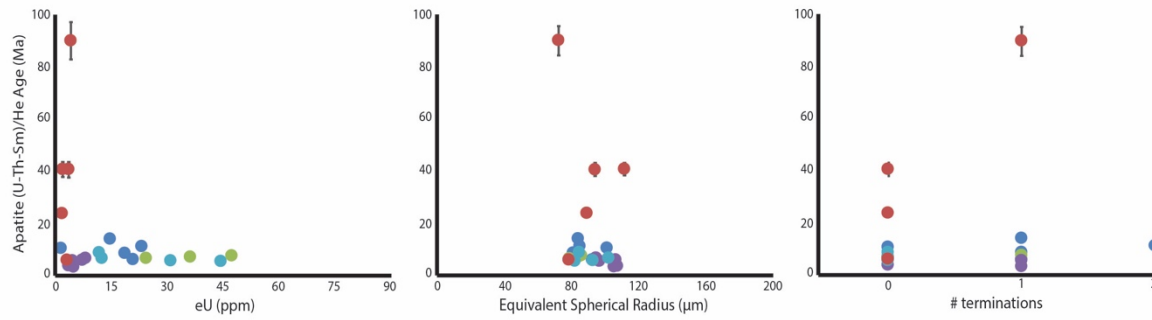
^c F_r is the α-ejection correction after Farley *et al.*, [1996].

^d Grain morphology - 0T = no crystal terminations, 1T = one crystal termination and 2T = 2 crystal terminations.

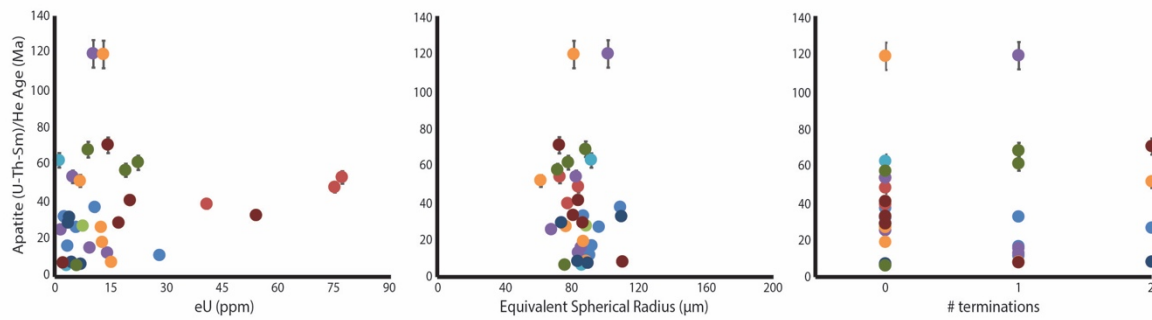
Beto Basin SW margin ● GR11 ● GR4



Beto Basin NW margin ● GR12 ● GR13 ● GR3 ● GR2 ● GR14



Amaro Horst 1st traverse ● AM3 ● AM4 ● AM5 ● AM6 ● AM7 ● AM8 ● AM9 ● AM10 ● AMSouth



Amaro Horst 2nd traverse ● AM5 ● AM2 ● AM1

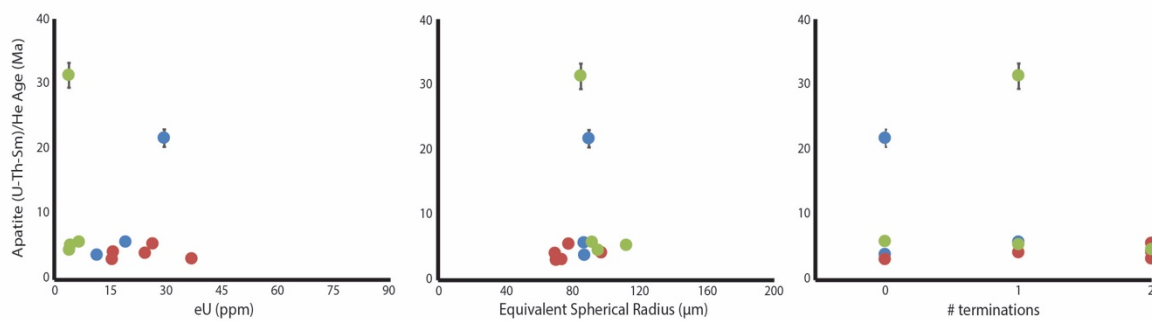


Figure 4.4. Plots of AHe age versus eU content (left column), AHe age versus equivalent spherical radius (middle column) and AHe age versus the number of apatite crystal terminations (right column) for samples from the Beto Basin SW margin (first row) and NW margin (second row) and Amaro Horst 1st traverse (third row) and 2nd traverse (fourth row). In general, AHe ages do not exhibit strong correlations with grain size, morphology or eU content. The few possible exceptions are discussed in Section 4.4.2.

4.4.2 Apatite (U-Th-Sm)/He analyses

AHe analyses from both the Beto Basin and Amaro Horst regions yield large inter-sample dispersion. A wide range of crystallographic and compositional characteristics have been shown to affect He diffusion in apatite, at times producing intra-sample He age dispersion greater than 50% [Wolfe and Stockli, 2010; Wildman *et al.*, 2016]. Of particular concern to this study, are grain size, morphology and accumulated radiation damage, all of which are considered during data interpretation and thermal history modelling. The potential effect grain size may have on resultant He age is assessed by proxy, equivalent spherical radius (R_s ; $R_s = [3 \cdot R \cdot L] / [2 \cdot \{R + L\}]$), allowing the comparison of 0, 1 and 2 termination grains. Similarly, radiation damage modified He diffusion effects are assessed by using effective uranium concentration (eU; $eU = U + 0.235 \times Th$) as a proxy for α -radiation damage, which considers the decay rate and abundance of radiogenic parent elements.

4.4.2.1 Beto Basin

AHe ages from the southwestern margin of the Beto Basin range from 31 ± 2 Ma to 10.7 ± 0.7 Ma and show no correlation to grain size or morphology. Of these, six 2T grains yielded ages between 39 ± 2 Ma and 13.1 ± 0.8 Ma. Sample GR11 (27 ± 2 to 10.7 ± 0.7 Ma; 5 grains) however, shows a strong positive correlation to eU content (Figure 4.4).

Apatite He analyses from the Beto Basin-bounding fault scarp yielded ages between 2.7 ± 0.2 Ma and 90 ± 6 Ma that exhibit a general trend of decreasing age towards the northeast (Table 4.2). Only one 2T grain was obtained (sample GR12) and yields an AHe age of 10.6 ± 0.7 Ma. Apart from GR13 (5 grains; 90 ± 6 to 5.3 ± 0.3 Ma), rocks from the northwest margin exhibit far less intra-sample AHe age dispersion (13.4 ± 0.8 to 2.7 ± 0.2 Ma). Apatites from GR13 appear to show a strong correlation between age and morphology, with its only 1T grain (90 ± 6 Ma) yielding a single grain age more than double the next oldest. The remaining grains (all 0T; 40 ± 3 to 5.3 ± 0.3 Ma) show a positive correlation with grain size.

4.4.2.1 Amaro Horst

Single grain AHe results from the Amaro Horst 1st traverse range from 5.5 ± 0.3 Ma to 119 ± 7 Ma. Upper 1st traverse samples show a general decrease in age and dispersion with elevation (Table 4.3), contrary to the AFT age-elevation trend for these samples (Table 4.1) and opposite of what might be expected for vertical transects [e.g. Fitzgerald *et al.*, 2006]. Within these samples, intra-sample single grain AHe age variation does not display any clear

relationships to morphology, grain size or eU concentration (Figure 4.4). The exceptions being AM6-*1sttr* (12.2 ± 0.8 to 119 ± 7 Ma), showing a positive age-grain size correlation, and AM4-*1sttr* (38 ± 2 to 53 ± 3 Ma), whose AHe ages slightly increase with eU content. The sole 2T grain analysed from upper 1st traverse samples (AM3-*1sttr*) records an AHe age of 26 ± 2 Ma. Lower 1st traverse samples, on the other hand, do not show a clear age-elevation relationship (Table 4.3). Single grain ages are more dispersed than their 1st traverse counterparts, ranging from 146 ± 9 Ma to 5.5 ± 0.3 Ma. Dispersion in the three 2T grains analysed from below the fault is reduced but substantial (70 ± 4 to 7.2 ± 0.4 Ma). Intra-sample age variation relationships are complex and ambiguous and the few instances where samples do exhibit a strong relationship with a certain parameter, it is often the opposite of expected trends (Figure 4). For example, sample AM8-*1sttr* shows negative correlations between AHe age and both grainsize and eU content. AM1-*South* yields predominantly older single grain ages (3 grains; 68 ± 4 to 56 ± 4 Ma), apart from one very young 0T grain (5.4 ± 0.3 Ma).

Samples from the Amaro Horst 2nd traverse yield younger, far less dispersed AHe ages between 31 ± 2 Ma to 2.8 ± 0.2 Ma that show no relationship to elevation (Table 4.3). 2T AHe ages (4 grains) exhibit very little dispersion and range from 5.2 ± 0.3 Ma to 2.9 ± 0.2 Ma. Sample AM5-*2ndtr* (3 grains; 22 ± 1 to 3.5 ± 0.2 Ma) shows a strong correlation with eU content (Figure 4.5). Apatites from sample AM1-*2ndtr* (4 grains; 31 ± 2 to 4.3 ± 0.3 Ma) exhibit negative age-eU and age- R_s trends.

4.4.3 Apatite (U-Th-Sm)/He age dispersion

Single grain AHe ages obtained in this study exhibit complex relationships with accumulated radiation damage, grain size and morphology (Figure 4.4). Intra-sample single grain AHe age variation relationships to eU content and R_s often differ from the positive correlations that may be expected for relatively slowly cooled rocks [Reiners and Farley, 2001; Flowers *et al.*, 2009]. These are likely the result of the competing effects of grain size and radiation damage decoupling the expected correlation [Brown *et al.*, 2013; Wildman *et al.*, 2016], perhaps compounded by other unquantified factors such as U, Th, and Sm zonation [Flowers and Kelley, 2011] and apatite chemistry [Gautheron *et al.*, 2013; Mbongo Djimbi *et al.*, 2015]. The analysis of fragmented apatite grains likely further convolutes these relationships by producing ages that differ from their whole grain equivalents [Wildman *et al.*, 2016]. One

termination (1T) fragments can produce older or younger ages than the whole grain age, while no termination (0T) grains are most likely to yield older ages [Brown *et al.*, 2013].

Similar complex age dispersion patterns have been observed in previous AHe studies from slowly cooled terranes [e.g. Flowers *et al.*, 2007; Kohn *et al.*, 2009]. Despite these complications, general relationships between the degree of single grain AHe age dispersion and the thermal history of a sample or group of samples have been identified. For instance, the effects of radiation damage and grain size on He diffusion has been shown to become more pronounced during thermal histories characterized by slow cooling, prolonged residence in the AHe PRZ or protracted reheating [Fitzgerald *et al.*, 2006; Brown *et al.*, 2013]. Whereas, datasets with low inter-aliquot age dispersion despite variation in eU content and grain size are indicative of rapid cooling or complete resetting during reheating [Flowers and Kelley, 2011]. By extension, the degree of dispersion observed for samples in this study can be used to infer information about the thermal evolution of the Beto Basin and Amaro Horst.

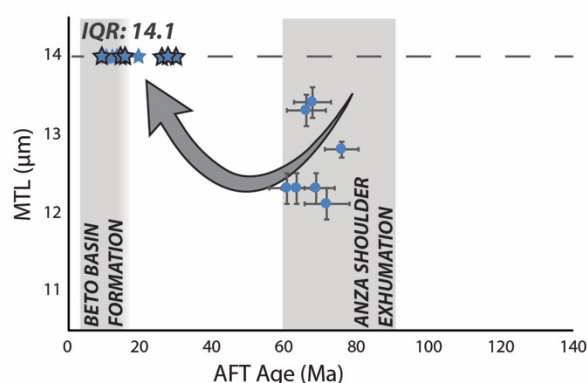
Brown *et al.* [2013] calculated intra-sample dispersion percentages as the range divided by the mean for a set of single grain AHe ages obtained from one sample. The calculated dispersion [%] was then used as a simple way to apply a value to the degree of intra-sample age variance experienced by a sample. However, for comparison of age dispersion between different groups of samples, with each group consisting of samples that share a similar thermal history, the statistical measure of dispersion suggested by Brown *et al.* [2013] is inappropriate because datasets with lower means will, by definition, have higher percentages of dispersion. In other words, AHe data from a basement block that experienced rapid uplift and cooling in the Neogene would be significantly more dispersed than AHe results from rocks that experienced the exact same duration and rate of uplift and cooling, but in the Cretaceous. Instead, we use the interquartile range (IQR):

$$IQR = Q_3 - Q_1$$

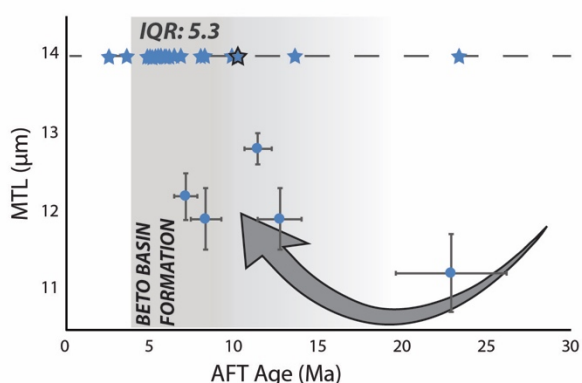
where Q_1 and Q_3 are the first and third quartiles, respectively. The IQR is a computationally simplistic robust scale estimator and, thus, resistant to outliers. Moreover, the IQR statistic does not assume a symmetric data distribution [Rousseeuw and Croux, 1993] and is independent of a dataset's mean. Assuming that the degree of spread in single grain AHe

ages is characteristic of their shared thermal history [*Fitzgerald et al.*, 2006; *Flowers and Kelley*, 2011; *Brown et al.*, 2013], the use of IQR allows for a first order comparison of the type of cooling history experienced by different samples, or groups of samples, independent of timing. For example, a group of samples that experienced rapid cooling through the AHe PRZ or complete resetting during reheating should record low inter-aliquot age dispersion [*Flowers and Kelley*, 2011] and, thus, a low IQR. Whereas, a group of rocks that experienced slow cooling, prolonged residence in the AHe PRZ or protracted reheating should record broad single grain age variance [*Fitzgerald et al.*, 2006; *Brown et al.*, 2013] and a large IQR.

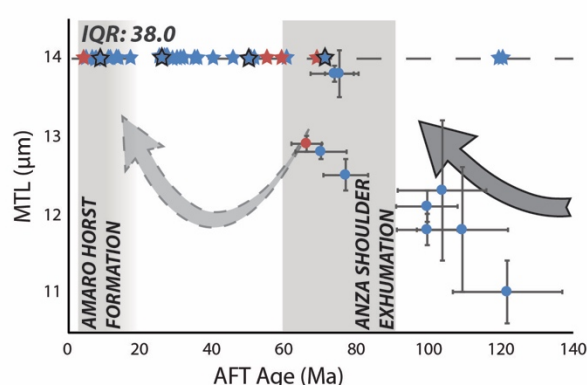
Beto Basin SW margin



Beto Basin NW margin



Amaro Horst 1st traverse (● AM1-South)



Amaro Horst 2nd traverse

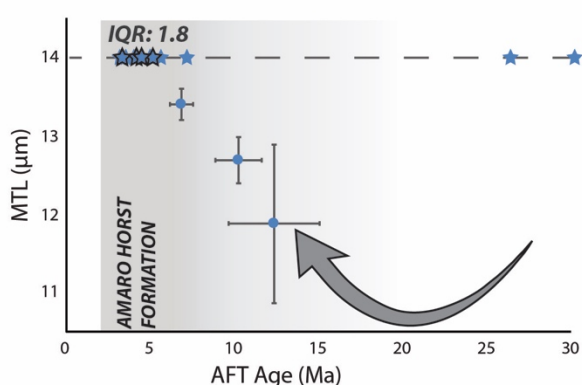


Figure 4.5. Boomerang plots of AFT age versus MTL (circles) for Beto Basin and Amaro Horst basement rock samples. Single grain AHe ages are displayed as stars, with 2T grains outlined in black. Note, three Beto Basin NW margin AHe single grain ages from sample GR13 (40.2 ± 2.5 Ma, 40.3 ± 2.5 Ma, and 89.7 ± 5.6 Ma) are not shown. Approximate duration of tectonothermal events recorded by AFT and AHe data are marked by grey bars. Dark grey arrows represent inferred boomerang curves. Beto Basin SW margins AFT data form the descending arm of a boomerang curve, suggesting these samples experienced an early period of Cretaceous cooling followed by residence or reheating into the AFT PAZ. A subsequent period of Neogene cooling and exhumation is required to have exposed samples with shortened MTL at the surface, consistent with seven Neogene AHe single grain ages. Similarly, Beto Basin NW margin and Amaro Horst 2nd traverse samples also record a period of earlier cooling, possibly also during the Cretaceous, overprinted by subsequent residence or reheating into the AFT PAZ and followed by pronounced Neogene cooling. However, these samples cooled from higher temperatures within the upper AFT PAZ, resulting in younger Neogene AFT ages that form ascending arms of their respective boomerang curves which approach predominantly middle Miocene-Pliocene single grain AHe ages. By contrast, Amaro Horst 1st traverse samples produced older Cretaceous AFT ages. These data form the ascending arm of a boomerang curve, indicative of a pronounced period of Cretaceous cooling. However, the lack of Cenozoic 1st traverse AFT ages suggests that during subsequent Cenozoic reheating (light grey arrow), indicated by 30 Paleocene-Miocene single grain AHe ages, these samples did not reach high enough temperatures for long enough to substantially reset their AFT ages. See Section 4.5 for description of a boomerang plot and data interpretation.

4.5. Discussion

AFT and AHe analyses from the Beto Basin and Amaro Horst yield important constraints on the thermal evolution of the BRZ of southern Ethiopia. The relationship between these two data sets is illustrated in Figure 4.5, where AFT ages are plotted against MTL data for groups of samples that have shared very similar thermal histories as determined in previous low-temperature thermochronology investigations [Philippon *et al.*, 2014; Balestrieri *et al.*, 2016]. These so-called boomerang plots usefully illustrate multi-stage thermal histories for sets of samples that have cooled from different paleotemperatures during a later period of cooling [Green *et al.*, 1989; Gallagher and Brown, 1997]. In these plots, samples that record an earlier phase of cooling but were negligibly affected by subsequent reheating and cooling will yield older ages with relatively long MTL ($> 14 \mu\text{m}$), as they had already been exposed to near-surface temperatures prior to renewed cooling. By contrast, samples that cooled from paleotemperatures greater than $\sim 110\text{--}120^\circ\text{C}$ during the final cooling stage will then record relatively long MTL ($> 14 \mu\text{m}$) and young AFT ages that approximate the time of a later cooling phase. Samples that cooled from temperatures within the PAZ ($\sim 110\text{--}60^\circ\text{C}$) before the final cooling episode will yield intermediate “mixed” AFT ages with shortened MTL.

Samples from the southwest margin of the Beto Basin form the descending portion of a boomerang curve (Figure 4.4), characteristic of an early period of pronounced cooling, such as Late Cretaceous-earliest Paleogene denudational cooling of the Anza Rift shoulders recorded regionally [e.g. Foster and Gleadow, 1992, 1993, 1996; Torres Acosta *et al.*, 2015], followed by reheating into and/or prolonged residence in the PAZ. Subsequent cooling and exhumation, is then required to expose samples with substantially annealed MTL to the surface. This is consistent with Oligocene-Miocene single grain AHe analyses from these samples that are moderately skewed towards younger ages and have an intermediate IQR of 14.1. By contrast, samples collected from the footwall of the Beto Basin-bounding master fault form the ascending portion of a boomerang curve, suggesting these samples cooled from higher paleotemperatures in the upper PAZ during this final cooling phase. This is consistent with the limited spread in AHe data and corresponding low IQR (5.3). By projecting the boomerang curve to longer MTL we can estimate that this later cooling phase likely occurred during Miocene-Pliocene time, consistent with 18 of 22 single grain AHe analyses (13.4 ± 0.8 to 2.7 ± 0.2 Ma) from these same rocks.

AFT analyses of Amaro Horst 1st traverse samples record little late Cenozoic cooling, as indicated by the corresponding boomerang plot and large IQR (38.0) of the corresponding AHe data. Instead, these data form the ascending portion of a boomerang trend that curves towards the relatively long MTL samples AM4-*Isttr* (75 ± 4 Ma; $13.8 \mu\text{m}$) and AM10-*Isttr* (73 ± 7 Ma; $13.8 \mu\text{m}$). The slightly younger ages and shorter MTL of samples AM6-*Isttr* (70 ± 7 Ma; $12.8 \mu\text{m}$) and AM1-*South* (66 ± 4 Ma; $12.9 \mu\text{m}$) suggest that these samples may form the very beginning of a younger boomerang curve which extends towards a subsequent period of late Cenozoic cooling. This is in strong accordance with Amaro Horst 1st traverse AHe analyses, 32 of 34 of which yield latest Cretaceous-Miocene ages (70 ± 4 to 5.5 ± 0.3 Ma).

Late Cenozoic cooling hinted at by the spread of Amaro Horst 1st traverse samples is more clearly evident in the age-MTL plot of 2nd traverse samples. Here, AFT data form a distinct ascending portion of a boomerang curve that extends towards long MTL ($>14 \mu\text{m}$) between 15 and 5 Ma. The substantially younger AFT ages (12 ± 3 to 6.8 ± 0.7 Ma) and relatively longer MTL (11.9 - $13.4 \mu\text{m}$) suggest that these samples cooled from elevated temperatures within the PAZ during later Miocene-Pliocene cooling. This is supported by the 2nd traverse AHe data, showing very low age dispersion (12 grains; IQR = 1.8) that skew towards latest Miocene-Pliocene ages, with ten single grains yielding ages between 6.7 ± 0.4 Ma to 3.6 ± 0.2 Ma.

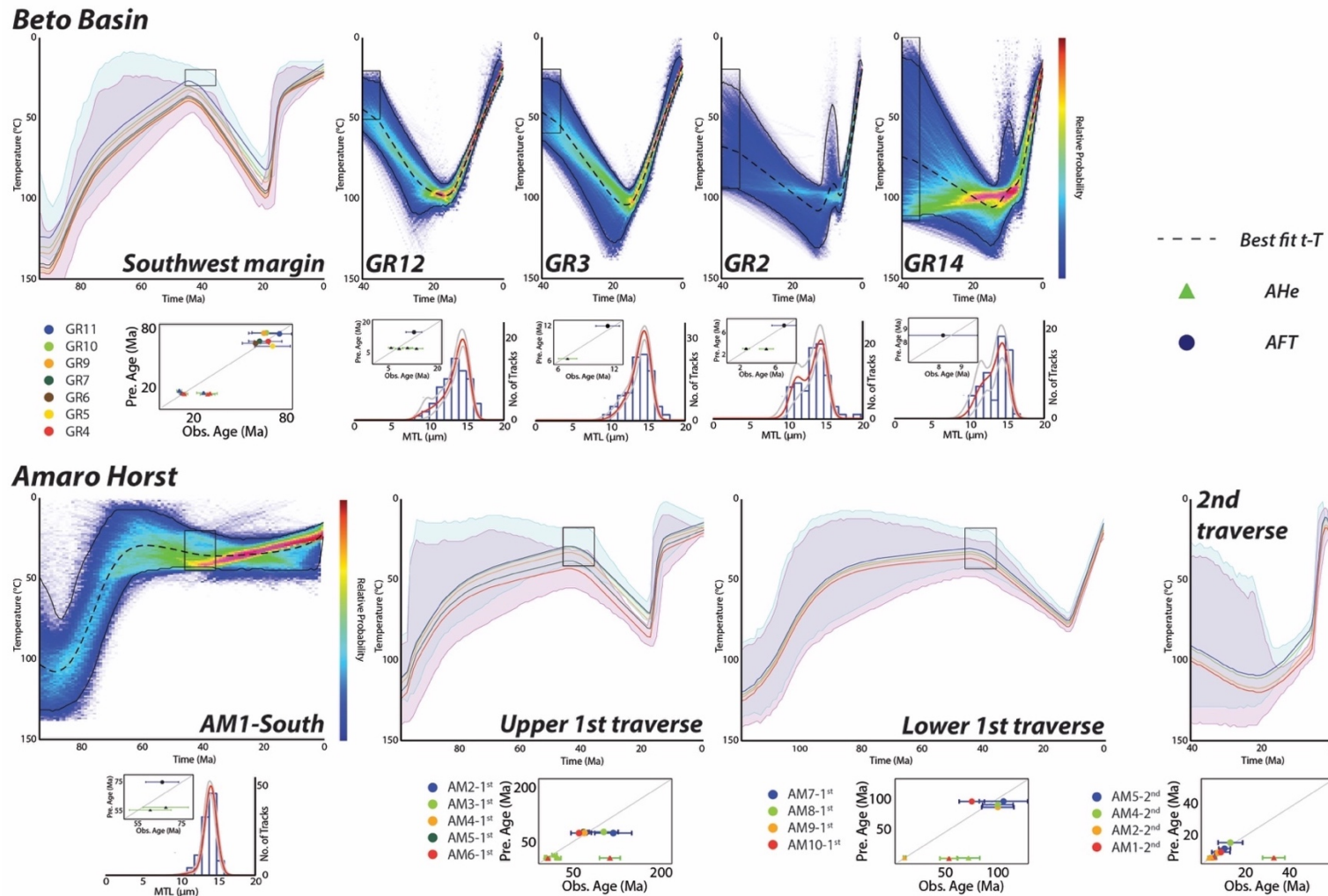


Figure 4.6. Thermal history models of Beto Basin and Amaro Horst samples generated using the QTQt software following the protocol presented in Section 6. Samples were grouped together during thermal history modelling as vertical transects, where possible (i.e. Beto Basin SW margin samples and Amaro Horst upper 1st traverse, lower 1st traverse and 2nd traverse samples). A plot of observed versus predicted AFT and AHe ages accompanies each thermal history model. AHe ages are displayed with resampling errors (see text). Single sample thermal history models are also accompanied by a plot of observed confined track distribution overlain by the predicted distribution (red curve) and the 95% confidence interval (grey curves) of the predicted distribution. Plots of observed and predicted confined track distributions for vertical transect thermal history models are presented in Supplemental Data C. The posterior probability distribution of accepted time-temperature paths is represented by the colour map, with warmer colours indicating higher probabilities. Black hashed lines represent the best fit time-temperature path, while solid black lines represent the 95% confidence interval limits.

4.6. Thermal History Modelling

In order to quantify thermal histories for the Beto Basin and Amaro Horst of the BRZ, AFT and AHe data were inversely modelled using *QTQt* (Figure 4.6) [Gallagher, 2012], with the multi-compositional fission track annealing model of *Ketcham et al.* [2007a]. Confined fission track lengths were normalised for annealing and etching anisotropy using the c-axis projection method [Ketcham et al., 2007b]. Differences in AHe grain-size were accounted for using the alpha ejection correction (F_T) [Farley et al., 1996]. Radiation damage accumulation and annealing models were employed for modelling of AHe data [Gautheron et al., 2009] to account for differences in radiation-modified diffusion kinetics recorded by age-eU relationships. However, these models are unable to account for all known sources of intrasample ZHe and AHe age dispersion, such as chemical zonation [Guenther et al., 2013] and apatite composition [Gautheron et al., 2013]. Therefore, during modelling *QTQt* was allowed to resample the He error for single grain ages, effectively decreasing their influence on determining thermal histories, as compared to data from the AFT system, whose thermal annealing kinetics are better understood [Gallagher, 2012]. For samples with fragmented AHe grains, only 1- and 2-termination grains were modelled in *QTQt* due to the current implementation of the *Brown et al.* [2013] fragmentation model's inability to process 0-termination grains.

Samples were initially modelled individually (e.g. AM1-*South*), unless they had a relatively low number of confined track measurements (< 60). AM1-*West* was not modelled at all, as no new data were acquired for this sample in this study. A constraint was applied to all models, requiring samples to currently be at surface temperatures (20 ± 10 °C). Samples then underwent thermal history modelling again with the addition of a second time-temperature constraint corresponding to the emplacement of the Gamo Amaro basalts (45-35 Ma) [Ebinger et al., 1993] overlying Precambrian basement in the Beto Basin and Amaro Horst regions, requiring samples to be at near-surface temperatures during that time. The temperature range for this constraint varied between samples and was determined using the difference in elevation between a sample and the base of the overlying Gamo Amaro basalt and a conservative paleogeothermal gradient of 30 °C /km, consistent with the average Proterozoic mobile belts of East Africa [Nyblade, 1990]. This constraint was omitted from samples from the Amaro Horst 2nd traverse, as any volcanic rocks that may have once

overlain those rocks have since been eroded away. Thermal history models for individual samples not included in Figure 4.6 are presented in Supplementary Data A (without the Eocene t-T constraint) and Supplementary Data B (with the Eocene t-T constraint). Where possible, samples were then modelled together as vertical profile groups to better constrain the thermal evolution of their respective basement blocks. Samples with low confined track measurement (< 60) were included in vertical transect models, as they were given little weight in determining probable time-temperature paths compared to more statistically robust data in their group. Vertical transect thermal history modelling was performed for samples from the Beto Basin southwest margin (GR4-11), upper Amaro Horst 1st traverse (AM2-1sttr- AM6-1sttr), lower Amaro Horst 1st traverse (AM7-1sttr- AM10-1sttr) and the Amaro Horst 2nd traverse (AM1-2ndtr- AM5-2ndtr).

4.6.1 Cretaceous-early Paleogene Anza Rifting and Eocene-early Miocene magmatism

Thermal history models from both the Beto Basin southwest margin and the Amaro Horst 1st traverse record an initial period of pronounced Cretaceous-early Paleogene cooling beginning between ~120-90 Ma and reaching near-surface temperatures by the late Eocene (45-35 Ma). Thermal history models for samples GR3, GR4, GR12, GR14 and the Amaro Horst 2nd traverse are not constrained prior to Eocene time due to the total, or near-total, resetting of AFT and AHe data during subsequent late Paleogene reheating. However, their proximity to samples whose thermal history models extend back to Cretaceous times and the lack of intermediate pre-EARS structures allows for the assumption that these also experienced protracted Cretaceous-late Eocene cooling. Thermal history models from the Beto Basin southwest margin and Amaro Horst 1st traverse suggest that Late Cretaceous-late Eocene cooling consisted of an initial period of rapid cooling (2.5-5 °C/Ma) before transitioning to protracted cooling phase (0.33-1.67 °C/Ma) sometime between 90-70 Ma.

The presence of a top basement erosional unconformity and the large decrease in paleotemperature (~80-90 °C) as indicated by t-T reconstructions are consistent with erosional denudation as the mechanism for observed cooling. A similar period of denudational cooling, both in timing and character, has been observed in low-temperature thermochronology studies performed throughout much of East Africa [e.g. Chapter 3; *Foster and Gleadow*, 1992, 1993, 1996; *Noble et al.*, 1997; *Torres Acosta et al.*, 2015; *Boone et al.*, 2018]. This pervasive period of cooling has been interpreted as recording rift margin

denudation during the development of the NW-SE trending, Cretaceous-early Paleogene Anza Rift [e.g. *Foster and Gleadow*, 1992].

Between 45 Ma and 35 Ma, the BRZ was uniformly covered by voluminous [2000-2300 km³; *Ebinger et al.*, 1993] Amaro Gamo basaltic flows, marking the earliest manifestation of EARS-related volcanism in East Africa. This is recorded in all thermal history models presented in this study by a transition to a period of marked reheating ca. 45 Ma. Time-temperature reconstructions suggest that basement samples from the Beto Basin and the Amaro Horst 1st and 2nd traverses experienced an increase in temperature of between ~45-60 °C by the early Miocene. If burial was the sole heating mechanism, these estimates indicate the Gamo Amaro basalt flows in these areas reach thicknesses between ~1.5-2 km, assuming a conservative paleogeothermal gradient of 30 °C/km. By comparison, the thermal history model for sample AM1-South records very little, if any (up to 10 °C), late Paleogene reheating. This is consistent with its retention of Cretaceous AFT and AHe ages and suggests that the thickness of Gamo Amaro basalt flows decreased substantially towards the south, likely reaching no more than 300 m in the Segen Basin area.

Thermal history models from the Beto Basin consistently indicate basement rocks reached maximum paleotemperatures ca. 18 Ma. By contrast, the timing of maximum Neogene paleotemperatures indicated by thermal history reconstructions for Amaro Horst samples is more variable, ranging between ~18 Ma (upper 1st traverse) and ~12 Ma (lower 1st traverse). If this apparent lag does in fact reflect continued reheating of lower 1st traverse samples into the middle Miocene, this may indicate further burial beneath the early-middle Miocene Getra-Kele basalts [18-11 Ma, *Woldegabriel et al.*, 1991; *Ebinger et al.*, 1993], absent in the Gofa Province further west.

4.6.2 Miocene-Recent tectono-thermal evolution of the Beto Basin

Thermal history models from samples in the footwall of the southwest Beto Basin margin-defining fault and the southernmost samples (GR12 and GR3) from the basin's northwest shoulder indicate that these rocks began to experience rapid cooling (~6-10 °C/Ma) in the early Miocene (20-18 Ma). Continued, albeit less extensive, volcanic activity in southern Ethiopia during the late Miocene-Pleistocene [*Zanettin et al.*, 1978; *Corti*, 2009] and the present-day high heat flow of the nearby Turkana crust [*Morley et al.*, 1999b] argue against a Neogene-Quaternary decrease in geothermal gradient as causing the observed cooling.

Instead, the highly eroded nature of the Gamo Amaro basalts capping the Beto Basin margins is consistent with rift-shoulder denudation as the primary cooling mechanism.

Circa 14 Ma, cooling rates of the southwest Beto margin appear to have substantially slowed (~ 0.7 °C/Ma), perhaps indicating a marked decrease or cessation of any vertical component of displacement accommodated along the southwestern margin-bounding fault. Rapid cooling along the southern NW basin margin (samples GR12 and GR3), however, continued unabated until Recent times. Coeval with the observed change in cooling rates along the Beto southwest margin, samples from the northern segment of the northwest basin margin (GR2 and GR14) began to cool, possibly reflecting the initiation of a new fault segment northeast of the town of Beto (Figure 2). At first, cooling rates were gradual (~ 2.5 -3 °C/Ma). Around 8-6 Ma however, cooling rates observed for the northern segment of the NW margin substantially increased (~ 9 -12.5 °C/Ma).

Such a north-eastward migration of strain along the NW margin is consistent with the triangular geometry and northward narrowing of the Beto Basin and the local, southward increase in the amount of extension in the Gofa province, which collectively imply large-scale counter clockwise rotation [Philippon *et al.*, 2014]. This model also predicts that the ~NW-SE trending faults delimitating the southwest margins of the Gofa Province basins must have acted as dextral transfer zones during large-scale rotation. Low-temperature thermochronology data, indicating a concurrent north-eastward migration of strain along the Beto Basin's NW margin and cessation or significant decrease in footwall uplift along its SW margin ca. 14 Ma, are consistent with this scenario and may indicate the onset of large-scale counter clockwise rotation.

4.6.3 Amaro Horst Miocene-Recent thermal evolution and the development of the Galana Basin

An early Miocene onset of rift-related denudation in the Amaro Horst is recorded by a period of rapid cooling (~ 10 °C/Ma) beginning ca. 18 Ma in the thermal history model of upper 1st traverse samples (AM2- to AM6-1st tr). Amaro Horst samples collected further north along the 2nd traverse also began to experience cooling around this time (~ 20 -18 Ma), albeit at a significantly slower rate (~ 2 °C/Ma).

Lower 1st traverse samples, however, continued to experience heating until ~ 13 Ma, by which time the difference in paleotemperatures of the lowermost upper 1st traverse sample (AM6-1st) and the uppermost lower 1st traverse sample (AM7-1st) was ~45 °C despite their present-day proximity (Figure 4.6). This suggests that the intervening normal fault must have accommodated a significant amount of throw (perhaps up to ~1.5 km) between ~18 Ma and 13 Ma to be able to decouple upper and lower traverse thermal histories to such a degree. It is perhaps this period of accommodation space creation which facilitated the deposition of lacustrine sediments and interbedded basaltic flows dated between 18-14 Ma in the SW Galana Basin [Woldegabriel *et al.*, 1991; Ebinger *et al.*, 1993].

At ~13 Ma, basement upper 1st traverse samples transitioned to a period of substantially reduced cooling (~0.5 °C/Ma) or thermal stability that has continued until the present day. At the same time, lower 1st traverse samples (AM7- to AM2-1st tr) began a pronounced period of cooling (~4.5 °C/Ma) that continued into the Quaternary. This marked change in southern Amaro Horst cooling history is interpreted as recording an eastward migration of strain in the southern Galana Basin, subsequently accommodated on a fault or faults basin-wards of the lower 1st traverse samples. Consequently, the location of the lower 1st traverse samples in the more proximal footwall of the newly active fault resulted in their rapid uplift and exhumation. By contrast, the more distal position of upper 1st traverse samples relative to the newly active fault resulted in reduced uplift and exhumation rates [Barnett *et al.*, 1987].

Around ~6 Ma, cooling rates recorded by 2nd traverse samples increased from ~2 °C/Ma to ~15 °C/Ma, suggesting a substantial increase in exhumation rates in the northern Amaro Horst during latest Miocene-Pliocene time. By contrast, cooling rates of the nearby Amaro Horst 1st traverse samples remained constant during this period, arguing against a decrease in geothermal gradient or climate-induced change in erosion rates as a plausible explanation. Instead, this change may reflect an important period of strain reorganisation in the normal fault system bounding the Galana Basin's western margin.

The thermal history model from sample AM1-South suggests this segment of the Segen Basin's eastern margin experienced no more than 20 °C of cooling, equivalent to a maximum of ~0.7 km of denudation, during Miocene to Recent times. However, this period of cooling is poorly constrained by its Cretaceous AFT and three Cretaceous-Paleogene AHe aliquot ages. Only one apatite grain (OT) from AM1-South yielded a Neogene age (5.4 ± 0.3 Ma)

consistent with inferred Miocene-Recent cooling. Despite this, a middle Miocene onset of Segen Basin development is consistent with the stratigraphic relationship of syn-rift lacustrine mudstones that onlap ~17-12 Ma phonolites within the basin [Ebinger *et al.*, 2000].

4.6.4 Comparison of newly acquired and previous thermal history models

In general, time-temperature reconstructions presented here are broadly consistent with the low-temperature thermal evolution previously suggested for the Beto Basin and Amaro Horst regions [Philippon *et al.*, 2014; Balestrieri *et al.*, 2016]. One important difference resulting from the application of multiple thermochronometers during thermal history modelling is an improvement in time-temperature resolution, elucidating polyphase Miocene-Recent cooling histories that were previously undetectable. However, the greatest disagreement between new and previous thermal history models concerns the timing of onset of Miocene cooling suggested for Beto Basin NW margin and Amaro Horst samples. The addition of newly acquired AHe analyses and additional confined AFT measurements have resulted in thermal history models suggesting a coeval early Miocene (~20-18 Ma) onset of cooling in these regions, significantly earlier than the middle-late Miocene initiation (14-10 Ma Beto Basin, 10-8 Ma Amaro Horst) suggested previously [Balestrieri *et al.*, 2016].

In certain cases, this difference can be attributed to the addition of AHe analyses during thermal history modelling. For example, the retention of Cretaceous AFT ages in Amaro Horst 1st traverse samples indicate that these rocks did not cool through the AFT PAZ (~120-60 °C) during Miocene-Recent times. Therefore, their AFT data alone poorly constrain the Miocene onset of rift-related denudation. The addition of data from the lower temperature AHe system (~90-30 °C) provides improved resolution over the range of temperatures corresponding to the maximum paleotemperatures experienced by Amaro Horst 1st traverse samples prior to Miocene cooling (~70-90 °C, Figure 4.6). Thus, the earlier onset of cooling ~20-18 Ma suggested by thermal history modelling of combined AFT and AHe data is considered more robust than previous models which were constrained by AFT data alone.

The robustness of thermal history models for samples from the Beto Basin NW margin and Amaro Horst 2nd traverse also improved due to increased low-temperature resolution provided by the addition of AHe analyses. However, the earlier onset of Miocene cooling indicated by thermal history models in this study is attributed to their greatly improved confined fission track measurement statistics resulting from supplemental analysis using the

^{252}Cf irradiation method. The number of confined tracks available for measurement improved substantially (by between 6 and 3750%). The most extreme example being sample GR2, whose number of confined tracks increased from 2 to 77 tracks. The application of the ^{252}Cf irradiation method to samples AM5-2nd tr, AM4-2nd tr, and AM1-2nd tr, which previously had no confined track length measurements, resulted in 76, 5, and 6 measurements, respectively. In all cases, MTL decreased from relatively long lengths [15.2-11.9 μm ; *Balestrieri et al.*, 2016] to more moderate values (13.4-11.2 μm), indicative of more prolonged residence in the AFT PAZ. Consequently, new thermal history results reveal these rocks began cooling through the PAZ earlier than previously suggested, albeit from similar maximum paleotemperatures (~120-110 °C).

4.7. Tectonic Implications

Low-temperature thermochronology of Precambrian basement rocks from the Beto Basin and Amaro Horst regions reveal complex tectono-thermal evolution trends related to the development of the EARS in southern Ethiopia. Moreover, the addition of AHe analyses and supplemental confined AFT measurements reveal higher time-temperature resolution, polyphase cooling histories for these basin margins, providing previously unavailable insights into fault propagation and strain migration trends within the BRZ.

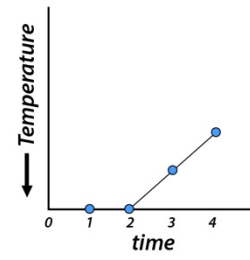
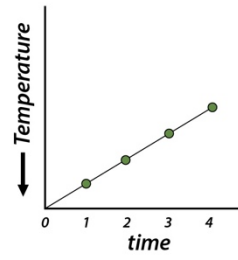
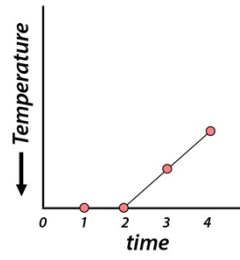
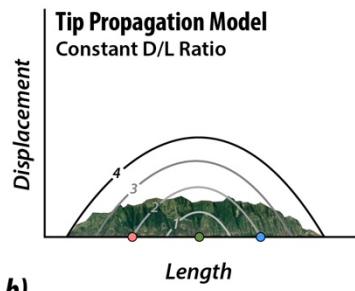
4.7.1 Thermochronological insights into normal fault array growth

Boundary fault systems grow via elongation and linkage with other fault segments and the accumulation of displacement [*Kim and Sanderson*, 2005; *Fossen and Rotevatn*, 2016]. Observations of outcropping and subsurface faults suggest that a scaling relationship exists between the length (L) of a fault and the maximum amount of displacement accommodated along it (D) [e.g. *Watterson*, 1986; *Walsh and Watterson*, 1988; *Marrett and Allmendinger*, 1991; *Cowie and Scholz*, 1992; *Dawers et al.*, 1993]. However, scatter in observed D-L relationships [*Cartwright et al.*, 1995] and syn-rift sedimentation patterns [e.g. *Gawthorpe et al.*, 1997; *Gupta et al.*, 1998; *Sharp et al.*, 2000] have led to the idea that crustal-scale fault systems evolve by one or a combination of two end-member mechanisms, the segment growth and linkage model [*Anders and Schlische*, 1994] and the constant fault length model [*Walsh et al.*, 2002, 2003].

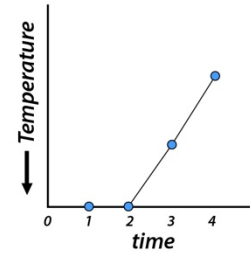
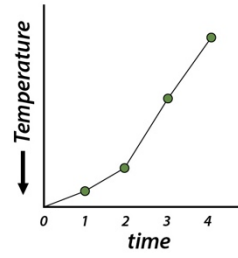
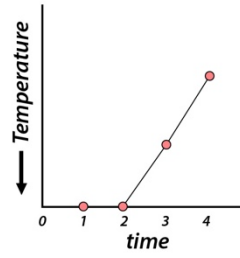
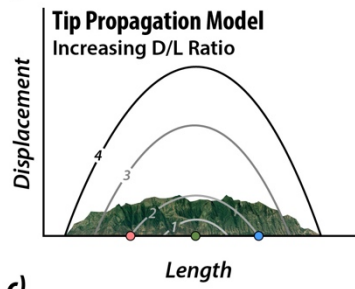
The segment growth and linkage model envisions an initial population of individual isolated fault segments which grow by tip propagation (Figures 4.7a and b), whereby the segments incrementally elongate and accumulate slip, often in association with earthquake events [Walsh and Watterson, 1988; Cowie and Scholz, 1992a, 1992b]. During growth by tip propagation, the D/L ratio of a fault may remain constant throughout its development, as sometimes suggested for small, isolated faults (Figure 4.7a) [Kim and Sanderson, 2005]. However, other authors have demonstrated that D/L ratios are relatively low during the early stages of fault growth and substantially increase later during fault development (Figure 4.7b) [e.g. Filbrandt *et al.*, 1994]. Prior to linking, the adjacent fault segments begin to mechanically interact, often producing asymmetric D-L profiles (Figure 4.7c, t2) [e.g. Peacock and Sanderson, 1994; Cartwright *et al.*, 1995; Gawthorpe and Leeder, 2000]. Eventually the individual fault segments link with one another to form a larger through-going fault [Trudgill and Cartwright, 1994]. The segment of the through-going fault coinciding with the area of paleo-fault segment linkage (green circle in Figure 4.7c) is initially under-displaced (Figure 4.7c, t3). It must therefore, rapidly accumulate displacement in order for the linked fault array to develop an appropriate D/L ratio (Figure 4.7c, t4) [Gawthorpe and Leeder, 2000].

By contrast, the constant fault length model (Figure 4.7d) suggests that normal fault arrays attain their full length relatively early in their development, after which growth is facilitated via displacement accumulation with little elongation [Morewood and Roberts, 1999]. Under this model, large faults may still form by the linkage of smaller faults. However, linkage occurs much more rapidly, with the fault system achieving its full, or near full, length prior to the accumulation of significant displacement. This fault evolution scheme is often applicable to reactivated structures, where fault lengths are inherited from pre-existing crustal discontinuities [e.g. Giba *et al.*, 2012]. The model may also apply where fault propagation is retarded by the interaction with other fault tips [e.g. Walsh *et al.*, 2002].

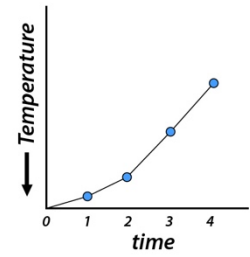
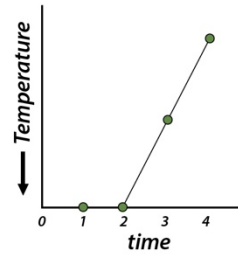
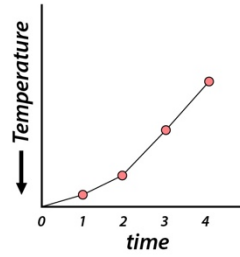
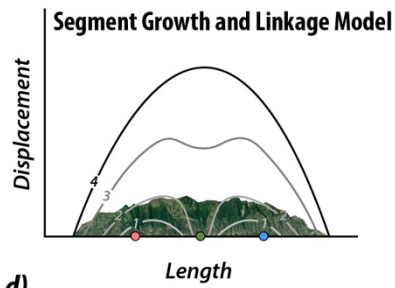
a)



b)



c)



d)

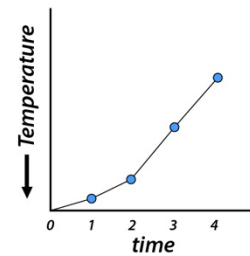
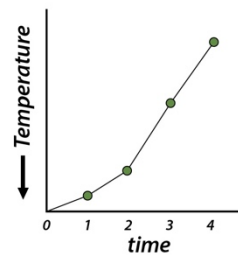
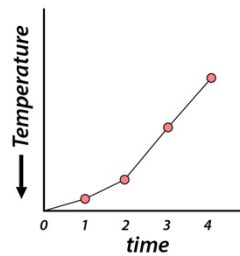
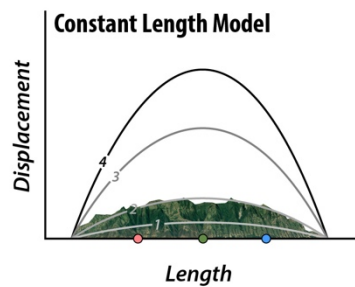


Figure 4.7 (Previous page). Idealised fault displacement (D) versus length (L) plots (left column) and theoretical along-strike thermal history trends (rightmost three columns) for different normal fault array growth models: **a)** Tip propagation constant D/L model generally reserved for small, isolated faults, where the ratio between fault length and displacement remain constant through time [Kim and Sanderson, 2005]; **b)** Tip propagation increasing D/L model, where fault displacement increases with respect to fault length over time [e.g. Filbrandt et al., 1994]; **c)** Fault growth and linkage model, where multiple (in this case two), initially independent fault segments grow via tip propagation before interacting and linking to form a through-going fault array [Anders and Schlische, 1994; Cartwright et al., 1995; Gawthorpe and Leeder, 2000; Gupta et al., 1998; Peacock and Sanderson, 1994; Trudgill and Cartwright, 1994]; **d)** Constant length model, in which a fault system approaches or obtains its full length prior to the accumulation of significant displacement [Morewood and Roberts, 1999; Walsh et al., 2002, 2003]. For each fault growth model, theoretical cooling histories are presented for three basement rock samples (red, green and blue circles) collected along strike from the base of the fault scarp. Note, temperature increases down the y-axis while time increases along the x-axis, following the convention of thermal history models presented in low-temperature thermochronology studies (see Figure 4.6). Relative temperatures for each rock sample are shown for time periods 0-4, corresponding to the four stages of fault growth illustrated in the D/L plot, whereby temperature is assumed to be inversely proportional to displacement. In nature, temporal and spatial variations in along-strike footwall cooling histories may be further complicated by factors such as fault strength heterogeneities [Cowie and Shipton, 1998], the efficacy of surface processes [Densmore et al., 2004; Densmore and Guidon, 2005] and antecedent topography [Densmore et al., 2009]. Time 0 represents the period prior to fault array formation. D/L plots adapted from Kim and Sanderson [2005].

As normal fault systems grow and accumulate displacement, their footwall experiences a predictable uplift and denudational history [Harbor, 1997; Densmore et al., 2007], resulting in the development of an expected pattern of topography [Densmore et al., 2004]. Thus, these fault growth-footwall exhumation relationships can be extrapolated to develop distinct, along-strike cooling history patterns, detectable by low-temperature thermochronology, that reflect the mechanism by which a given normal fault system has evolved (Figure 4.7) [Curry et al., 2016; Mortimer et al., 2016; Brown et al., 2017]. Importantly, this assumes that the amount of footwall exhumation associated with faulting is sufficient to cool rocks from within or through the temperature sensitivities of the AFT and AHe systems. For instance, if a fault system develops via the constant length model (Figure 4.7d), it would attain its total, or near-total, length very early in its evolution and subsequently begin to accumulate displacement along its entire length synchronously, albeit at different rates [Walsh et al., 2002]. This, in turn, would be reflected by a concurrent initiation of cooling along strike [Curry et al., 2016]. Initial cooling rates may be slow but as the fault subsequently accumulates more displacement, with the highest displacement rate being recorded somewhere in the centre of the fault segment [Dawers et al., 1993], cooling rates would

increase (Figure 4.7d, t₂). Following this transition, the central footwall segment would record faster cooling rates and greater total cooling compared to near the fault tips.

If, however, a fault system develops via the segment growth and linkage model, its footwall will record a temporally and spatially variable onset of cooling along strike (Figure 4.7c) [Curry *et al.*, 2016; Mortimer *et al.*, 2016]. Initially, isolated, low-displacement fault segments would develop, with slow displacement rates initially prevailing as extension is distributed over many smaller fault segments [Gupta *et al.*, 1998]. Accordingly, cooling rates may be initially protracted (Figure 4.7c, red & blue circles, t₀-2). Subsequent elongation, interaction and linkage of the isolated fault segments would be reflected by footwall segments that correspond to regions where paleo-fault segments once interacted and linked (Figure 4.7c, green circles) recording a relatively later onset of cooling. The development of a hard linkage between fault segments is then marked by a rapid increase in uplift and cooling rates [Gupta *et al.*, 1998], as displacement along the fault increases to rectify its D/L deficiency (Figure 4.7c, green circles, t₂-4) [Gawthorpe and Leeder, 2000; Mortimer *et al.*, 2016].

Expected relationship between fault growth and along-strike footwall cooling histories, however, can be complicated by natural factors, such as fault strength heterogeneities [Cowie and Shipton, 1998], the efficacy of surface processes [Densmore *et al.*, 2004; Densmore and Guidon, 2005] and antecedent topography [Densmore *et al.*, 2009]. Moreover, the ability of combined AFT and AHe low-temperature thermochronology to resolve small differences in cooling behaviour between samples is limited by the quality of data, the accuracy of the AFT annealing and AHe diffusion kinetic models used to produce time-temperature reconstructions [Flowers *et al.*, 2015], the biases inherent in the inverse modelling approach of the thermal history modelling software employed [Vermeesch and Tian, 2014] and the modelling parameters [Gallagher and Ketcham, 2018]. Nevertheless, marked along-strike differences in the onset, degree and rate of cooling experienced by footwall rocks has been used to successfully differentiate between end-member fault growth mechanisms [Curry *et al.*, 2016; Mortimer *et al.*, 2016; Brown *et al.*, 2017].

4.7.2 Insights into normal fault system growth patterns in the BRZ as revealed by combined AFT and AHe thermochronology

Combined AFT and AHe thermochronology of the Beto and Galana basin margins reveal spatially and temporally variable, along-strike denudational cooling patterns that are interpreted to reflect fault system growth and displacement trends. The concurrent early Miocene (~20-18 Ma) initiation of observed cooling in the Amaro Horst upper 1st and 2nd traverses, is consistent with this portion of the Galana basin-bounding normal fault attaining its full length prior to accumulating significant displacement. Such an along-strike, footwall thermal history trend is consistent with the expected pattern for a fault array that developed via the constant length model (Figure 4.7d) [Walsh *et al.*, 2002]. Extending this interpretation to the entire Galana Basin bounding fault system is problematic, as the 12 km distance between 1st and 2nd traverses spans only a portion of the more than 90 km length of the fault array. Nevertheless, the long, continuous and linear geometry of the normal fault system bounding the eastern Amaro Horst is compatible with this interpretation. A similar concomitant onset of cooling observed in AHe data along-strike of the Jackson and Pine Forest Ranges of the northwest United States was used by Curry *et al.* [2016] to propose that the fault systems bounding those ranges also developed via this mechanism.

At ~13 Ma, strain accommodated in the central Galana master fault system migrated basin-wards, as suggested by marked changes in cooling behaviour of upper and lower 1st traverse thermal history models (Figure 4.6 and Section 4.6.3). Similar patterns of strain localising towards the rift centre during progressive rifting is well-documented throughout the EARS [e.g. Chapman *et al.*, 1978; Morley *et al.*, 1990; Bosworth, 1994; Boccaletti *et al.*, 1998; Bonini *et al.*, 2005; Keir *et al.*, 2006]. Indeed, the shift of deformation from large boundary faults to within a narrow trough in the rift axis as extension progresses is characteristic of the Ethiopian Rift [Hayward and Ebinger, 1996; Corti, 2009]. Progressive rifting led to the eventual abandonment of large border faults as the stress required to drive slip along these deeply penetrating faults exceeded that required to initiate slip on new faults within the weakened lithosphere of the rift valley [Forsyth, 1992]. Circa 2 Ma, this process led to the development of a long chain of overlapping, right stepping en-echelon fault segments called the Wonji Fault Belt which runs from the Afar to the southern MER, where the Wonji faults are considerably less well developed [Mohr, 1962; Boccaletti *et al.*, 1998; Ebinger and Casey, 2001; Wolfenden *et al.*, 2004]. While Wonji faults have yet to develop in the Galana Basin, the basin-ward migration of rapid rift-related cooling observed in Amaro Horst 1st

traverse samples around 13 Ma suggests the shift of deformation towards a narrow, interior trough may have been a gradual process that started relatively early in the development of these rifts segments.

By contrast, samples from the Beto Basin NW margin exhibit footwall cooling histories with substantial spatial and temporal variation along-strike. Rift-related denudation began in the early Miocene around ~18 Ma in the faulted margins bounding the southwest end of the basin. The concurrent initiation of cooling ca. 18 Ma, similar monotonic cooling behaviour and cooling rates (~4.5 °C/Ma) recorded by sample GR3 and GR12 thermal history models suggests that the fault segments bounding the southern Beto Basin were mechanically linked prior to the accumulation of significant displacement. Only later, at ~14 Ma did denudational cooling of the northern NW margin (samples GR2 and 14) begin. By then, samples GR3 and GR12 had experienced a significant amount of cooling (~10 °C). Thus, the delayed onset of cooling in the northern Beto Basin NW margin is interpreted to represent the formation of a new, initially independent fault segment north of the Beto township. In this scenario, the rapid increase in cooling rates observed for samples GR2 and GR14 ca. 8-6 Ma reflect increased late Miocene-Recent displacement rates along the northern fault segment following its interaction and linkage to the already established fault system further south (samples GR3, 12 and 13), resulting in the development of a through-going boundary fault system along the Beto Basin NW margin. By this time, southern NW margin samples (GR12 and 3) had cooled ~50 °C (equivalent to ~1.5-2 km of exhumation).

Such a temporally and spatially variable along-strike footwall thermal history trend for the Beto Basin is consistent with what may be expected for a fault array that developed via the segment growth and linkage model (Figure 4.7c), whereby tension is first accommodated by the initiation of numerous small displacement faults that evolve into a longer, linked, and higher displacement fault array [Cartwright *et al.*, 1995; Gawthorpe and Leeder, 2000; Gawthorpe *et al.*, 2003; Cowie *et al.*, 2006]. The initiation of the Beto Basin bounding master fault as a collection of initially isolated, smaller fault segments is consistent with its present-day, highly segmented geometry (Figure 4.2) [Moore Jr and Davidson, 1978; Philippon *et al.*, 2014]. A similarly disjointed, along-strike cooling history was observed in the Livingstone Fault footwall, in the northern Malawi Rift, where along strike variation in AFT and AHe ages were interpreted to record the linkage of two initially isolated fault segments [Mortimer *et al.*, 2016].

4.7.3 The potential role of structural inheritance in disparate BRZ basin evolution patterns

The pronounced contrast between the structural evolution of the Beto and Galana basins suggests that longitudinal differences in kinematic, mechanical and thermal parameters may have existed within the BRZ during its Neogene-Recent rifting history. While it is unclear if the kinematic boundary conditions have remained constant during the Miocene-Recent evolution of the Ethiopian Rift [Corti, 2009 and references therein], any such changes should have affected the relatively proximal Beto and Galana Basins equally during their coeval evolutions. Their contrasting fault geometries are not easily explained by differences in mechanical stratigraphy [Walsh *et al.*, 2003; Schöpfer *et al.*, 2006] either, as both master fault systems developed in the Neoproterozoic basement of the Mozambique Belt and overlying Gamo Amaro basalts. Possible west to east differences in the paleo-thermal state of the southern Ethiopian lithosphere are more difficult to determine, in part due to the scarcity of subsurface data in the region. However, indications of major differences in crustal thickness or mantle structure across the modern BRZ that might be suggestive of east-to-west discrepancies in paleo-heat flow are lacking. Instead, estimates of present-day crustal thicknesses across the BRZ are similar [around 30 km; Keranen and Klemperer, 2008], albeit poorly constrained. Furthermore, seismic tomography reveals that the broad low-velocity region in the upper mantle beneath Ethiopia extends below the BRZ across its entire width [e.g. Bastow *et al.*, 2008] and late Paleogene-early Neogene volcanism appears to have been similarly pronounced throughout the region [Woldegabriel *et al.*, 1991; Ebinger *et al.*, 1993, 2000].

One potential parameter that may have led to the contrasting structural evolutionary styles exhibited by the Beto and Galana Basins is the asymmetric distribution of pre-existing basement weaknesses in the BRZ. Moore and Davidson [1978] noted that in the Gofa Province (i.e. Beto Basin), the arcuate pattern of basins and ranges reflect local foliation trends [$\sim$ N65°E and $\sim$ N155°E; Philippon *et al.*, 2014] in the basement gneisses with faults tending to develop parallel or orthogonal to the basement structures. The markedly similar orientation of the Beto Basin-bounding faults along its northwest ($\sim$ N45°E) and southwest ($\sim$ N150°E) margins suggest that its rift-related structural evolution was strongly controlled by the pre-existing basement fabric. A similar relationship between fault development and inherited basement heterogeneities has also been proposed for the Chew Bahir Basin, $\sim$ 50 km to the south [Philippon *et al.*, 2014].

By contrast, metamorphic foliations within the Precambrian basement of the eastern BRZ (i.e. Galana Basin) trend N-S [Ebinger *et al.*, 1993]. Moreover, a deep-seated, ~N-S structure may be indicated by a significant change in crustal thickness beneath the surrounding eastern (38-40 km) and western (34-36 km) rift shoulders [Keranen and Klemperer, 2008], corresponding to the position of the Galana Basin and Southern MER. This jump in crustal thickness has been interpreted as a boundary between two different crustal terranes, corresponding to the location of a Neoproterozoic N-NNE trending suture between East and West Gondwana [Berhe, 1990; Stern *et al.*, 1990; Keranen and Klemperer, 2008]. Thus, the long and linear nature of the N-S trending fault system bounding the Galana Basin may reflect both the absence of transversal basement foliations and the local prevalence of both shallow and deep N-S lithospheric fabrics.

Inherited lithospheric weak zones and structures play an important role in controlling the location and evolution of intracontinental extensional deformation [e.g. Burke and Dewey, 1973; Zoback and Zoback, 1980; Cheadle *et al.*, 1987; Morley *et al.*, 2004; Wijk, 2005; Morley, 2016]. This is particularly true for the EARS, which preferentially follows a trend of Late Archean-Proterozoic mobile belts, circumventing stronger cratonic regions [Mohr, 1962; McConnell, 1972; Daly *et al.*, 1989; Smith and Mosley, 1993; Hetzel and Strecker, 1994]. In part, this is thought to be because deforming thinner, hotter mobile belt lithosphere is more energetically efficient than rifting thicker, colder, and more rigid cratonic terranes [e.g. McConnell, 1972; Sykes, 1978; Vauchez *et al.*, 1997]. However, mechanical anisotropies, such as steep basement shear zones and sutures within the mobile belts, may also guide principal stress during rifting [Daly *et al.*, 1989; Smith and Mosley, 1993]. The importance of deep-seated lithospheric anisotropies in the structural evolution of rift segments has been documented throughout the EARS [Morley, 1999b], from the Red Sea [Dixon *et al.*, 1987], through the Ethiopian [Stuart *et al.*, 2006; Keranen and Klemperer, 2008; Keranen *et al.*, 2009] and Kenyan Rifts [Smith and Mosley, 1993; Hetzel and Strecker, 1994], to numerous segments in the western branch of the EARS [Versfelt and Rosendahl, 1989], including the nascent Okavango rift zone in Botswana [Kinabo *et al.*, 2007]. These penetrative fabrics have provided a matrix of mechanical anisotropies prone to reactivation during extension normal to foliation planes [Vauchez *et al.*, 1997, 1998]. Their pervasiveness throughout the East African Proterozoic mobile belts may explain why an overwhelming majority of crustal-scale normal faults in the EARS, like the Galana Basin bounding fault array, are thought to have

evolved via the constant length model, achieving significant elongation prior to the accumulation of substantial displacement [e.g. *Morley*, 1999c, 2002].

4.7.4 EARS evolution and the important role of rheological heterogeneities

The EARS records a complex and spatially disparate late Paleogene-Recent evolution, with associated magmatism and extensional deformation initiating diachronously along its length [*Zeyen et al.*, 1997; *Morley*, 2002b; *Torres Acosta et al.*, 2015]. Its irregular propagation of rifting is thought to have been controlled by rheological heterogeneities and crustal structures inherited from earlier periods of Proterozoic orogenesis and Phanerozoic rifting [e.g. *Morley et al.*, 1992; *Hetzl and Strecker*, 1994; *Vétel et al.*, 2005; *Bastow et al.*, 2008; *Keranen and Klemperer*, 2008].

In general, volcanism preceded extensional basin formation during the development of the EARS [e.g. *Morley*, 1994; *Morley et al.*, 1999c; *Corti*, 2009]. The earliest known manifestation of EARS-related volcanism occurred in the BRZ and northernmost Turkana Depression, as expressed by the emplacement of the ~45-35 Ma Amaro-Gamo basalt flows [*Davidson and Rex*, 1980; *Ebinger et al.*, 1993; *George et al.*, 1998], ~15 Myr before the eruption of flood basalts in the Afar ca. 31-29 Ma [*Baker et al.*, 1996; *Hofmann et al.*, 1997; *George et al.*, 1998] and a pronounced period of Turkana volcanism ca. 30 Ma [*Morley et al.*, 1992]. By Oligocene time, EARS-related extension had begun with the development of the ~N-S trending Lokichar and Lothidok basins in the Turkana Depression [*Morley et al.*, 1999b; *Boone et al.*, in review]. Both the early onset of volcanism and rift basin formation in the Turkana Depression were likely associated with the regionally thinned lithosphere, resulting from earlier Cretaceous-early Paleogene Anza-South Sudan rifting [*Ebinger et al.*, 2000]. Consequently, the pre-deformed Turkana lithosphere was perhaps innately weaker [*Morley et al.*, 1992] and/or its shallower lithosphere-asthenosphere boundary allowed for ponding of plume material during the Paleogene [*Ebinger and Sleep*, 1998].

The general lack of Paleogene aged sedimentary rocks, parallelism between Paleogene and Miocene lava flows and the relative consistency in thicknesses of volcanic flows in the BRZ argue against extensional basin development in southern Ethiopia prior to ~20 Ma [*Ebinger et al.*, 2000]. Combined AFT and AHe thermochronology presented in this study suggest an approximately coeval onset of rift-related denudation in the Beto Basin margin and the Amaro Horst during early Miocene time (~20-18 Ma), earlier than the middle-late Miocene

(14-10 Ma Beto Basin, 10-8 Ma Amaro Horst) initiation suggested by earlier thermal history modelling of AFT data alone [*Balestrieri et al.*, 2016] (Section 4.6.4). Eruptive centres for trachybasalts and tephrites [18.8-12.7 Ma; *Ebinger et al.*, 2000] that overlie the Gamo Amaro basalts have been identified along normal faults bounding the Gamo-Gidole Horst along the western Chamo Basin margin, suggesting those faults too had formed prior to ~ 18Ma. While a similar early Miocene age has been suggested for the Chew Bahir Basin in the southern BRZ [*Pik et al.*, 2008], other evidence for early Miocene rift basin development elsewhere in the region has been sparse. Instead, widespread extensional basin development in the BRZ was thought to have begun later in the middle-late Miocene [*Corti*, 2009]. However, the refined tectono-thermal evolutions of the Beto and Galana Basin margins presented herein suggest that early Miocene rift basin development in the BRZ may have been more widespread than previously thought, perhaps representing a northward migration of Turkana-related extension into the region [*Bonini et al.*, 2005].

Rift basin formation in the northern MER [~11 Ma; *Wolfenden et al.*, 2004] and central MER [~5-6 Ma; *Bonini et al.*, 2005] only began later in the late Miocene. Perhaps surprisingly, the MER developed on the eastern edge of the wide (~500 km) mantle upwelling beneath Ethiopia, not above its centre [*Bastow et al.*, 2008]. Instead, the location and initial evolution of the MER was probably strongly controlled by the same N-NE trending, lithospheric-scale heterogeneities associated with the Neoproterozoic suture between East and West Gondwana [*Bastow et al.*, 2008; *Keranen and Klemperer*, 2008; *Keranen et al.*, 2009] that governed the development of the Galana Basin.

During Miocene-Recent times, extensional basin formation and magmatism in the Turkana Depression has propagated eastward [*Morley et al.*, 1992, 1999b; *Wescott et al.*, 1999], possibly due to strain hardening of the lithosphere during extension [*Kusznir and Park*, 1987] caused by relatively slow strain rates [between 10^{-16} s^{-1} and 10^{-15} s^{-1} ; *Morley*, 1994]. However, the general eastward decrease in age of deformation that is so apparent in central Turkana becomes less clear further north, where all northern Turkana basins may host late Paleogene-Miocene to Pliocene deposits [Figure 26 in *Morley et al.*, 1999b]. Similarly, thermochronology data (this study) and hydrography and geomorphological data [*Woldegabriel and Aronson*, 1987; *Philippon et al.*, 2014] suggest extensional deformation has continued to affect the breadth of the BRZ from Miocene to Recent time.

During the late Pliocene-Pleistocene, faulting and magmatism in the MER and northern Kenyan rifts localized within narrow zones of minor fault swarms, fissures and eruptive centres [Ebinger and Casey, 2001; Vétel et al., 2005; Vétel and Le Gall, 2006; Corti, 2009]. The fundamental change in structural style from fault-bounded grabens to minor fault swarms may be related to significant crustal thinning followed by upwelling and shallow emplacement of magma beneath the fault swarms [Hayward and Ebinger, 1996; Ebinger, 2005; Corti, 2009]. By contrast, border faults in the BRZ and southernmost MER are still accommodating some degree of strain. Extension in the Gofa Province has largely ceased apart from in its southern sector (e.g. the Beto Basin) where seismicity, hydrography data and master fault scarp morphology, characterized by prominent triangular facets, indicate a degree of ongoing fault activity [Woldegabriel and Aronson, 1987; Philippon et al., 2014]. Late Pleistocene-Holocene radiocarbon ages [$\sim$ 26-23 kyr and 7 kyr ; Philippon et al., 2014] from fault gouge collected along the western Galana Basin margin suggest that activity along these structures is also ongoing. Yet, the step-fault arrangement of the western Galana Basin margin and thermal history modelling results suggest a degree of strain localisation towards a narrower zone of rifting in the eastern BRZ and southernmost MER that is absent further west. Nonetheless, while modern-day extension in the MER, eastern Turkana Depression and Kenyan Rift is accommodated across a narrow (< 100 km) corridor, strain in the BRZ is still accommodated across much of its ~ 300 km breadth in a diffuse zone of deformation. As lithospheric strength is fundamentally controlled by crustal thickness [Kusznir and Park, 1987], the unique style of modern deformation in the BRZ may be explained by the distinct regional rheological characteristics resulting from the northward thickening of the crust from the Turkana Depression [~ 20 km, Benoit et al., 2006a] to the BRZ [~ 30 km; Keranen and Klemperer, 2008], as lithospheric strength is fundamentally controlled by crustal thickness [Kusznir and Park, 1987]. Kusznir and Park [1987] showed that above a critical lithospheric extension factor (β) of 1.5, slow strain rates lead to rapid strain hardening. Therefore, the reduced degree of crustal thinning experienced by the BRZ [maximum β of 1.2, Ebinger et al., 1989] compared to that of Turkana [maximum β of 1.55-1.65, Hendrie et al., 1994] may have thus far prevented the onset of significant strain hardening. Nevertheless, the hot, weak lithosphere of the BRZ still promotes widely-distributed deformation. Alternatively, the pre-existing basement structures of the eastern BRZ may not be as well developed or as ideally situated as those in the eastern Turkana Depression to encourage a narrowing of the rift zone and abandonment of boundary faults throughout the region.

4.8. Conclusions

Low-temperature thermochronology data constrain the timing of Beto and Galana Basin formation to be early Miocene in age (20-18 Ma). These findings, along with previously published AHe data interpreted as indicating a concurrent (~20 Ma) formation of the Chew Bahir Basin, suggest that EARS-related extension began concurrently across the width of the BRZ in the early Miocene, at least 7 Myr prior to the formation of the late Miocene MER further north. The earlier initiation of rifting in the BRZ was likely related to the inherent weakness of the regionally thinned lithosphere, a consequence of earlier Cretaceous-early Paleogene Anza-South Sudan rifting. The pre-existing lithospheric structure of the region was also responsible for Neogene EARS extension being accommodated by a diffuse zone of deformation, as opposed to the MER's narrow rift morphology exhibited above the colder, thicker Ethiopian Dome.

The increased time-temperature resolution provided by combined AFT and AHe thermochronology presented herein reveal spatially and temporally variable basin margin denudation patterns, and thus insights into master fault system growth and displacement trends. Along strike differences in thermal history models suggests that the normal fault array bounding the Beto Basin's NW margin initially developed as isolated and kinematically independent fault segments. These segments only linked relatively late in their history, after accumulating significant displacement, consistent with the segment growth and linkage model. By contrast, low-temperature thermochronology data suggest that the Galana Basin-bounding fault system had approached at least a considerable portion of its final length prior to the accumulation of significant displacement, more compatible with the constant fault length model. This interpretation is consistent with the long (> 90 km), linear nature of the basin-bounding fault arrays of the eastern BRZ and southern MER. At ~13 Ma, strain in the Galana Basin began to migrate towards the rift valley with the formation of a new normal fault in the former hanging wall. This marked the beginning of a pattern of strain localisation and progressive narrowing of the zone of deformation in the eastern BRZ, not exhibited further west in the Gofa province. However, unlike in the MER and northern Kenyan Rift, where large boundary faults were abandoned in favour of low-displacement fault swarms and associated intrusive complexes, border faults across the BRZ remain active.

The longitudinal contrast in boundary fault system growth reflects differences in the distribution of pre-existing basement fabrics in the BRZ. The highly-segmented fault system bounding the Beto Basin reflects local basement foliation trends in the west of the region, while the long, linear faults defining the Galana western margin mirror an ~N-S trending, Neoproterozoic suture zone in the eastern BRZ. This strongly suggests that the disparate structural evolution of the BRZ was controlled by inherited lithospheric heterogeneities.

References

- Abebe, T., M. L. Balestrieri, and G. Bigazzi (2010), The Central Main Ethiopian Rift is younger than 8 Ma: confirmation through apatite fission-track thermochronology, *Terra Nov.*, 22(6), 470–476, doi:10.1111/j.1365-3121.2010.00968.x.
- Anders, M. H., and R. W. Schlische (1994), Overlapping Faults, Intrabasin Highs, and the Growth of Normal Faults, *J. Geol.*, 102, 165–180.
- Arambourg, C., and R. G. Wolff (1969), Nouvelles données paléontologique sur l'âge des “Gres du Lubur” (Turkana Grits) al'ouest du lac Rodolphe, *Comptes Rendus Société géologique Fr.*, 9, 190–202.
- Baker, B. H., P. A. Mohr, and L. A. J. Williams (1972), Geology of the Eastern Rift System of Africa, in *Geological Society of America Special Papers*, vol. 136, pp. 1–68.
- Baker, J., L. Snee, and M. Menzies (1996), A brief Oligocene period of flood volcanism in Yemen: implications for the duration and rate of continental flood volcanism at the Afro-Arabian triple junction, *Earth Planet. Sci. Lett.*, 138, 39–55.
- Balestrieri, M. L., F. M. Stuart, C. Persano, E. Abbate, and G. Bigazzi (2005), Geomorphic development of the escarpment of the Eritrean margin, southern Red Sea from combined apatite fission-track and (U–Th)/He thermochronometry, *Earth Planet. Sci. Lett.*, 231, 97–110, doi:10.1029/2001JB001009.
- Balestrieri, M. L., M. Bonini, G. Corti, F. Sani, and M. Philippon (2016), A refinement of the chronology of rift-related faulting in the Broadly Rifted Zone, southern Ethiopia, through apatite fission-track analysis, *Tectonophysics*, 671, 42–55, doi:10.1016/j.tecto.2016.01.012.
- Barbarand, J., A. Carter, I. Wood, and T. Hurford (2003), Compositional and structural control of fission-track annealing in apatite, *Chem. Geol.*, 198(1–2), 107–137, doi:10.1016/S0009-2541(02)00424-2.
- Barberi, F., H. Tazieff, and J. Varet (1972), Volcanism in the Afar Depression: Its Tectonics and Magmatic Significance, *Tectonophysics*, 15(1–2), 19–29.
- Barnett, J. A., J. Mortimer, J. H. Rippon, J. J. Walsh, and J. Watterson (1987a), Displacement geometry in the volume containing a single normal fault, *Am. Assoc. Pet. Geol. Bull.*, 71(8), 925–937.
- Barnett, J. A. M., J. Mortimer, J. H. Rippon, J. J. Walsh, and J. Watterson (1987b), Displacement geometry in the volume containing a single normal fault, *Am. Assoc. Pet. Geol. Bull.*, 71, 925–937.
- Bastow, I. D., G. W. Stuart, J.-M. Kendall, and C. J. Ebinger (2005), Upper-mantle seismic structure in a region of incipient continental breakup: northern Ethiopian rift, *Geophys. J. Int.*, 162, 479–493, doi:10.1111/j.1365-246X.2005.02666.x.
- Bastow, I. D., A. A. Nyblade, G. W. Stuart, T. O. Rooney, and M. H. Benoit (2008), Upper mantle seismic structure beneath the Ethiopian hot spot: Rifting at the edge of the African low-velocity anomaly, *Geochemistry, Geophys. Geosystems*, 9(12), doi:10.1029/2008GC002107.
- Bastow, I. D., D. Keir, and E. Daly (2011), The Ethiopia Afar Geoscientific Lithospheric Experiment (EAGLE): Probing the transition from continental rifting to incipient seafloor spreading, *Geol. Soc. Am. Spec. Pap.*, 478, doi:10.1130/2011.2478(04).
- Bendick, R., S. McClusky, R. Bilham, L. Asfaw, and S. Klemperer (2006), Distributed Nubia–Somalia relative motion and dike intrusion in the Main Ethiopian Rift, *Geophys. J. Int.*, 165, 303–310, doi:10.1111/j.1365-246X.2006.02904.x.
- Benoit, M. H., A. A. Nyblade, and M. E. Pasyanos (2006), Crustal thinning between the Ethiopian and East African plateaus from modeling Rayleigh wave dispersion, *Geophys.*

- Res. Lett.*, doi:10.1029/2006GL025687.
- Berhe, S. M. (1990), Ophiolites in Northeast and East Africa: implications for Proterozoic crustal growth, *J. Geol. Soc. London*, *147*, 41–57.
- Boccaletti, M., M. Bonini, R. Mazzuoli, B. Abebe, L. Piccardi, and L. Tortorici (1998), Quaternary oblique extensional tectonics in the Ethiopian Rift (Horn of Africa), *Tectonophysics*, *287*, 97–116.
- Bonini, M., T. Souriot, M. Boccaletti, and J. P. Brun (1997), Successive orthogonal and oblique extension episodes in a rift zone: Laboratory experiments with application to the Ethiopian Rift, *Tectonics*, *16*(2), 347–362.
- Bonini, M., G. Corti, F. Innocenti, P. Manetti, F. Mazzarini, T. Abebe, and Z. Pecskey (2005), Evolution of the Main Ethiopian Rift in the frame of Afar and Kenya rifts propagation, *Tectonics*, *24*(1), n/a-n/a, doi:10.1029/2004TC001680.
- Boone, S. C., C. Seiler, B. P. Kohn, A. J. W. Gleadow, D. A. Foster, and L. Chung (2018), Influence of Rift Superposition on Lithospheric Response to East African Rift System Extension: Lapur Range, Turkana, Kenya, *Tectonics*, *37*, 1–26, doi:10.1002/2017TC004575.
- Bosworth, W. (1994), a Model for the 3-Dimensional Evolution of Continental Rift Basins, Northeast Africa, *Geol. Rundschau*, *83*(4), 671–688.
- Bosworth, W., and C. K. Morley (1994), Structural and stratigraphic evolution of the Anza rift, Kenya, *Tectonophysics*, *236*(1–4), 93–115, doi:10.1016/0040-1951(94)90171-6.
- Bosworth, W., M. R. Strecker, and P. M. Blisniuk (1992), Integration of East African Paleostress and Present-Day Stress Data: Implications for Continental Stress Field Dynamics, *J. Geophys. Res.*, *97*(B8), 11,851–11,865, doi:10.1029/90JB02568.
- Brown, R. W., R. Beucher, S. Roper, C. Persano, F. Stuart, and P. Fitzgerald (2013), Natural age dispersion arising from the analysis of broken crystals. Part I: Theoretical basis and implications for the apatite (U-Th)/He thermochronometer, *Geochim. Cosmochim. Acta*, *122*(120), 478–497, doi:10.1016/j.gca.2013.05.041.
- Brown, S. J., J. R. Thigpen, J. A. Spotila, W. C. Krugh, L. M. Tranel, and D. A. Orme (2017), Onset Timing and Slip History of the Teton Fault, Wyoming: A Multidisciplinary Reevaluation, *Tectonics*, *36*, doi:10.1002/2016TC004462.
- Browne, S. E., and J. D. Fairhead (1983), Gravity study of the Central African Rift system: A model of continental disruption 1. The Ngaoundere and Abu Gabra Rifts, *Tectonophysics*, *94*(1–4), 187–203, doi:10.1016/0040-1951(83)90016-1.
- Bürgmann, R., D. D. Pollard, and S. J. Martel (1994), Slip distribution on faults: effects of stress gradients, inelastic deformation, heterogeneous host-rock stiffness, and fault interaction, *J. Struct. Geol.*, *16*(12), 1675–1690.
- Burke, K., and J. F. Dewey (1973), Plume-Generated Triple Junctions: Key Indicators in Applying Plate Tectonics to Old Rocks, *J. Geol.*, *81*, 406–433.
- Cartwright, J. A., B. D. Trudgill, and C. S. Mansfield (1995), Fault growth by segment linkage: an explanation for scatter in maximum displacement and trace length data from the Canyonlands Grabens of SE Utah, *J. Struct. Geol.*, *17*(9), 1319–1326.
- Chapman, G. R., S. J. Lippard, and J. E. Martyn (1978), The stratigraphy and structure of the Kamasia Range, Kenya Rift Valley, *J. Geol. Soc. London*, *135*(3), 265–281, doi:10.1144/gsjgs.135.3.0265.
- Cheadle, M. J., S. McGeary, M. R. Warner, and D. H. Matthews (1987), Extensional structures on the western UK continental shelf: a review of evidence from deep seismic profiling, *Geol. Soc. Spec. Publ.*, *28*, 445–465.
- Childs, C., T. Manzocchi, J. J. Walsh, C. G. Bonson, A. Nicol, and M. P. J. Schöpfer (2009), A geometric model of fault zone and fault rock thickness variations, *J. Struct. Geol.*, *31*, 117–127, doi:10.1016/j.jsg.2008.08.009.

- Chorowicz, J. (2005), The East African rift system, *J. African Earth Sci.*, 43(1), 379–410, doi:10.1016/j.jafrearsci.2005.07.019.
- Civiero, C. et al. (2016), Small-scale thermal upwellings under the northern East African Rift from S travel time tomography, *J. Geophys. Res. Solid Earth*, 121, 7395–7408, doi:10.1002/2016JB013070. Received.
- Colgan, J. P., T. A. Dumitru, P. W. Reiners, J. L. Wooden, and E. L. Miller (2006), Cenozoic Tectonic Evolution of the Basin and Range Province in Northwestern Nevada, *Am. J. Sci.*, 306, 616–654, doi:10.2475/08.2006.02.
- Commings, D., S. Gupta, and J. Cartwright (2005), Deformed streams reveal growth and linkage of a normal fault array in the Canyonlands graben, Utah, *Geology*, 33(8), 645–648, doi:10.1130/G21433.1.
- Corti, G. (2009), Continental rift evolution: From rift initiation to incipient break-up in the Main Ethiopian Rift, East Africa, *Earth-Science Rev.*, 96(1–2), 1–53, doi:10.1016/j.earscirev.2009.06.005.
- Cowie, P. A., and C. H. Scholz (1992a), Displacement-length scaling relationship for faults: data synthesis and discussion, *J. Struct. Geol.*, 14(10), 1149–1156, doi:10.1016/0191-8141(92)90066-6.
- Cowie, P. A., and C. H. Scholz (1992b), Growth of Faults by Accumulation of Seismic Slip, *J. Geophys. Res.*, 97(B7), 11,085–11,095, doi:10.1029/92JB00586.
- Cowie, P. A., and Z. O. E. K. Shipton (1998), Fault tip displacement gradients and process zone dimensions, *J. Struct. Geol.*, 20(8), 983–997.
- Cowie, P. A., S. Gupta, and N. H. Dawers (2000), Implications of fault array evolution for synrift depocentre development: insights from a numerical fault growth model, *Basin Res.*, 12, 241–261.
- Cowie, P. A., M. Attal, G. E. Tucker, A. C. Whittaker, M. Naylor, A. Ganas, and G. P. Roberts (2006), Investigating the surface process response to fault interaction and linkage using a numerical modelling approach, *Basin Res.*, 18, 231–266, doi:10.1111/j.1365-2117.2006.00298.x.
- Curry, M. A. E., J. B. Barnes, and J. P. Colgan (2016), Testing fault growth models with low-temperature thermochronology in the northwest Basin and Range, USA, *Tectonics*, 35, 2467–2492, doi:10.1002/2016TC004211. Received.
- Daly, M. C., J. Chorowicz, and J. D. Fairhead (1989), Rift basin evolution in Africa: The influence of reactivated steep basement shear zones, *Geol. Soc. Spec. Publ.*, 44, 309–334, doi:10.1144/GSL.SP.1989.044.01.17.
- Danišik, M., R. F. Sachsenhofer, V. A. Privalov, E. A. Panova, W. Frisch, and C. Spiegel (2008), Low-temperature thermal evolution of the Azov Massif (Ukrainian Shield–Ukraine) — Implications for interpreting (U–Th)/He and fission track ages from cratons, *Tectonophysics*, 456, 171–179, doi:10.1016/j.tecto.2008.04.022.
- Danišik, M., B. I. A. McInnes, C. L. Kirkland, B. J. McDonald, N. J. Evans, and T. Becker (2017), Seeing is believing: Visualization of He distribution in zircon and implications for thermal history reconstruction on single crystals, *Sci. Adv.*, 3, 1–9.
- Davidson, A. (1983), *The Omo River Project: reconnaissance geology and geochemistry of parts of Ilubabor, Kefa, Gemu Gofa and Sidamo, Ethiopia*.
- Davidson, A., and D. C. Rex (1980), Age of volcanism and rifting in southwestern Ethiopia, *Nature*, 283, 657–658.
- Davies, S. J., N. H. Dawers, A. McLeod, and J. R. Underhill (2000), The structural and sedimentological evolution of early synrift successions: the Middle Jurassic Tarbert Formation, North Sea, *Basin Res.*, 12, 343–365.
- Dawers, N. H., and J. R. Underhill (2000), The role of fault interaction and linkage in controlling synrift stratigraphic sequences: Late Jurassic, Stratfjord East Area, northern

- North Sea, *Am. Assoc. Pet. Geol. Bull.*, 84(1), 45–64.
- Dawers, N. H., M. H. Anders, and C. H. Scholz (1993), Growth of normal faults: Displacement-length scaling, *Geology*, 21, 1107–1110.
- Densmore, A. L., and R. Guidon (2005), What sets topographic relief in extensional footwalls?, *Geology*, 33(6), 453–456, doi:10.1130/G21440.1.
- Densmore, A. L., N. H. Dawers, S. Gupta, R. Guidon, and T. Goldin (2004), Footwall topographic development during continental extension, *J. Geophys. Res.*, 109, 1–16, doi:10.1029/2003JF000115.
- Densmore, A. L., S. Gupta, P. A. Allen, and N. H. Dawers (2007), Transient landscapes at fault tips, *J. Geophys. Res.*, 112, 1–16, doi:10.1029/2006JF000560.
- Densmore, A. L., R. Hetzel, S. Ivy-ochs, W. C. Krugh, N. Dawers, P. Kubik, and R. Range (2009), Spatial variations in catchment-averaged denudation rates from normal fault footwalls, *Geology*, 37(12), 1139–1142, doi:10.1130/G30164A.1.
- Dixon, T. H., R. J. Stern, and I. M. Hussein (1987), Control of Red Sea Rift Geometry by Precambrian Structures, *Tectonics*, 6(5), 551–571, doi:10.1029/TC006i005p00551.
- Donelick, R. A., P. B. O’Sullivan, and R. A. Ketcham (2005), Apatite Fission-Track Analysis, *Rev. Mineral. Geochemistry*, 58(1), 49–94, doi:10.2138/rmg.2005.58.3.
- Dugda, M. T., A. A. Nyblade, J. Julia, C. A. Langston, C. J. Ammon, and S. Simiyu (2005), Crustal structure in Ethiopia and Kenya from receiver function analysis: Implications for rift development in eastern Africa, *J. Geophys. Res.*, 110, 1–15, doi:10.1029/2004JB003065.
- Ebinger, C. (2005), Continental break-up: The East African perspective, *Astron. Geophys.*, 46(2), 2–16.
- Ebinger, C., T. D. Bechtel, D. W. Forsyth, and C. O. Bowin (1989), Effective elastic plate thickness beneath the East African and Afar domes, *J. Geophys. Res.*, 94(B3), 2883–2901.
- Ebinger, C. J., and M. Casey (2001), Continental breakup in magmatic provinces: An Ethiopian example, *Geology*, 29(6), 527–530.
- Ebinger, C. J., and N. H. Sleep (1998), Cenozoic magnetism throughout east Africa resulting from impact of a single plume, *Nature*, 395, 788–791.
- Ebinger, C. J., T. Yemane, G. Woldegabriel, J. L. Aronson, and R. C. Walter (1993), Late Eocene-Recent volcanism and faulting in the southern main Ethiopian rift, *J. Geol. Soc. London.*, 150(1), 99–108, doi:10.1144/gsjgs.150.1.0099.
- Ebinger, C. J., T. Yemane, D. J. Harding, S. Tesfaye, S. Kelley, and D. C. Rex (2000a), Rift deflection, migration, and propagation: Linkage of the Ethiopian and Eastern rifts, Africa, *GSA Bull.*, 112(2), 163–176.
- Ebinger, C. J., T. Yemane, D. Harding, S. Tesfaye, S. Kelley, and D. C. Rex (2000b), Rift deflection, migration, and propagation: Linkage of the Ethiopian and Eastern rifts, Africa, *GSA Bull.*, 112(2), 163–176.
- Farley, K. A. (2000), Helium diffusion from apatite: General behavior as illustrated by Durango fluorapatite, *J. Geophys. Res.*, 105(B2), 2903–2914.
- Farley, K. A., and D. F. Stockli (2002), (U-Th)/He Dating of Phosphates: Apatite, Monazite, and Xenotime, *Rev. Mineral. Geochemistry*, 48(1), 559–577, doi:10.2138/rmg.2002.48.15.
- Farley, K. A., R. A. Wolf, and L. T. Silver (1996), The effects of long alpha-stopping distances on (U-Th)/He ages, *Geochim. Cosmochim. Acta*, 60(21), 4223–4229, doi:10.1016/S0016-7037(96)00193-7.
- Fernandes, R. M. S., B. A. C. Ambrosius, R. Noomen, L. Bastos, L. Combrinck, J. M. Miranda, and W. Spakman (2004), Angular velocities of Nubia and Somalia from continuous GPS data: implications on present-day relative kinematics, *Earth Planet. Sci.*

- Lett.*, 222, 197–208, doi:10.1016/j.epsl.2004.02.008.
- Filbrandt, J. M., P. D. Richard, and R. C. M. W. Franssen (1994), Growth and coalescence of faults: numerical simulations and sand-box experiments, in *TSG Special Meeting on Fault Populations, Edinburgh, UK, Extended Abstract, Tectonic Studies Group*.
- Fishwick, S. (2010), Surface wave tomography: Imaging of the lithosphere–asthenosphere boundary beneath central and southern Africa?, *Lithos*, 120, 63–73, doi:10.1016/j.lithos.2010.05.011.
- Fishwick, S., and I. D. Bastow (2011), Towards a better understanding of African topography: A review of passive-source seismic studies of the African crust and upper mantle, *Geol. Soc. London, Spec. Publ.*, 357, 343–371, doi:10.1144/SP357.19.
- Fitzgerald, P. G. (1992), The Transantarctic Mountains of Southern Victoria Land: The Application of Apatite Fission Track Analysis to a Rift Shoulder Uplift, *Tectonics*, 11(3), 634–662.
- Fitzgerald, P. G., S. L. Baldwin, L. E. Webb, and P. B. O’Sullivan (2006), Interpretation of (U–Th)/He single grain ages from slowly cooled crustal terranes: A case study from the Transantarctic Mountains of southern Victoria Land, *Chem. Geol.*, 225(1–2), 91–120, doi:10.1016/j.chemgeo.2005.09.001.
- Fleischer, R., P. Price, and R. Walker (1975), *Nuclear Tracks in Solids: principles and applications*, University of California Press, Berkeley.
- Flowers, R. M., and S. A. Kelley (2011), Interpreting data dispersion and “inverted” dates in apatite (U–Th)/He and fission-track datasets: An example from the US midcontinent, *Geochim. Cosmochim. Acta*, 75(18), 5169–5186, doi:10.1016/j.gca.2011.06.016.
- Flowers, R. M., D. L. Shuster, B. P. Wernicke, and K. A. Farley (2007), Radiation damage control on apatite (U–Th)/He dates from the Grand Canyon region, Colorado Plateau, *Geology*, 35(5), 447–450, doi:10.1130/G23471A.1.
- Flowers, R. M., R. A. Ketcham, D. L. Shuster, and K. A. Farley (2009), Apatite (U–Th)/He thermochronometry using a radiation damage accumulation and annealing model, *Geochim. Cosmochim. Acta*, 73(8), 2347–2365, doi:10.1016/j.gca.2009.01.015.
- Flowers, R. M., K. A. Farley, and R. A. Ketcham (2015), A reporting protocol for thermochronologic modeling illustrated with data from the Grand Canyon, *Earth Planet. Sci. Lett.*, 432, 425–435, doi:10.1016/j.epsl.2015.09.053.
- Forsyth, D. W. (1992), Finite extension and low-angle normal faulting, *Geology*, 20, 27–30, doi:10.1130/0091-7613(1992)020<0027.
- Fosdick, J. C., and J. P. Colgan (2008), Miocene extension in the East Range, Nevada: A two-stage history of normal faulting in the northern Basin and Range, *Geol. Soc. Am. Bull.*, 120(9), 1198–1213, doi:10.1130/B26201.1.
- Fossen, H., and A. Rotevatn (2016), Fault linkage and relay structures in extensional settings — A review, *Earth Sci. Rev.*, 154, 14–28, doi:10.1016/j.earscirev.2015.11.014.
- Foster, A. N., and J. A. Jackson (1998), Source parameters of large African earthquakes: implications for crustal rheology and regional kinematics, *Geophys. J. Int.*, 134, 422–448.
- Foster, D. A., and A. J. W. Gleadow (1992), The morphotectonic evolution of rift-margin mountains in central Kenya: Constraints from apatite fission-track thermochronology, *Earth Planet. Sci. Lett.*, 113(1–2), 157–171, doi:10.1016/0012-821X(92)90217-J.
- Foster, D. A., and A. J. W. Gleadow (1993), Episodic Denudation in East Africa: A Legacy of Intracontinental Tectonism, *Geophys. Res. Lett.*, 20(21), 2395–2398.
- Foster, D. A., and J. W. Gleadow (1996), Structural framework and denudation history of the flanks of the Kenya and Anza Rifts, East Africa, *Tectonics*, 15(2), 258–271, doi:10.1029/95TC02744.
- Gallagher, K. (1995), Evolving temperature histories from apatite fission-track data, *Earth*

- Planet. Sci. Lett.*, 136(3–4), 421–435.
- Gallagher, K. (2012), Transdimensional inverse thermal history modeling for quantitative thermochronology, *J. Geophys. Res. Solid Earth*, 117, doi:10.1029/2011JB008825.
- Gallagher, K., and R. Brown (1997), The onshore record of passive margin evolution, *J. Geol. Soc. London*, 154, 451–457, doi:10.1144/gsjgs.154.3.0451.
- Gallagher, K., and R. A. Ketcham (2018), Comment on “Thermal history modelling: HeFTy vs. QTQt” by Vermeesch and Tian, *Earth-Science Reviews* (2014), 139, 279–290, *Earth-Science Rev.*, 176, 387–394, doi:10.1016/j.earscirev.2017.11.001.
- Gautheron, C., L. Tassan-Got, J. Barbarand, and M. Pagel (2009), Effect of alpha-damage annealing on apatite (U-Th)/He thermochronology, *Chem. Geol.*, 266, 157–170, doi:10.1016/j.chemgeo.2009.06.001.
- Gautheron, C., J. Barbarand, R. A. Ketcham, L. Tassan-Got, P. van der Beek, M. Pagel, R. Pinna-Jamme, F. Couffignal, and M. Fialin (2013), Chemical influence on α -recoil damage annealing in apatite: Implications for (U-Th)/He dating, *Chem. Geol.*, 351, 257–267, doi:10.1016/j.chemgeo.2013.05.027.
- Gawthorpe, R. L., and M. R. Leeder (2000), Tectono-sedimentary evolution of active extensional basins, *Basin Res.*, 12, 195–218.
- Gawthorpe, R. L., A. J. Fraser, and R. E. L. I. Collier (1994), Sequence stratigraphy in active extensional basins: implications for the interpretation of ancient basin-fills, *Mar. Pet. Geol.*, 11(6), 642–658.
- Gawthorpe, R. L., I. Sharp, J. R. Underhill, and S. Gupta (1997), Linked sequence stratigraphic and structural evolution of propagating normal faults, *Geology*, 25(9), 795–798, doi:10.1144/GSL.SP.2000.172.01.09.
- Gawthorpe, R. L., C. A. Jackson, M. J. Young, I. R. Sharp, A. R. Moustafa, and C. W. Leppard (2003), Normal fault growth, displacement localisation and the evolution of normal fault populations: the Hammam Faraun fault block, Suez rift, Egypt, *J. Struct. Geol.*, 25, 883–895.
- George, R., N. Rogers, and S. Kelley (1998), Earliest magnetism in Ethiopia: Evidence for two mantle plumes in one flood basalt province, *Geology*, 26(10), 923–926.
- George, R. M., and N. W. Rogers (2002), Plume dynamics beneath the African plate inferred from the geochemistry of the Tertiary basalts of southern Ethiopia, *Contrib. to Mineral. Petrol.*, 144, 286–304, doi:10.1007/s00410-002-0396-z.
- Giba, M., J. J. Walsh, and A. Nicol (2012), Segmentation and growth of an obliquely reactivated normal fault, *J. Struct. Geol.*, 39, 253–267, doi:10.1016/j.jsg.2012.01.004.
- Gleadow, A., M. Harrison, B. Kohn, R. Lugo-zazueta, and D. Phillips (2015), The Fish Canyon Tuff: A new look at an old low-temperature thermochronology standard, *Earth Planet. Sci. Lett.*, 424, 95–108, doi:10.1016/j.epsl.2015.05.003.
- Gleadow, A. J. W., and I. R. Duddy (1981), A Natural Long-Term Track Annealing Experiment for Apatite, *Nucl. Tracks*, 5(1–2), 169–174.
- Gorynski, K. E., D. F. Stockli, and J. D. Walker (2013), Thermochronometrically constrained anatomy and evolution of a Miocene extensional accommodation zone and tilt domain boundary: The southern Wassuk Range, Nevada, *Tectonics*, 32, 516–539, doi:10.1002/tect.20044.
- Green, P. F., I. R. Duddy, A. J. W. Gleadow, and J. F. Lovering (1989), Apatite Fission-Track Analysis as a Paleotemperature Indicator for Hydrocarbon Exploration, in *Thermal History of Sedimentary Basins, Methods and Case Histories*, edited by N. D. Naeser and T. H. McCulloh, pp. 181–196, Springer-Verlag, New York.
- Guenther, W. R., P. W. Reiners, R. A. Ketcham, L. Nasdala, and G. Giester (2013), Helium diffusion in natural zircon: radiation damage, anisotropy, and the interpretation of zircon (U-TH)/He thermochronology, *Am. J. Sci.*, 313(3), 145–198, doi:10.2475/03.2013.01.

- Guiraud, R., and W. Bosworth (1997), Senonian basin inversion and rejuvenation of rifting in Africa and Arabia : synthesis and implications to plate-scale tectonics, *Tectonophysics*, 282, 39–82.
- Guiraud, R., and J.-C. Maurin (1992), Early Cretaceous rifts of Western and Central Africa: an overview, *Tectonophysics*, 213, 153–168.
- Gupta, S., P. A. Cowie, N. H. Dawers, and J. R. Underhill (1998), A mechanism to explain rift-basin subsidence and stratigraphic patterns through fault-array evolution, *Geology*, 26(7), 595–598, doi:10.1130/0091-7613(1998)026<0595.
- Harbor, D. J. (1997), Landscape evolution at the margin of the Basin and Range, *Geology*, 25(12), 1111–1114.
- Hayward, N. J., and C. J. Ebinger (1996), Variations in the along-axis segmentation of the Afar Rift system, *Tectonics*, 15(2), 244–257.
- Hendrie, D. B., N. J. Kusznir, C. K. Morley, and C. J. Ebinger (1994), Cenozoic extension in northern Kenya: a quantitative model of rift basin development in the Turkana region, *Tectonophysics*, 236(1–4), 409–438, doi:10.1016/0040-1951(94)90187-2.
- Hetzel, R., and M. R. Strecker (1994), Late Mozambique Belt structures in western Kenya and their influence on the evolution of the Cenozoic Kenya Rift, *J. Struct. Geol.*, 16(2), 189–201, doi:10.1016/0191-8141(94)90104-X.
- Hofmann, C., V. Courtillot, G. Féraud, P. Rochette, G. Yirgu, E. Ketefo, and R. Pik (1997), Timing of the Ethiopian flood basalt event and implications of Plume birth and global change, *Nature*, 389, 838–841, doi:10.1038/39853.
- Ibrahim, A. E., C. J. Ebinger, and J. D. Fairhead (1991), Interpretation of the Central African Rift system in Sudan based on new gravity and aeromagnetic data, *Eos (Washington, DC)*, 462.
- Idleman, B. D., P. K. Zeitler, and K. T. McDannell (2017), Characterization of helium release from apatite by continuous ramped heating, *Chem. Geol.*, doi:10.1016/j.chemgeo.2017.11.019.
- Jackson, C. A., and A. Rotevatn (2013), 3D seismic analysis of the structure and evolution of a salt-influenced normal fault zone: A test of competing fault growth models, *J. Struct. Geol.*, 54, 215–234, doi:10.1016/j.jsg.2013.06.012.
- Jackson, J., and D. McKenzie (1983), The geometrical evolution of normal fault systems, *J. Struct. Geol.*, 5(5), 471–482.
- Keir, D., C. J. Ebinger, G. W. Stuart, E. Daly, and A. Ayele (2006), Strain accommodation by magmatism and faulting as rifting proceeds to breakup: Seismicity of the northern Ethiopian rift, *J. Geophys. Res.*, 111, 1–17, doi:10.1029/2005JB003748.
- Keranen, K., and S. L. Klemperer (2008), Discontinuous and diachronous evolution of the Main Ethiopian Rift: Implications for development of continental rifts, *Earth Planet. Sci. Lett.*, 265, 96–111, doi:10.1016/j.epsl.2007.09.038.
- Keranen, K. M., S. L. Klemperer, J. Julia, J. F. Lawrence, and A. A. Nyblade (2009), Low lower crustal velocity across Ethiopia: Is the Main Ethiopian Rift a narrow rift in a hot craton?, *Geochemistry, Geophys. Geosystems*, 10, 1–21, doi:10.1029/2008GC002293.
- Ketcham, R. A. (2005), Forward and Inverse Modeling of Low- Temperature Thermochronometry Data, *Rev. Mineral. Geochemistry*, 58, 275–314, doi:10.2138/rmg.2005.58.11.
- Ketcham, R. A., A. Carter, R. A. Donelick, J. Barbarand, and A. J. Hurford (2007a), Improved measurement of fission-track annealing in apatite using c-axis projection, *Am. Mineral.*, 92(5–6), 789–798, doi:10.2138/am.2007.2280.
- Ketcham, R. A., A. Carter, R. A. Donelick, J. Barbarand, and A. J. Hurford (2007b), Improved modeling of fission-track annealing in apatite, *Am. Mineral.*, 92, 799–810, doi:10.2138/am.2007.2281.

- Kim, Y., and D. J. Sanderson (2005), The relationship between displacement and length of faults: a review, *Earth-Science Rev.*, 68, 317–334, doi:10.1016/j.earscirev.2004.06.003.
- Kinabo, B. D., E. A. Atekwana, J. P. Hogan, M. P. Modisi, D. D. Wheaton, and A. B. Kampunzu (2007), Early structural development of the Okavango rift zone, NW Botswana, *J. African Earth Sci.*, 48, 125–136, doi:10.1016/j.jafrearsci.2007.02.005.
- Kohn, B. P., A. J. W. Gleadow, R. W. Brown, K. Gallagher, M. Lorencak, and W. P. Noble (2005), Visualizing Thermotectonic and Denudational Histories Using Apatite Fission Track Thermochronology, *Rev. Mineral. Geochemistry*, 58, 527–565, doi:10.2138/rmg.2005.58.20.
- Kohn, B. P., M. Lorencak, A. J. W. Gleadow, F. Kohlmann, A. Raza, K. G. Osadetz, and P. Sorjonen-ward (2009), A reappraisal of low-temperature thermochronology of the eastern Fennoscandia Shield and radiation-enhanced apatite fission-track annealing, in *Thermochronological Methods: From Palaeotemperature Constraints to Landscape Evolution Models*, edited by F. Lisker, B. Ventura, and U. A. Glasmacher, pp. 193–216, Geological Society of London, London.
- Kreuser, T. (1995), Rift to drift evolution in Permian-Jurassic basins of East Africa, *Geol. Soc. London, Spec. Publ.*, 80(1), 297–315.
- Kröner, a, and R. J. Stern (2005), Pan-African Orogeny, *Encycl. Of Geol.*, 1(2004), 1–12, doi:10.1016/S1342-937X(05)70162-3.
- Krugh, W. (2008), Low-temperature thermochronologic constraints on fault array evolution and patterns of range-scale denudation, ETH Zurich.
- Kusznir, N. J., and R. G. Park (1987), The extensional strength of the continental lithosphere: its dependence on geothermal gradient, and crustal composition and thickness, *Geol. Soc. Spec. Publ.*, (28), 35–52.
- Laslett, G. M., P. F. Green, I. R. Duddy, and A. J. W. Gleadow (1987), Thermal annealing of fission tracks in apatite 2 . A quantitative analysis, *Chem. Geol. (Isotope Geosci. Sect.)*, 65, 1–13, doi:10.1016/0168-9622(87)90057-1.
- Levitte, D., J. Columba, and P. Mohr (1974), Reconnaissance Geology of the Amaro Horst, Southern Ethiopian Rift, *Geol. Soc. Am. Bull.*, 85, 417–422.
- Maguire, P. K. H. et al. (2006), Crustal structure of the northern Main Ethiopian Rift from the EAGLE controlled-source survey; a snapshot of incipient lithospheric break-up, *Geol. Soc. London Spec. Publ.*, 259, 269–291.
- Marrett, R., and R. W. Allmendinger (1991), Estimates of strain due to brittle faulting: sampling of fault populations, *J. Struct. Geol.*, 13(6), 735–738.
- Mbongo Djimbi, D., C. Gautheron, J. Roques, L. Tassan-Got, C. Gerin, and E. Simoni (2015), Impact of apatite chemical composition on (U-Th)/He thermochronometry: An atomistic point of view, *Geochim. Cosmochim. Acta*, 167, doi:10.1016/j.gca.2015.06.017.
- McConnell, R. B. (1972), Geological Development of the Rift System of Eastern Africa, *Geol. Soc. Am. Bull.*, 83, 2549–2572.
- Mcleod, A. E., N. H. Dawers, and J. R. Underhill (2000), The propagation and linkage of normal faults: Insights from the Strathspey-Brent-Statfjord fault array, northern North Sea, *Basin Res.*, 12(3–4), 263–284.
- Meesters, A. G. C. A., and T. J. Dunai (2002), Solving the production – diffusion equation for finite diffusion domains of various shapes: Part II. Application to cases with α -ejection and nonhomogenous distribution of the source, *Chem. Geol.*, 186, 347–363, doi:10.1016/S0009-2541(01)00423-5.
- Mengesha, T., C. Tadiwos, and H. Workineh (1996), Explanation of the geological map of Ethiopia,
- Mickus, K., K. Tadesse, G. R. Keller, and B. Oluma (2007), Gravity analysis of the main

- Ethiopian rift, *J. African Earth Sci.*, 48, 59–69, doi:10.1016/j.jafrearsci.2007.02.008.
- Mohr, P. (1983), Ethiopian flood basalt province, *Nature*, 303, 577–584.
- Mohr, P. A. (1962), *The geology of Ethiopia*, University College Addis Ababa Press, Addis Ababa.
- Moore Jr, J. M., and A. Davidson (1978), Rift Structure in Southern Ethiopia, *Tectonophysics*, 46, 159–173.
- Morewood, N. C., and G. P. Roberts (1999), Lateral propagation of the surface trace of the South Alkyonides normal fault segment, central Greece: its impact on models of fault growth and displacement-length relationships, *J. Struct. Geol.*, 21, 635–652.
- Morley, C. K. (1994), Interaction of deep and shallow processes in the evolution of the Kenya rift, *Tectonophysics*, 236, 81–91.
- Morley, C. K. (1999a), Basin Evolution Trends in East Africa, in *Geoscience of Rift Systems-Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 131–150.
- Morley, C. K. (1999b), Influence of Preexisting Fabrics on Rift Structure, in *Geoscience of Rift Systems-Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 151–160.
- Morley, C. K. (1999c), Patterns of Displacement Along Large Normal Faults: Implications for Basin Evolution and Fault Propagation, Based on Examples from East Africa, *Am. Assoc. Pet. Geol. Bull.*, 83(4), 613–634.
- Morley, C. K. (2002a), Evolution of large normal faults: Evidence from seismic reflection data, *Am. Assoc. Pet. Geol. Bull.*, 86(6), 961–978.
- Morley, C. K. (2002b), Tectonic settings of continental extensional provinces and their impact on sedimentation and hydrocarbon prospectivity, *SEPM Spec. Publ.*, 73, 22–56.
- Morley, C. K. (2016), The impact of multiple extension events, stress rotation and inherited fabrics on normal fault geometries and evolution in the Cenozoic rift basins of Thailand, *Geol. Soc. London, Spec. Publ.*, 439, doi:10.1144/SP439.3.
- Morley, C. K., and W. A. Wescott (1999), Sedimentary Environments and Geometry of Sedimentary Bodies Determined from Subsurface Studies in East Africa, in *Geoscience of Rift Systems-Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 211–231.
- Morley, C. K., R. A. Nelson, T. L. Patton, and S. G. Munn (1990), Transfer zones in the East African Rift System and their relevance to hydrocarbon exploration in rifts, *Am. Assoc. Pet. Geol. Bull.*, 74(8), 1234–1253.
- Morley, C. K., W. a. Wescott, D. M. Stone, R. M. Harper, S. T. Wigger, and F. M. Karanja (1992), Tectonic evolution of the northern Kenyan Rift, *J. Geol. Soc. London.*, 149(3), 333–348, doi:10.1144/gsjgs.149.3.0333.
- Morley, C. K., W. Bosworth, R. A. Day, R. Lauck, R. Bosher, D. M. Stone, S. T. Wigger, W. A. Wescott, D. Haun, and N. Bassett (1999a), Geology and Geophysics of the Anza Graben, in *Geoscience of Rift Systems-Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 67–90.
- Morley, C. K., D. M. Stone, R. M. Harper, and S. T. Wigger (1999b), Geology and Geophysics of the Western Turkana Basins, Kenya, in *Geoscience of Rift Systems—Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 19–54.
- Morley, C. K., D. K. Ngenoh, and J. K. Ego (1999c), Introduction to the East African Rift System, in *Geoscience of Rift Systems-Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 1–18.
- Morley, C. K., N. Wonganan, A. Kornawan, W. Phoosongsee, C. Haranya, and S. Pongwapee (2004), Activation of rift oblique and rift parallel pre-existing fabrics during

- extension and their effect on deformation style: examples from the rifts of Thailand, *J. Struct. Geol.*, 26(10), 1803–1829, doi:10.1016/j.jsg.2004.02.014.
- Mortimer, E., L. A. Kirstein, F. M. Stuart, and M. R. Strecker (2016), Spatio-temporal trends in normal-fault segmentation recorded by low-temperature thermochronology: Livingstone fault scarp, Malawi Rift, East African Rift System, *Earth Planet. Sci. Lett.*, 455, 62–72, doi:10.1016/j.epsl.2016.08.040.
- Noble, W. P., D. A. Foster, and A. J. W. Gleadow (1997), The post-Pan-African thermal and extensional history of crystalline basement rocks in eastern Tanzania, *Tectonophysics*, 275(4), 331–350, doi:10.1016/S0040-1951(97)00026-7.
- Nyblade, A. A. (1990), Terrestrial heat flow in east and southern Africa, *J. Geophys. Res. Solid Earth*, 95(B11), 17371–17384.
- Peacock, D., and D. J. Sanderson (1994), Geometry and Development of Relay Ramps in Normal Fault Systems, *Am. Assoc. Pet. Geol. Bull.*, 78(2), 147–165, doi:10.1306/BDF9046-1718-11D7-8645000102C1865D.
- Philippon, M., G. Corti, F. Sani, M. Bonini, M. Balestrieri, P. Molin, E. Willingshofer, D. Sokoutis, and S. Cloetingh (2014), Evolution, distribution, and characteristics of rifting in southern Ethiopia, *Tectonics*, 33, 1–24, doi:10.1002/2013TC003430.
- Pik, R., B. Marty, J. Carignan, G. Yirgu, and T. Ayalew (2008), Timing of East African Rift development in southern Ethiopia : Implication for mantle plume activity and evolution of topography, *Geology*, 36(2), 167–170, doi:10.1130/G24233A.1.
- Reeves, C. V., J. P. Teasdale, and E. S. Mahanjane (2016), Insight into the Eastern Margin of Africa from a new tectonic model of the Indian Ocean, *Geol. Soc. London Spec. Publ.*, 431, doi:10.1144/SP431.12.
- Reeves, C. V., F. M. Karanja, and I. N. MacLeod (1987), Geophysical evidence for a failed Jurassic rift and triple junction in Kenya, *Earth Planet. Sci. Lett.*, 81(2–3), 299–311, doi:10.1016/0012-821X(87)90166-X.
- Reiners, P. W., and M. T. Brandon (2006), Using Thermochronology to Understand Orogenic Erosion, *Annu. Rev. Earth Planet. Sci.*, 34, 419–466, doi:10.1146/annurev.earth.34.031405.125202.
- Reiners, P. W., and K. A. Farley (2001), Influence of crystal size on apatite (U-Th)/ He thermochronology: an example from the Bighorn Mountains, Wyoming, *Earth Planet. Sci. Lett.*, 188, 413–420.
- Reiners, P. W., T. A. Ehlers, and P. K. Zeitler (2005), Past, present, and future of thermochronology, *Revi.*, 58, 1–18, doi:10.2138/rmg.2005.58.1.
- Rogers, N. W. (2006), Basaltic magmatism and the geodynamics of the East African Rift System, *Geol. Soc. London, Spec. Publ.*, 259, 77–93, doi:10.1144/GSL.SP.2006.259.01.08.
- Rousseeuw, P. J., and C. Croux (1993), Alternatives to the Median Absolute Deviation, *J. Am. Stat. Assoc.*, 88(424), 1273–1283.
- Royer, J.-Y., R. G. Gordon, and B. C. Horner-Johnson (2006), Motion of Nubia relative to Antarctica since 11 Ma: Implications for Nubia-Somalia, Pacific–North America, and India-Eurasia motion, *Geology*, 34(6), 501–504, doi:10.1130/G22463.1.
- Schlische, R. W. (1991), Half-graben basin filling models: New constraints on continental extensional basin development, *Basin Res.*, 3, 123–141.
- Schlische, R. W., and M. H. Anders (1996), Stratigraphic effects and tectonic implications of the growth of normal faults and extensional basins, *Geol. Soc. Am. Spec. Pap.*, 303, 183–203.
- Schöpfer, M. P. J., C. Childs, and J. J. Walsh (2006), Localisation of normal faults in multilayer sequences, *J. Struct. Geol.*, 28(5), 816–833.
- Schull, T. J. (1988), Rift Basins of Interior Sudan: Petroleum Exploration and Discovery, ,

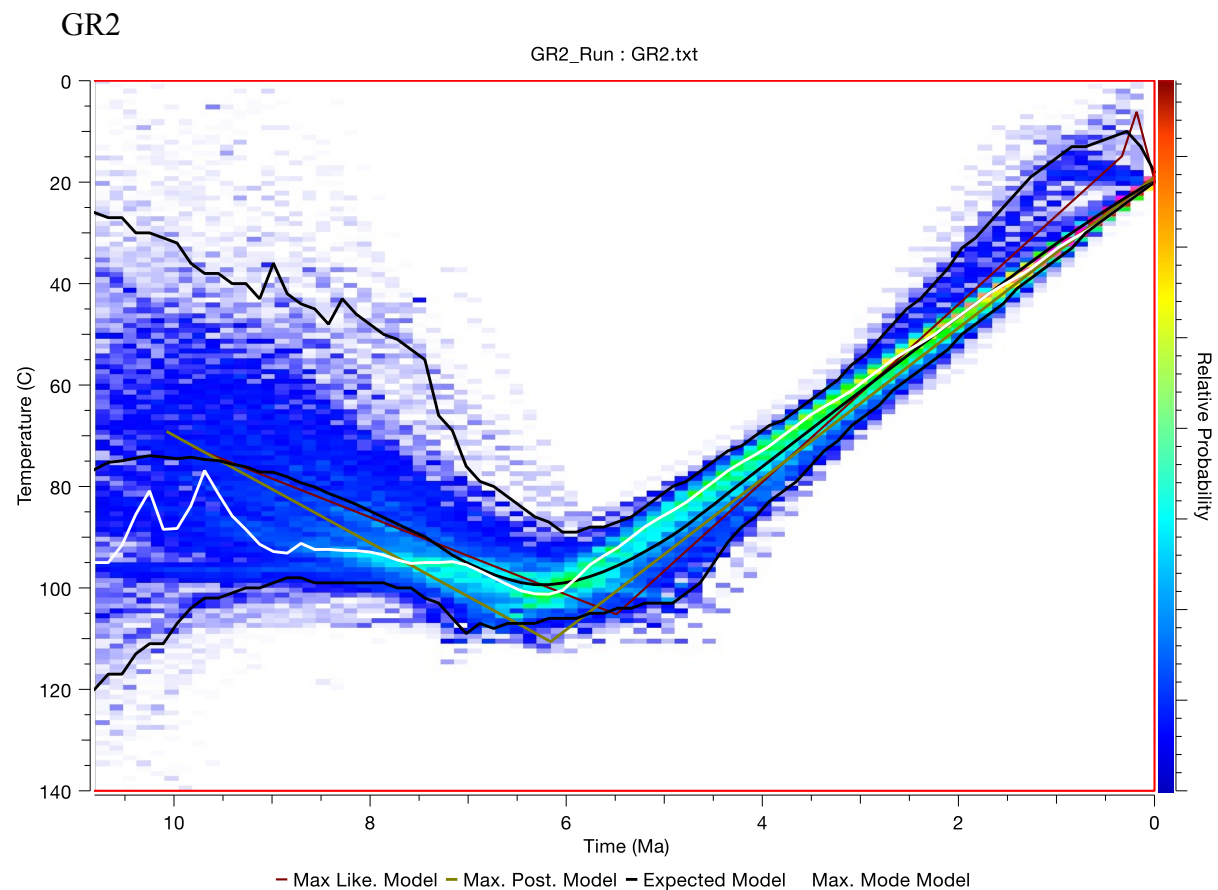
- 72(10), 1128–1142.
- Seiler, C., A. J. W. Gleadow, J. M. Fletcher, and B. P. Kohn (2009), Thermal evolution of a sheared continental margin: Insights from the Ballenas transform in Baja California, Mexico, *Earth Planet. Sci. Lett.*, 285(1–2), 61–74, doi:10.1016/j.epsl.2009.05.043.
- Seiler, C., J. M. Fletcher, B. P. Kohn, A. J. W. Gleadow, and A. Raza (2011), Low - temperature thermochronology of northern Baja California , Mexico : Decoupled slip - exhumation gradients and delayed onset of oblique rifting across the Gulf of California, *Tectonics*, 30(1), doi:10.1029/2009TC002649.
- Sharp, I. R., R. L. Gawthorpe, B. Armstrong, and J. R. Underhill (2000), Propagation history and passive rotation of mesoscale normal faults: implications for synrift stratigraphic development, 2000, 12, 285–305.
- Shuster, D. L., R. M. Flowers, and K. A. Farley (2006), The influence of natural radiation damage on helium diffusion kinetics in apatite, *Earth Planet. Sci. Lett.*, 249(3–4), 148–161, doi:10.1016/j.epsl.2006.07.028.
- Simmons, N. A., A. M. Forte, and S. P. Grand (2007), Thermochemical structure and dynamics of the African superplume, *Geophys. Res. Lett.*, 34, 1–5, doi:10.1029/2006GL028009.
- Smith, M., and P. Mosley (1993), Crustal Heterogeneity and Basement Influence on the Development of the Kenya Rift, East Africa, *Tectonics*, 12(2), 591–606.
- Spiegel, C., B. P. Kohn, D. X. Belton, and A. J. W. Gleadow (2007), Morphotectonic evolution of the central Kenya rift flanks: Implications for late Cenozoic environmental change in East Africa, *Geology*, 35(5), 427–430, doi:10.1130/G23108A.1.
- Steen, Ø., and A. Andresen (1999), Effects of lithology on geometry and scaling of small faults in Triassic sandstones, East Greenland, *J. Struct. Geol.*, 21, 1351–1368.
- Stern, R. J. (2002), Crustal evolution in the East African Orogen: a neodymium isotopic perspective, *J. African Earth Sci.*, 34, 109–117.
- Stern, R. J., K. C. Nielsen, E. Best, M. Sultan, R. E. Arvidson, and A. Kröner (1990), Orientation of late Precambrian sutures in the Arabian-Nubian Shield, *Geology*, 18, 1103–1106, doi:10.1130/0091-7613(1990)018<1103.
- Stockli, D. F. (2005), Application of Low-Temperature Thermochronometry to Extensional Tectonic Settings, *Rev. Mineral. Geochemistry*, 58, 411–448, doi:10.2138/rmg.2005.58.16.
- Stockli, D. F., K. A. Farley, and T. A. Dumitru (2000), Calibration of the apatite (U-Th)/He thermochronometer on an exhumed fault block, White Mountains, California, *Geology*, 28(11), 983–986.
- Storti, F., M. L. Balestrieri, F. Balsamo, and F. Rossetti (2008), Structural and thermochronological constraints to the evolution of the West Antarctic Rift System in central Victoria Land, *Tectonics*, 27, 1–21, doi:10.1029/2006TC002066.
- Stuart, G. W., I. D. Bastow, and C. J. Ebinger (2006), Crustal structure of the northern main Ethiopian rift from receiver function studies, *Geol. Soc. London Spec. Publ.*, 259, 253–267, doi:10.1144/GSL.SP.2006.259.01.20.
- Sullivan, P. B. O., and R. R. Parrish (1995), The importance of apatite composition and single-grain ages when interpreting fission track data from plutonic rocks: a case study from the Coast Ranges, British Columbia, *Earth Planet. Sci. Lett.*, 132, 213–224.
- Sykes, L. R. (1978), Intraplate Seismicity, Reactivation of Preexisting Zones of Weakness, Alkaline Magmatism, and Other Tectonism Postdating Continental Fragmentation, *Rev. Geophys. Sp. Phys.*, 16(4), 621–688.
- Torres Acosta, V., A. Bande, E. R. Sobel, M. Parra, T. F. Schildgen, F. Stuart, and M. R. Strecker (2015), Cenozoic extension in the Kenya Rift from low-temperature thermochronology: Links to diachronous spatiotemporal evolution of rifting in East

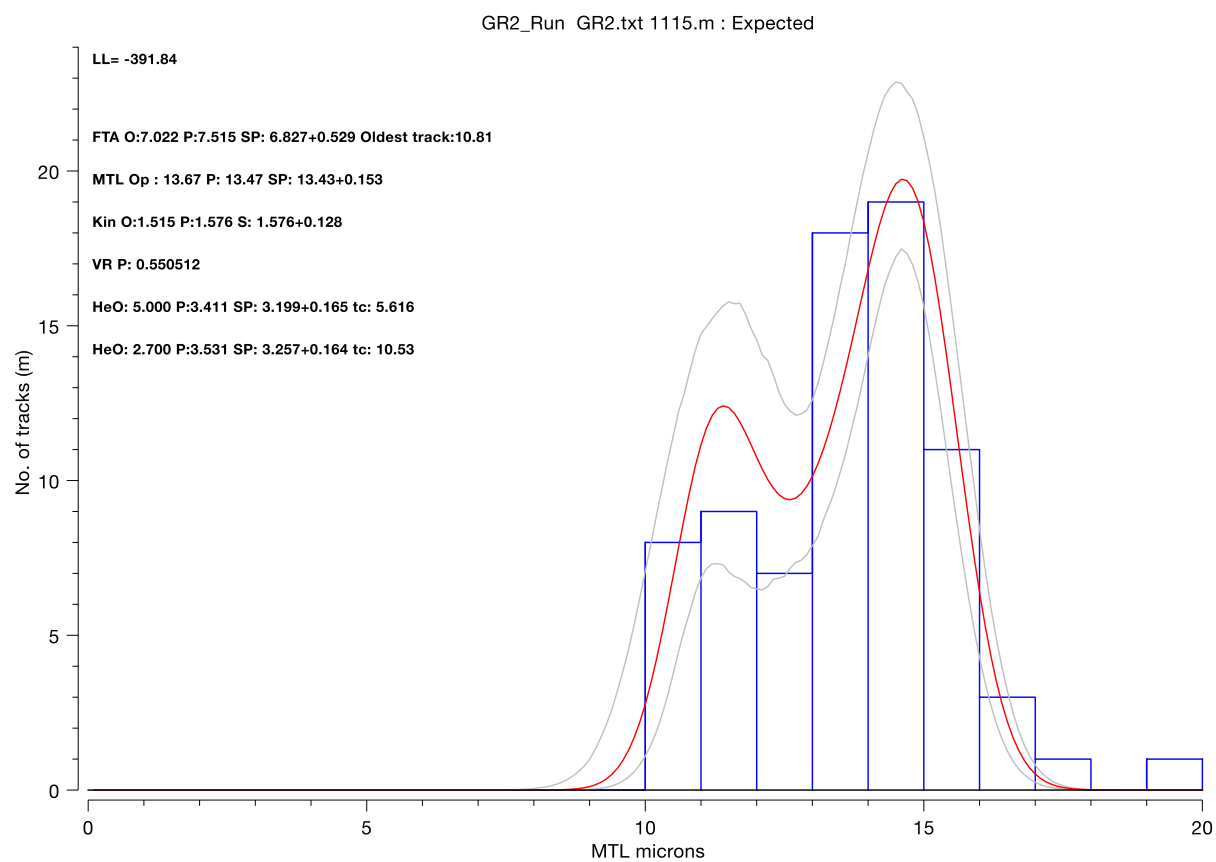
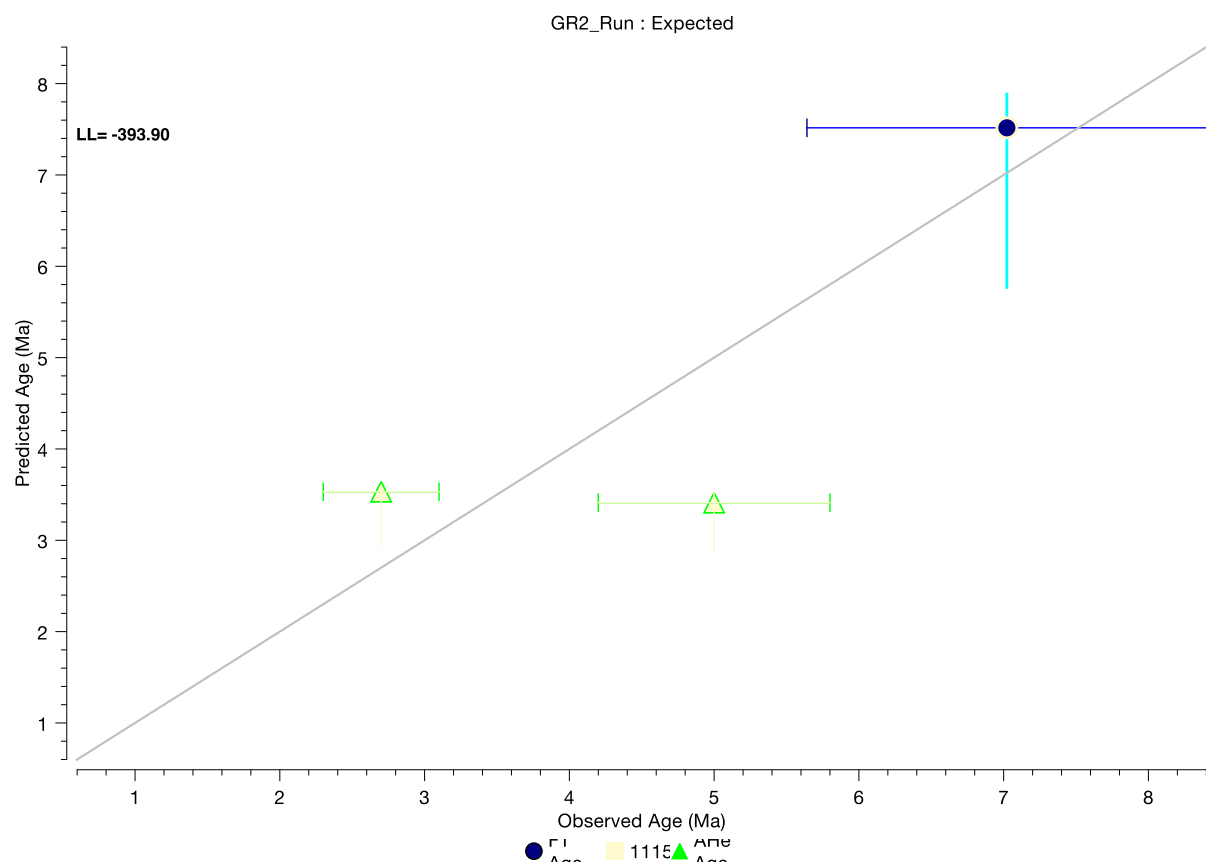
- Africa, *Tectonics*, 34(12), 2367–2386, doi:10.1002/2015TC003949.
- Tripathy-Lang, A., M. Fox, and D. L. Shuster (2015), Zircon 4He/3He thermochronometry, *Geochim. Cosmochim. Acta*, 166, 1–14, doi:10.1016/j.gca.2015.05.027.
- Trudgill, B., and J. Cartwright (1994), Relay-Ramp forms and normal-fault linkages, Canyonlands National Park, Utah, *Geol. Soc. Am. Bull.*, 106, 1143–1157.
- Vaucher, A., G. Barruol, and A. Tommasi (1997), Why do continents break-up parallel to ancient orogenic belts?, *Terra Nov.*, 9, 62–66.
- Vaucher, A., A. Tommasi, and G. Barruol (1998), Rheological heterogeneity, mechanical anisotropy and deformation of the continental lithosphere, *Tectonophysics*, 296, 61–86.
- Vermeesch, P., and Y. Tian (2014), Thermal history modelling: HeFTy vs. QTQt, *Earth-Science Rev.*, 139, 279–290, doi:10.1016/j.earscirev.2014.09.010.
- Versfelt, J., and B. R. Rosendahl (1989), Relationships between pre-rift structure and rift architecture in Lakes Tanganyika and Malawi, East Africa, *Nature*, 337, 354–357.
- Vétel, W., and B. Le Gall (2006), Dynamics of prolonged continental extension in magmatic rifts: the Turkana Rift case study (North Kenya), *Geol. Soc. London Spec. Publ.*, 259, 209–233.
- Vétel, W., B. Le Gall, and J. J. Walsh (2005), Geometry and growth of an inner rift fault pattern: the Kino Sogo Fault Belt, Turkana Rift (North Kenya), *J. Struct. Geol.*, 27, 2204–2222, doi:10.1016/j.jsg.2005.07.003.
- Wagner, G., and P. Van den Haute (1992), *Fission-Track Dating*, Kluwer Academic Publishers, Dordrecht.
- Walsh, J. J., and J. Watterson (1988), Analysis of the relationship between displacements and dimensions of faults, *J. Struct. Geol.*, 10(3), 239–247.
- Walsh, J. J., A. Nicol, and C. Childs (2002), An alternative model for the growth of faults, *J. Struct. Geol.*, 24, 1669–1675.
- Walsh, J. J., W. R. Bailey, C. Childs, A. Nicol, and C. G. Bonson (2003), Formation of segmented normal faults: a 3-D perspective, *J. Struct. Geol.*, 25, 1251–1262.
- Watterson, J. (1986), Fault Dimensions, Displacements and Growth, *Pure Appl. Geophys.*, 124(1/2), 365–373.
- Weeraratne, D. S., D. W. Forsyth, K. M. Fischer, and A. A. Nyblade (2003), Evidence for an upper mantle plume beneath the Tanzanian craton from Rayleigh wave tomography, *J. Geophys. Res.*, 108(B9), 1–14, doi:10.1029/2002JB002273.
- Wescott, W. A., S. T. Wigger, D. M. Stone, and C. K. Morley (1999), Geology and Geophysics of the Lotikipi Plain, *Geosci. Rift Syst. East Africa*, (44), 55–66.
- Wijk, J. W. van (2005), Role of weak zone orientation in continental lithosphere extension, *Geophys. Res. Lett.*, 32, 1–4, doi:10.1029/2004GL022192.
- Wildman, M., R. Brown, R. Beucher, C. Persano, F. Stuart, K. Gallagher, J. Schwanethal, and A. Carter (2016), The chronology and tectonic style of landscape evolution along the elevated Atlantic continental margin of South Africa resolved by joint apatite fission track and (U-Th-Sm)/He thermochronology, *Tectonics*, 35(3), 511–545, doi:10.1002/2015TC004042.
- Wilkins, S. J., and M. R. Gross (2002), Normal fault growth in layered rocks at Split Mountain, Utah: influence of mechanical stratigraphy on dip linkage, fault restriction and fault scaling, *J. Struct. Geol.*, 24, 1413–1429.
- Woldegabriel, G., and J. Aronson (1987), Chew Bahir Rift: A “failed” rift in southern Ethiopia, *Geology*, 15, 430–433.
- Woldegabriel, G., T. Yemmar, G. Suw, T. Whrre, and B. Asfaw (1991), Age of volcanism and rifting in the Burji-Soyoma area, Amaro Horst, southern Main Ethiopian Rift: geo- and biochronologic data, *J. African Earth Sci.*, 13(3/4), 437–447.
- WoldeGabriel, G., N. Hessen, and T. Yemane (1997), Agere Maryam Sheet (NB37-10),

- Geological Survey of Ethiopia, 1:250,000 scale,
- Wolfe, M. R., and D. F. Stockli (2010), Zircon (U-Th)/He thermochronometry in the KTB drill hole, Germany, and its implications for bulk He diffusion kinetics in zircon, *Earth Planet. Sci. Lett.*, 295(1–2), 69–82, doi:10.1016/j.epsl.2010.03.025.
- Wolfenden, E., C. Ebinger, G. Yirgu, A. Deino, and D. Ayalew (2004), Evolution of the northern Main Ethiopian rift: birth of a triple junction, *Earth Planet. Sci. Lett.*, 224(1–2), 213–228, doi:10.1016/j.epsl.2004.04.022.
- Yemane, T., and T. Yohunie (1987), *Geological Report on subsheets A, B, and C (Agere Mariam, NG-37-10)*, Addis Ababa.
- Zanettin, B., E. Justin-Visentin, M. Nicoletti, and C. Petrucciani (1978), The evolution of the Chenchä escarpment and the Ganjuli graben (lake Abaya) in the southern Ethiopian rift, *Neues Jahrb. für Geol. und Paläontologie*, 8, 473–490.
- Zeitler, P. K., E. Enkelmann, J. B. Thomas, E. B. Watson, L. D. Ancuta, and B. D. Idleman (2017), Solubility and trapping of helium in apatite, *Geochim. Cosmochim. Acta*, 209, 1–8, doi:10.1016/j.gca.2017.03.041.
- Zeyen, H., F. Volker, V. Wehrle, K. Fuchs, S. V. Sobolev, and R. Altherr (1997), Styles of continental rifting: crust-mantle detachment and mantle plumes, *Tectonophysics*, 278(1–4), 329–352, doi:10.1016/S0040-1951(97)00111-X.
- Zoback, M. Lou, and M. D. Zoback (1980), Faulting Patterns in North-Central Nevada and Strength of the Crust, *J. Geophys. Res.*, 85(B1), 275–284.

Supplementary Data A: Individual sample thermal history models without t-T constraints

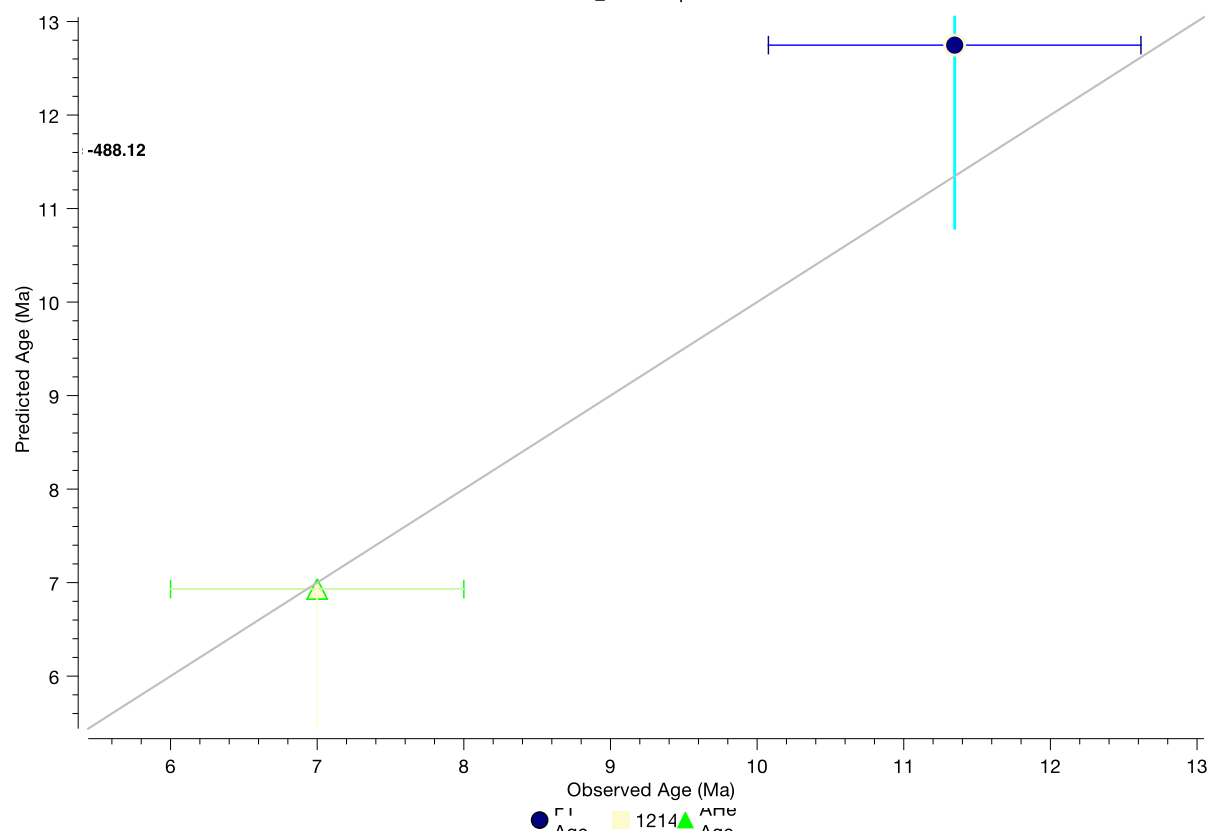
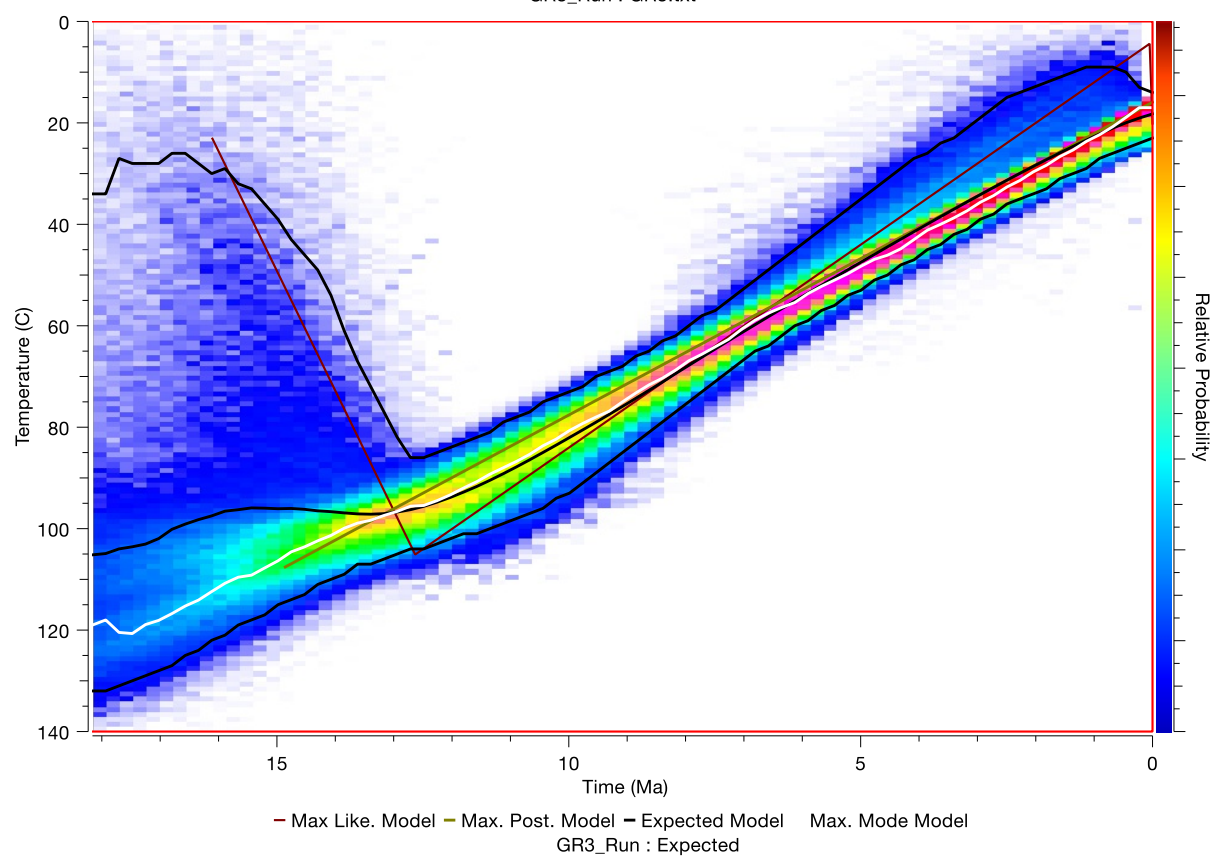
Thermal history models displayed below were generated in the software *QTQt* using AFT and AHe data (when available), following the procedure outlined in section 4.6. Only samples with a statistically robust number of confined track length measurements (> 60) were modelled. A sole time-temperature constraint was applied, requiring samples to be at near-surface temperatures (20 ± 10 °C) by the present-day. Plots of observed versus predicted AFT ages and confined track distributions are displayed below each thermal history model. All plots were generated by *QTQt*. For descriptions of the various plots, see the caption for Figure 4.7.

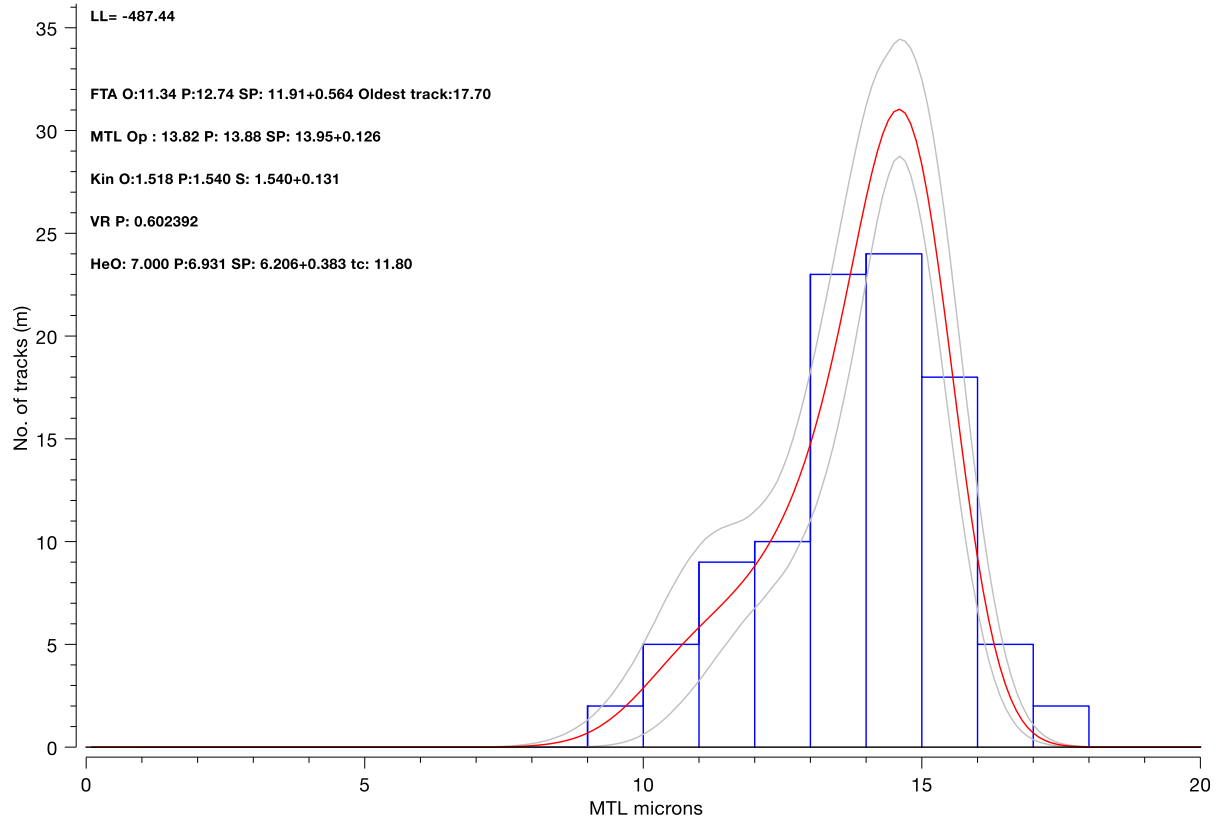




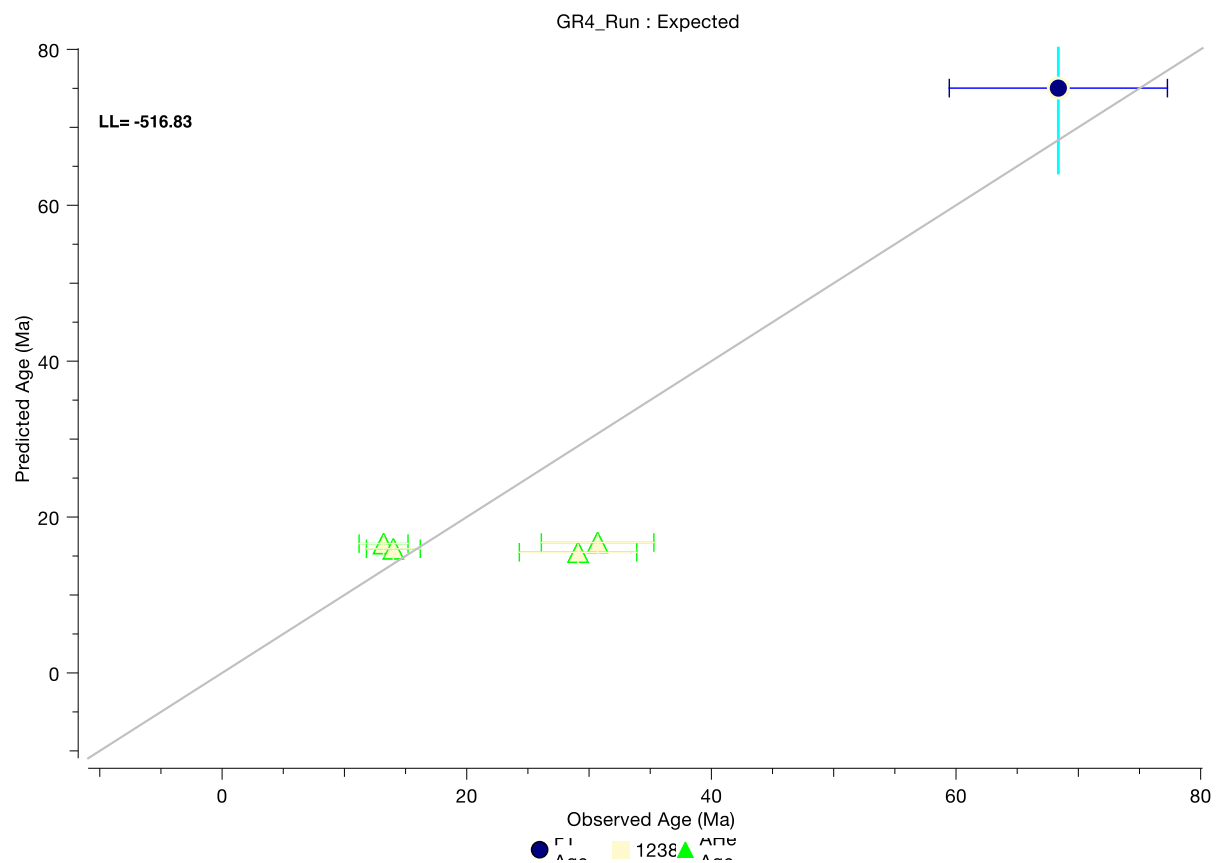
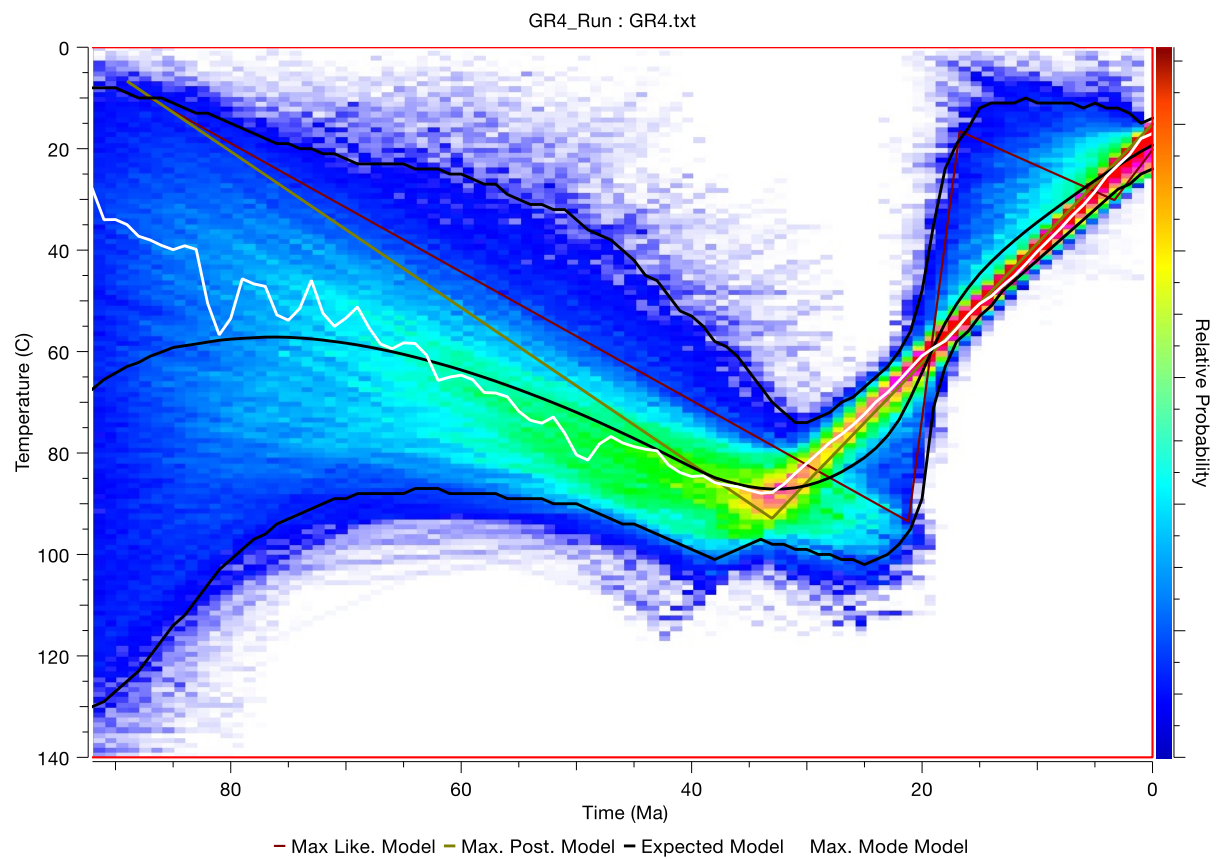
GR3

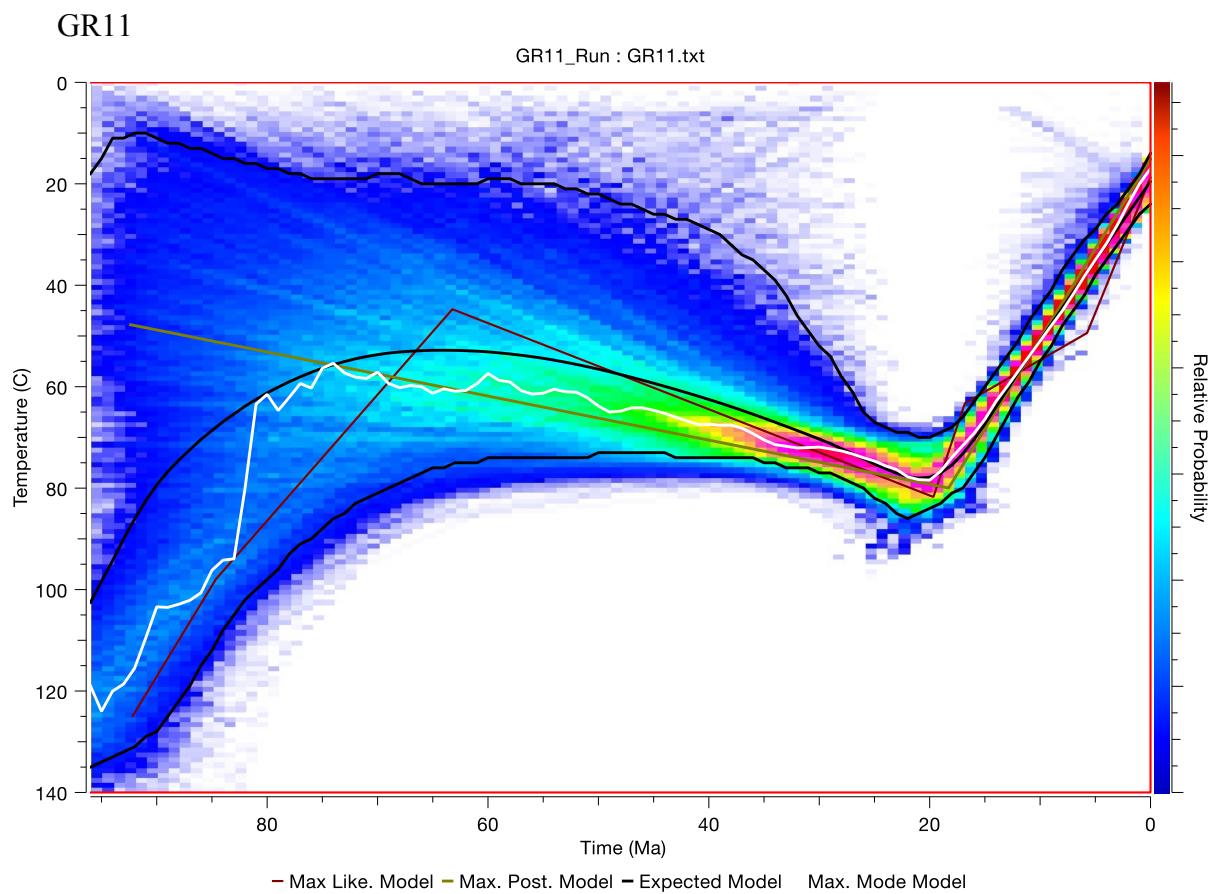
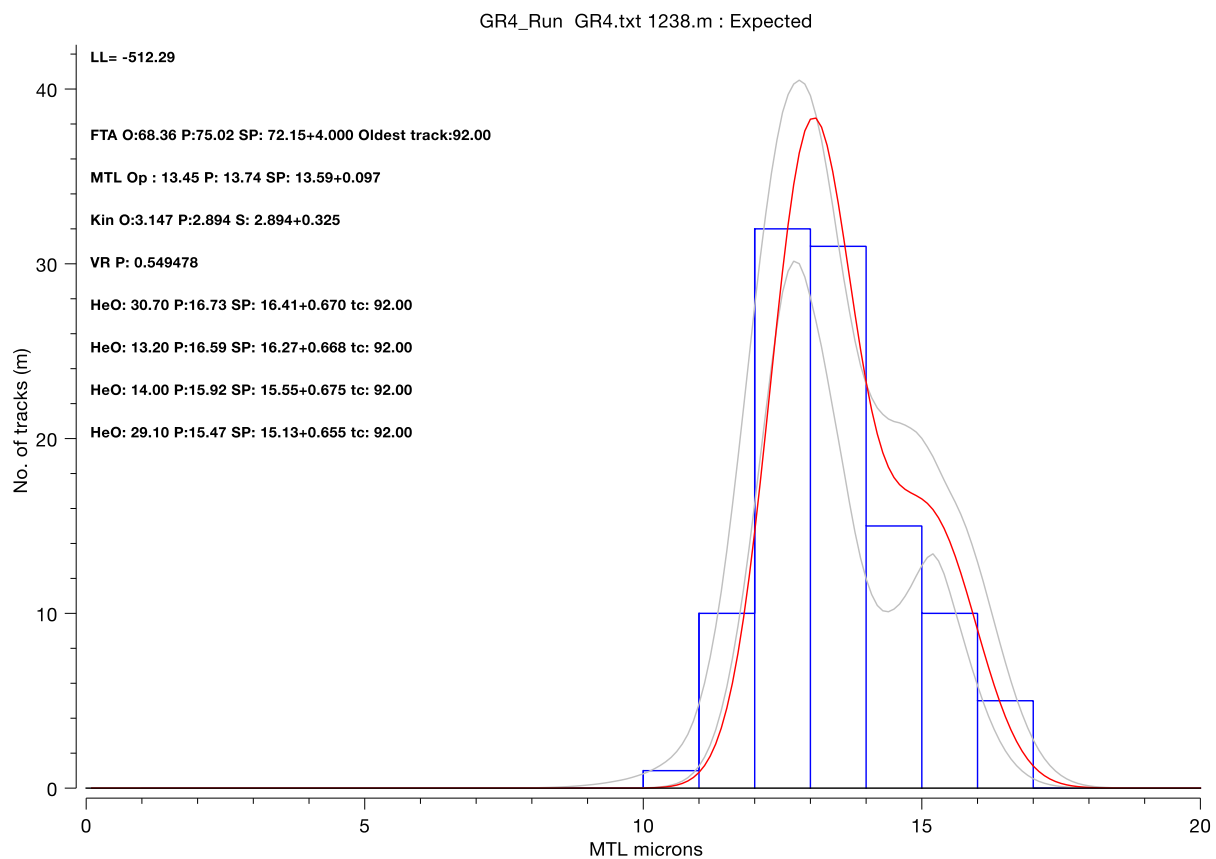
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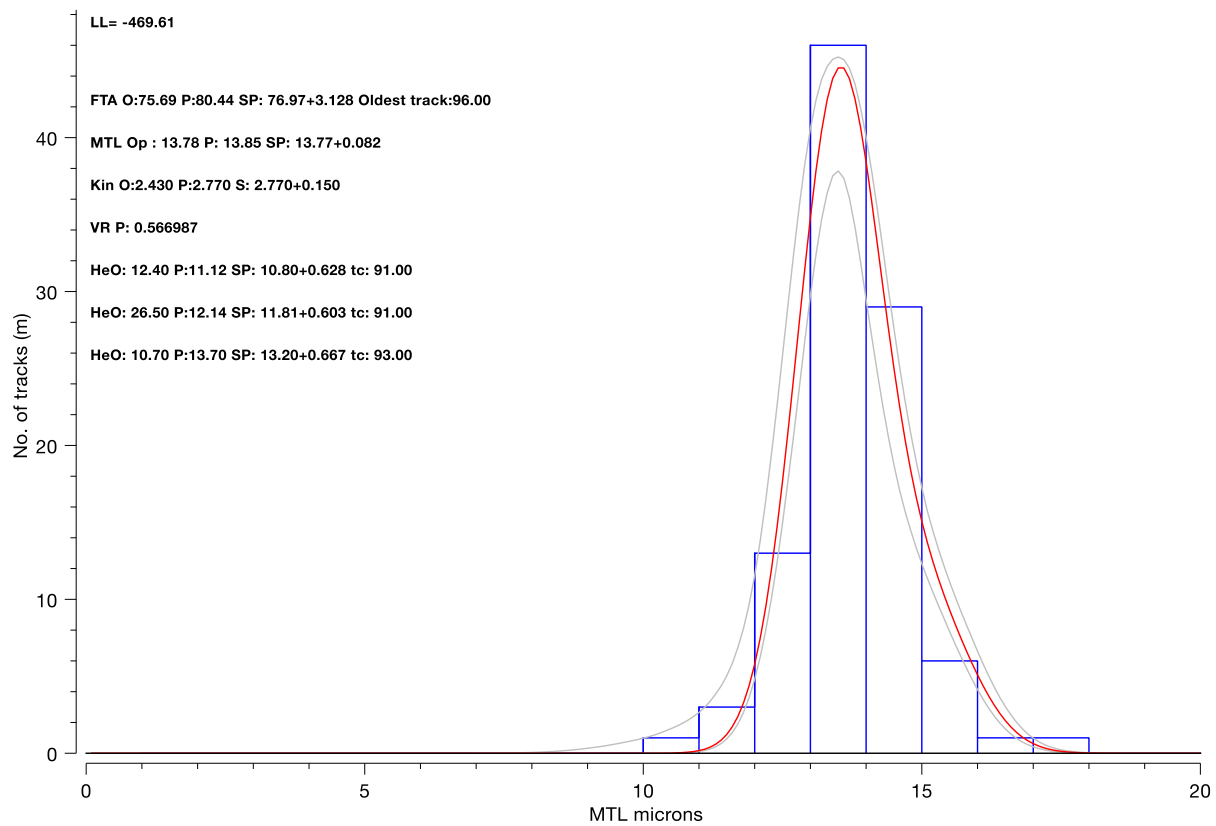
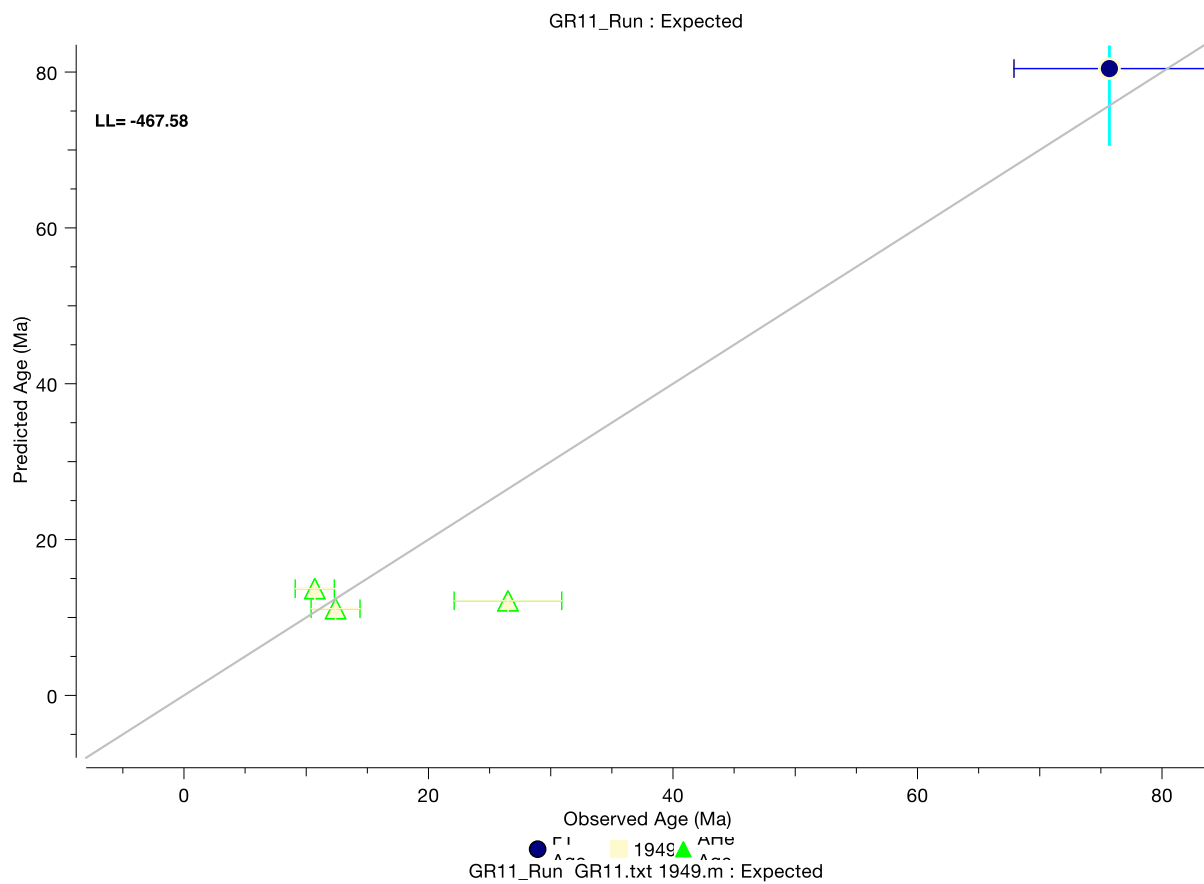




GR4

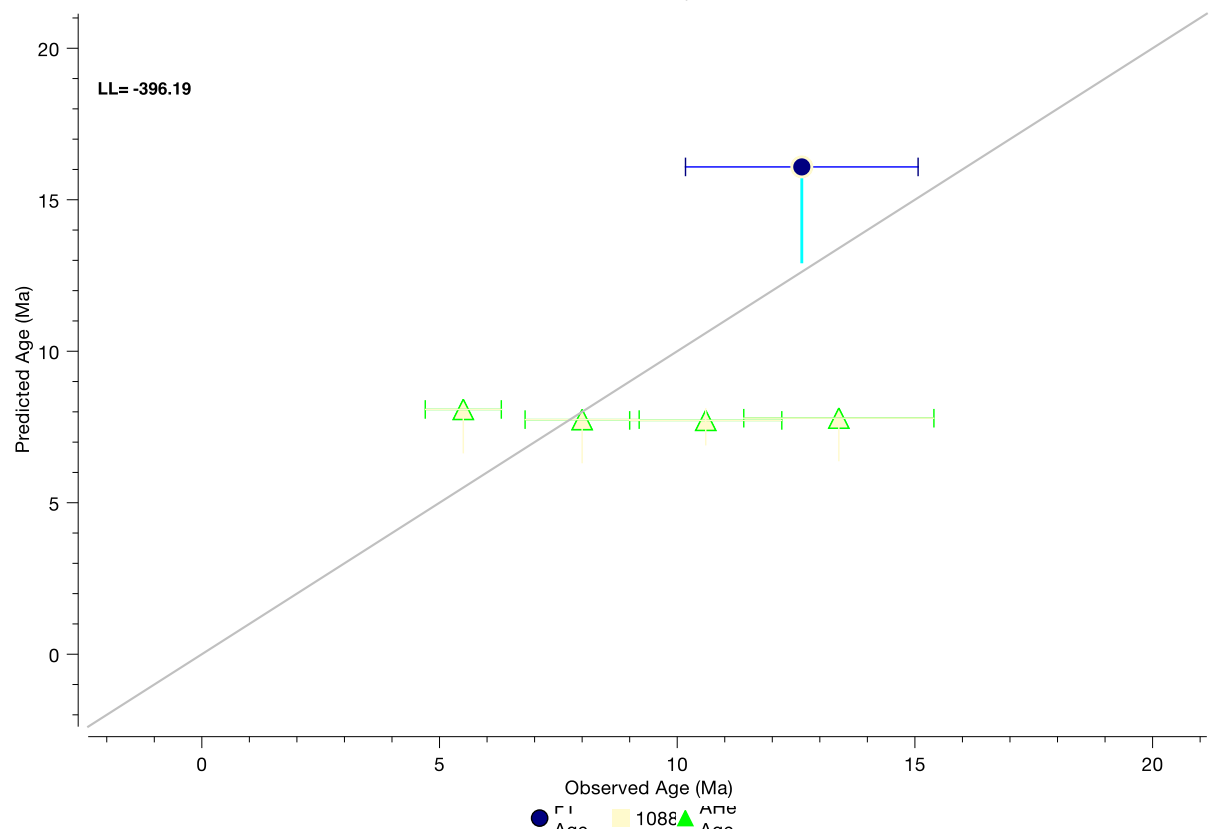
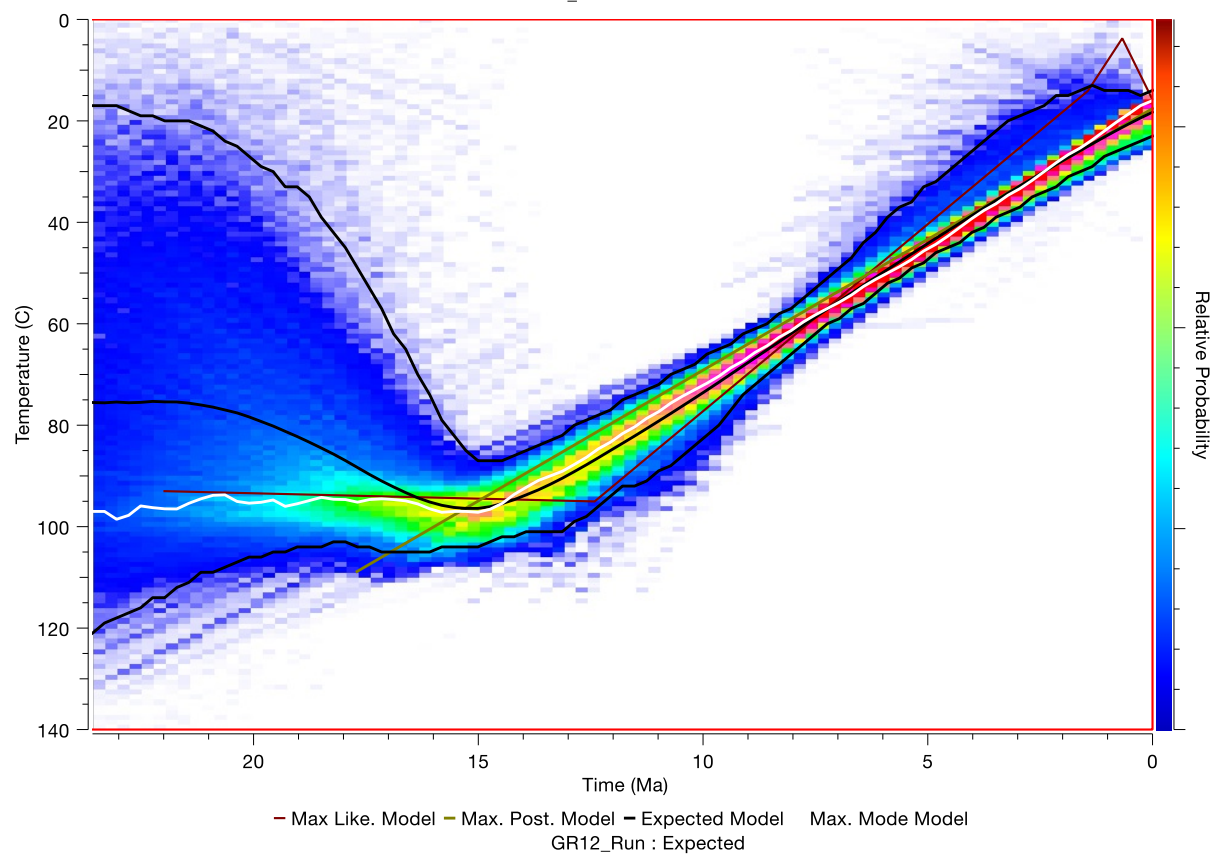


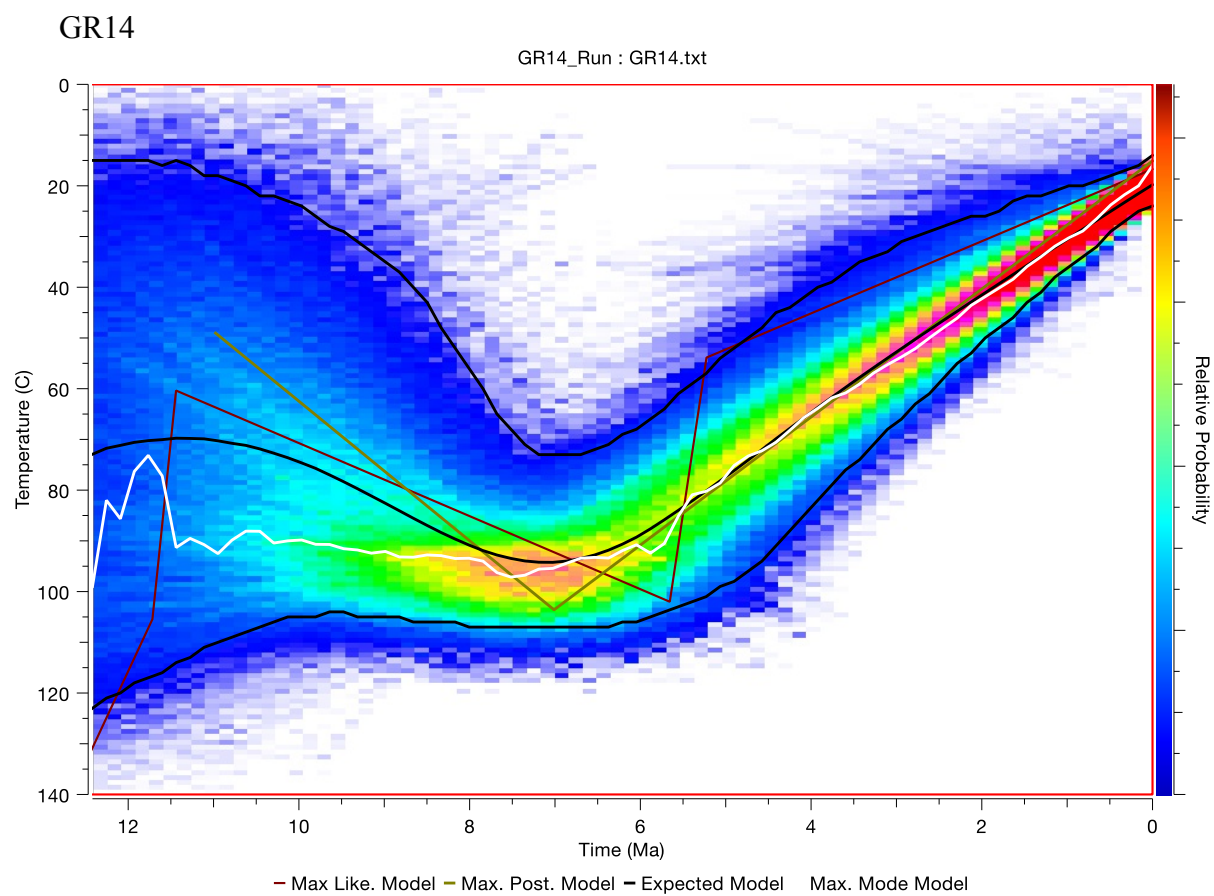
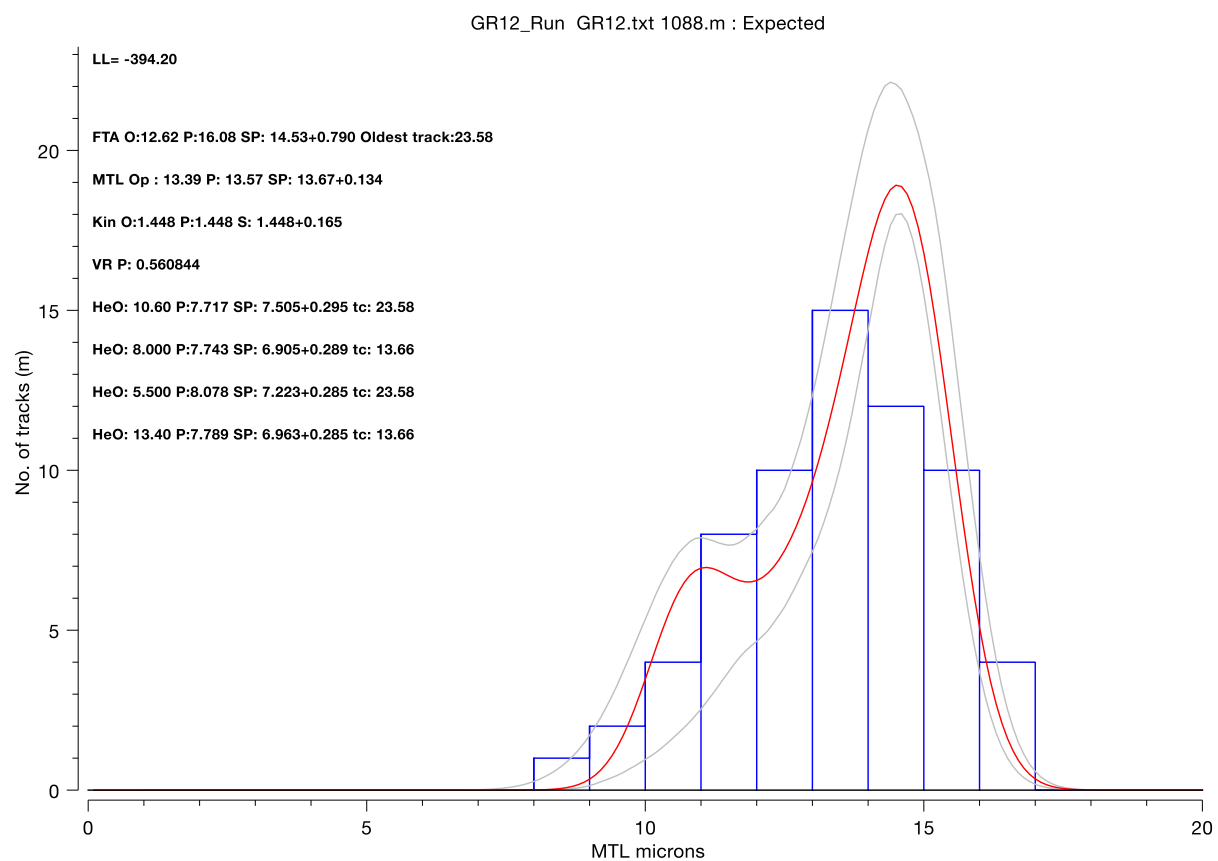


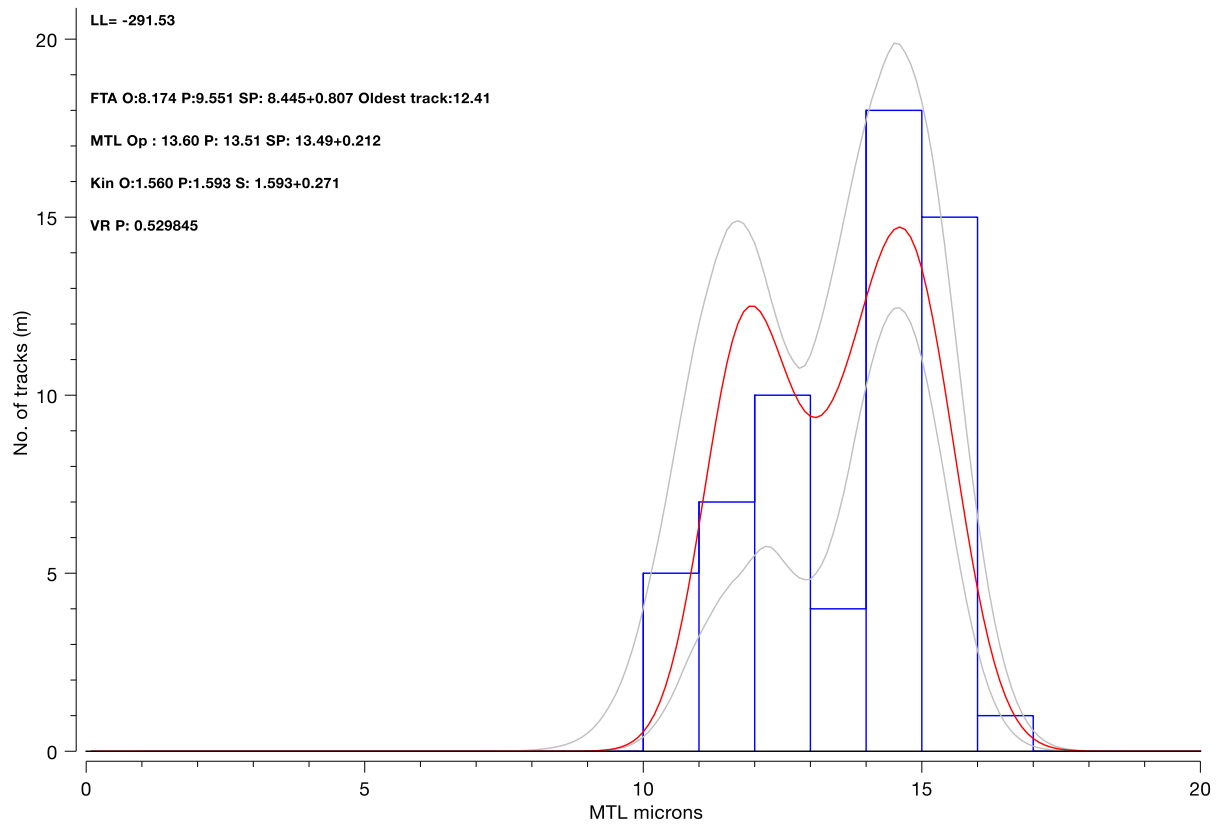
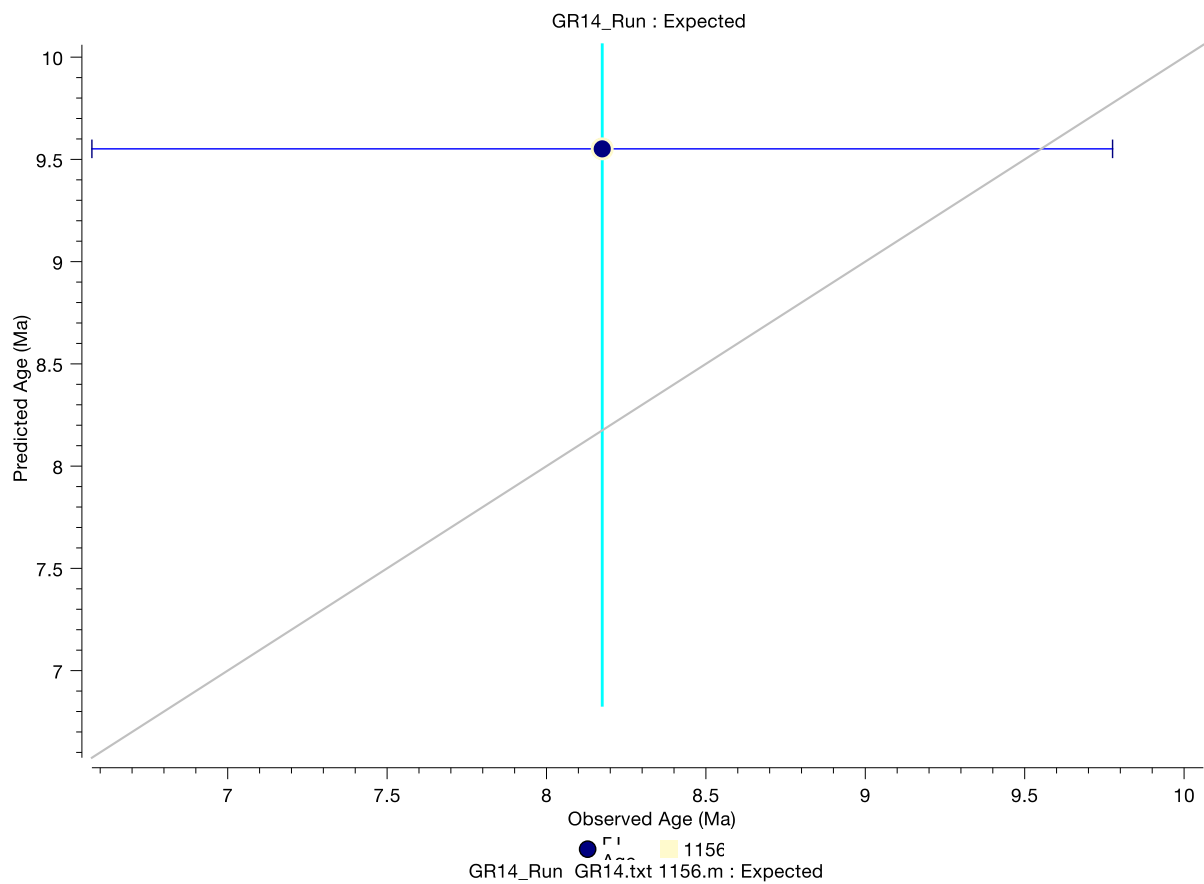


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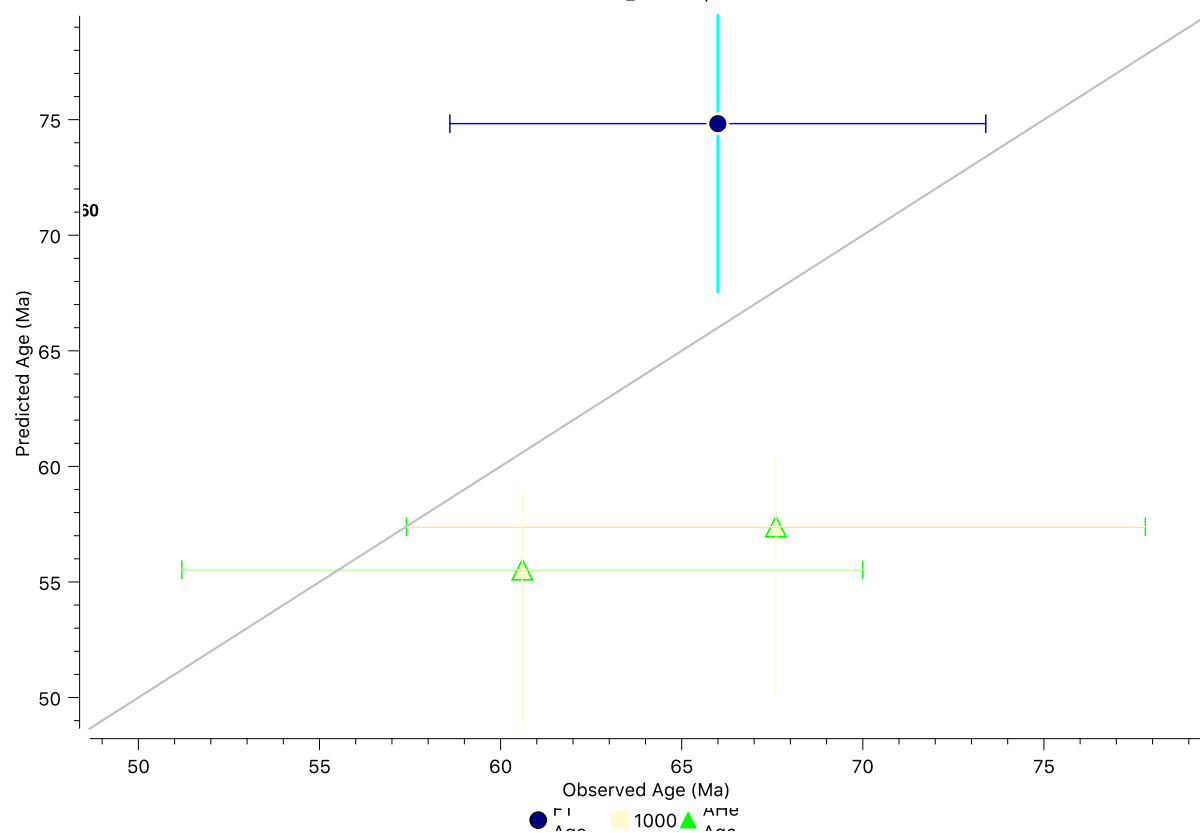
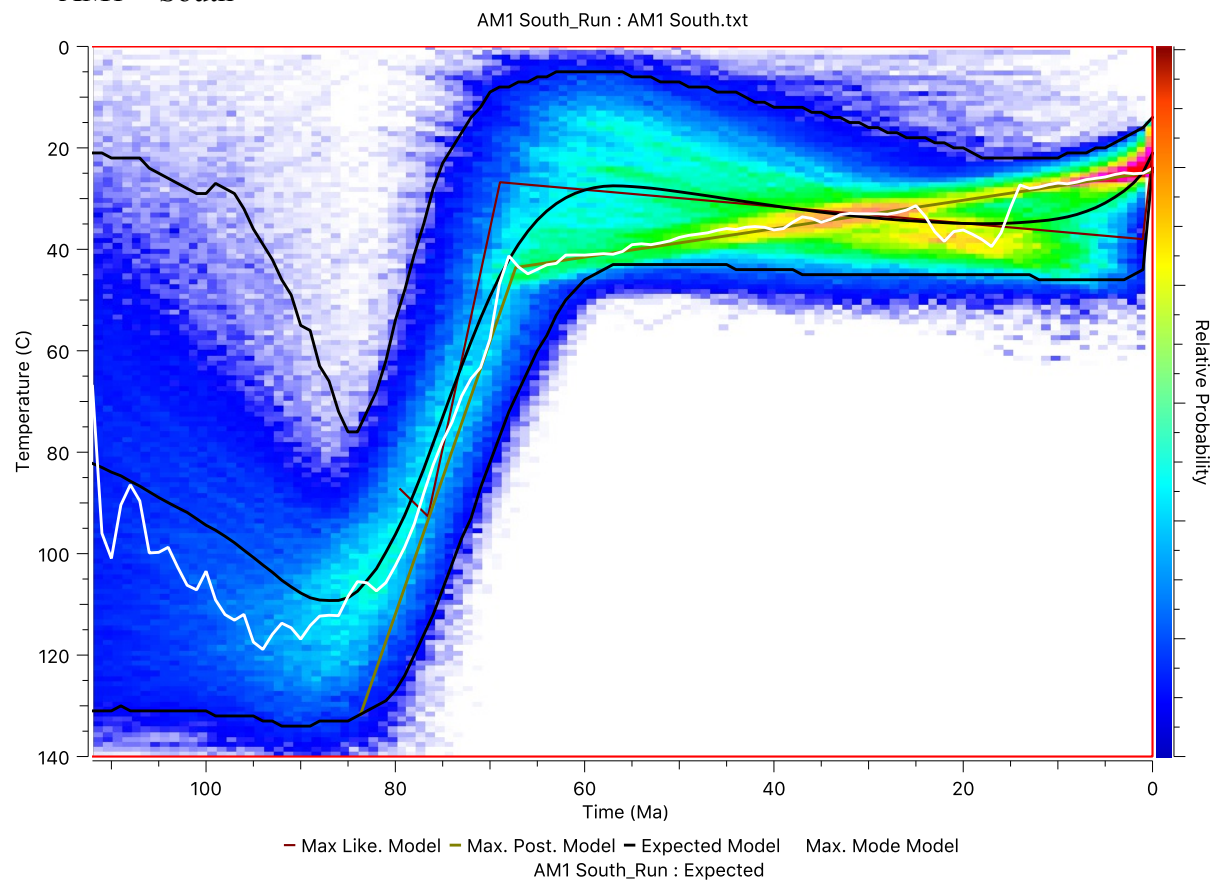
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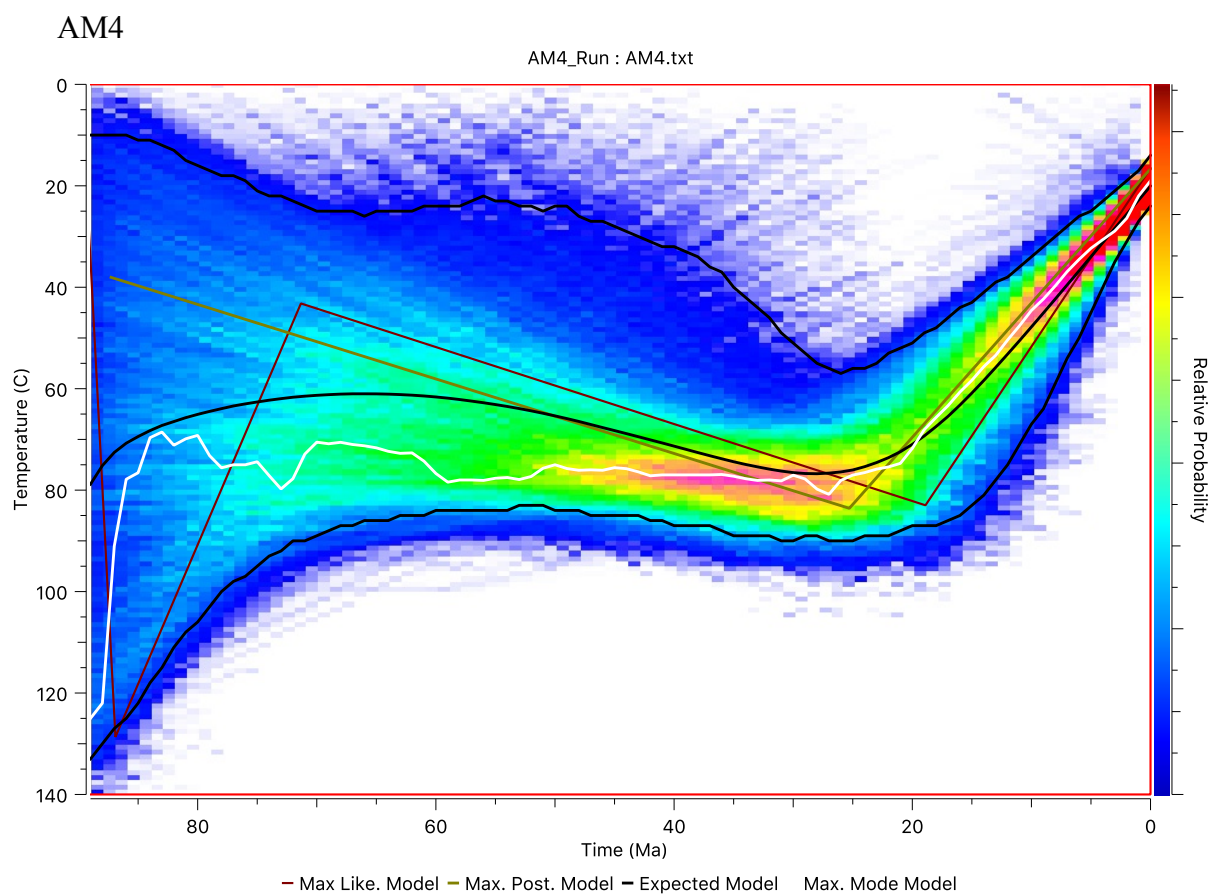
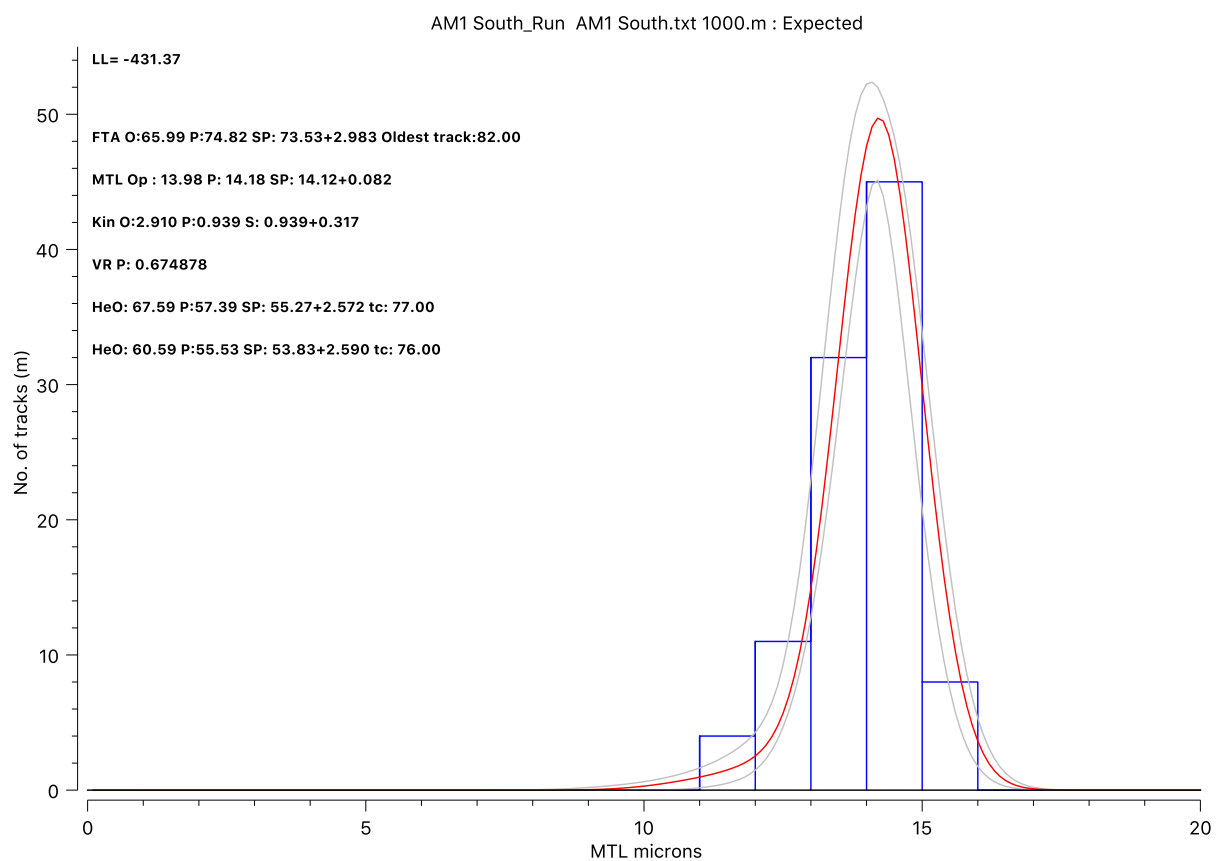


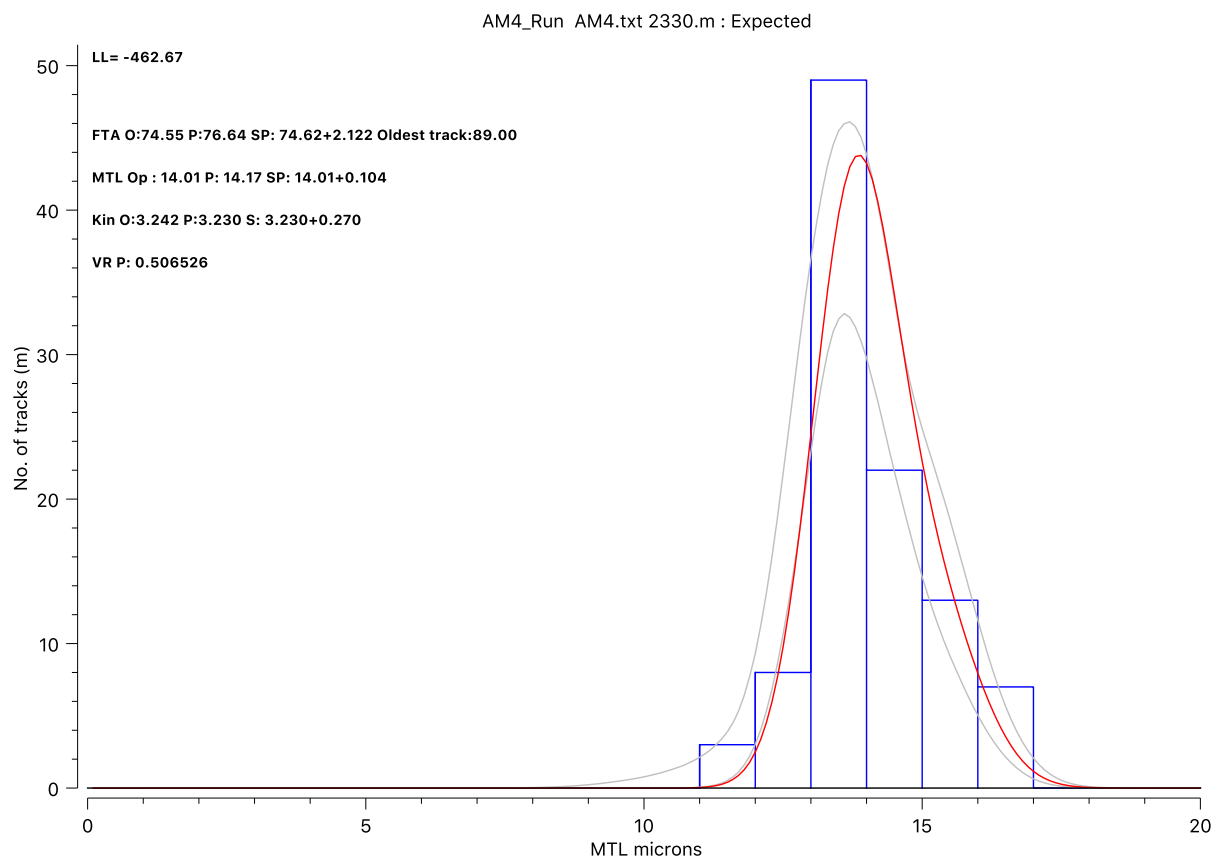
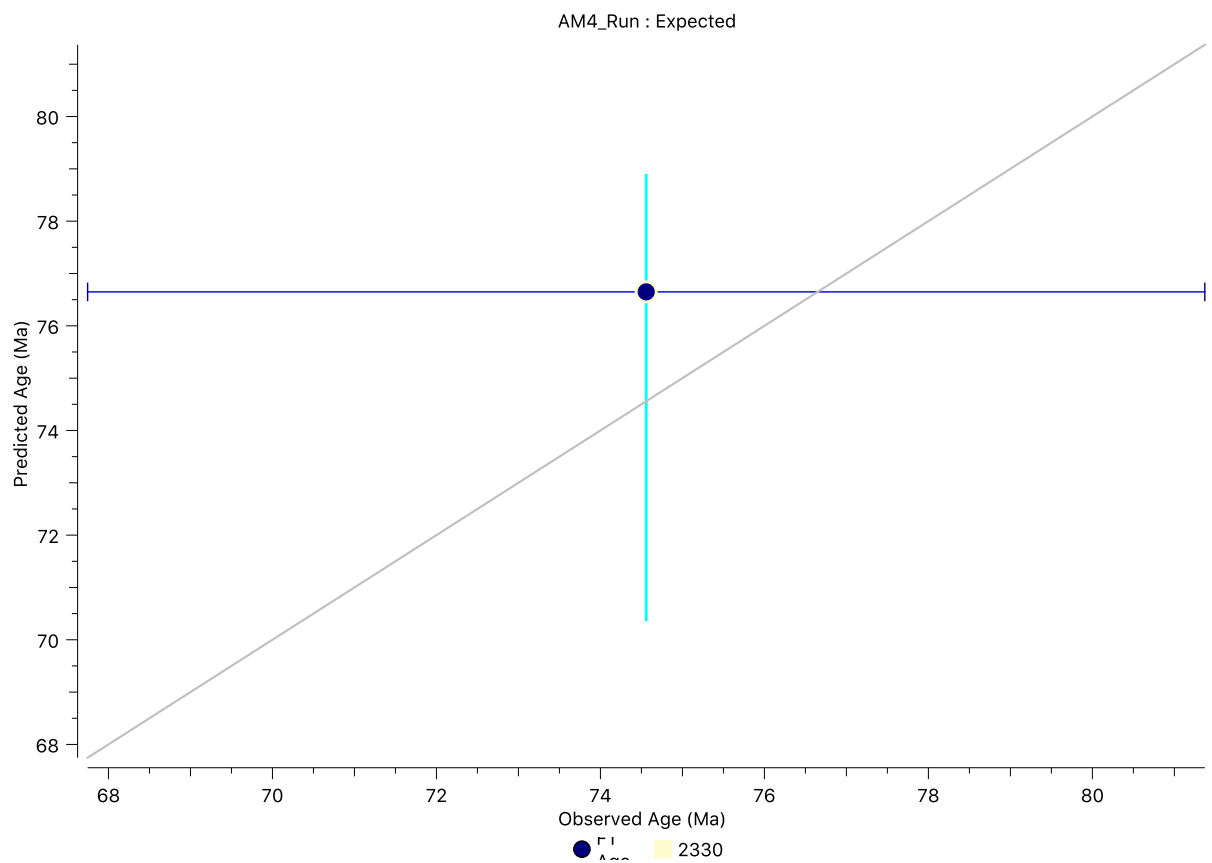




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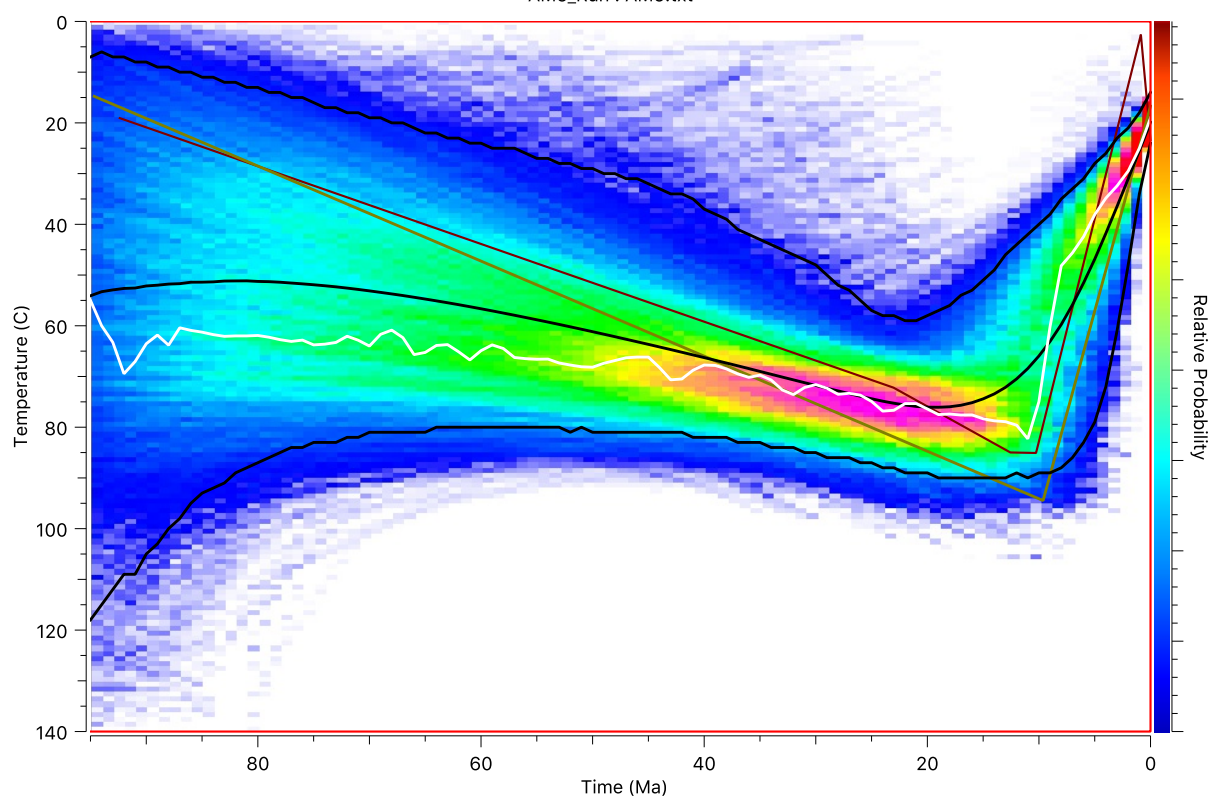




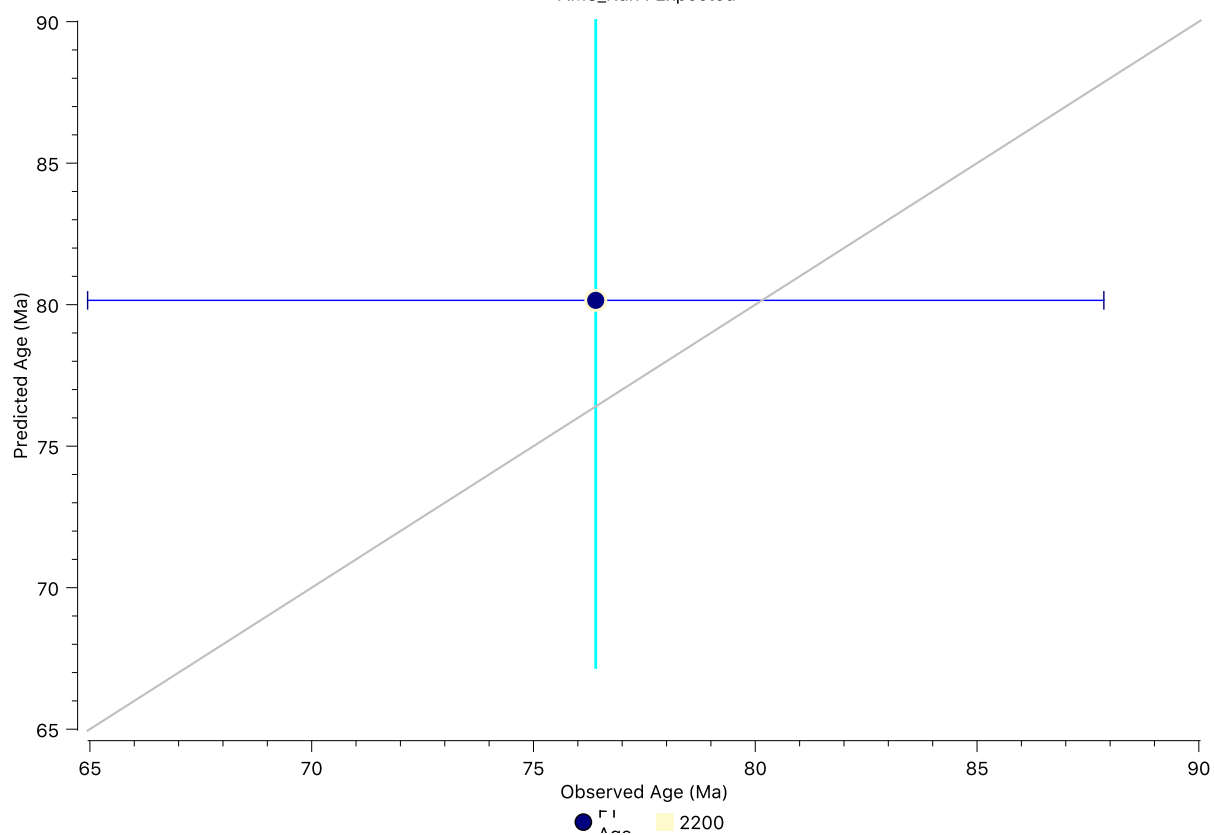


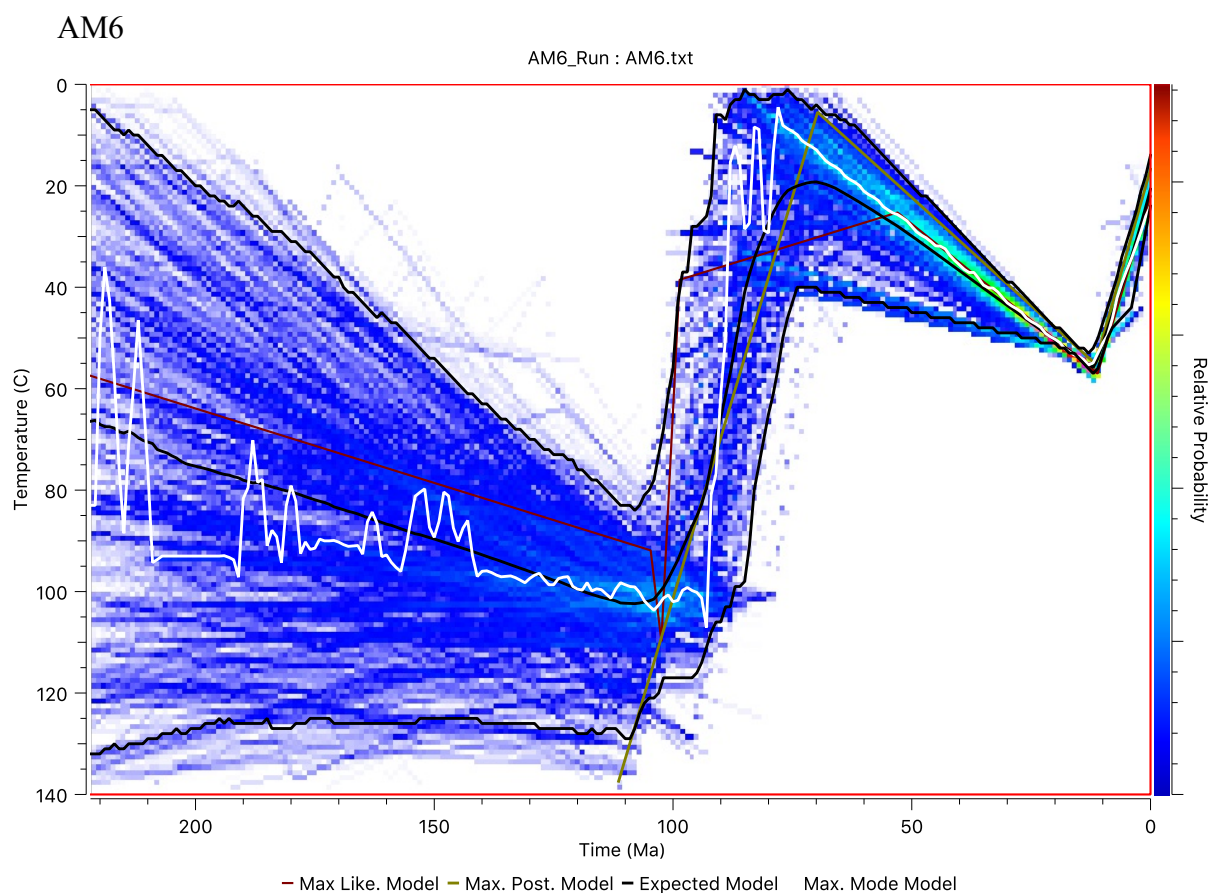
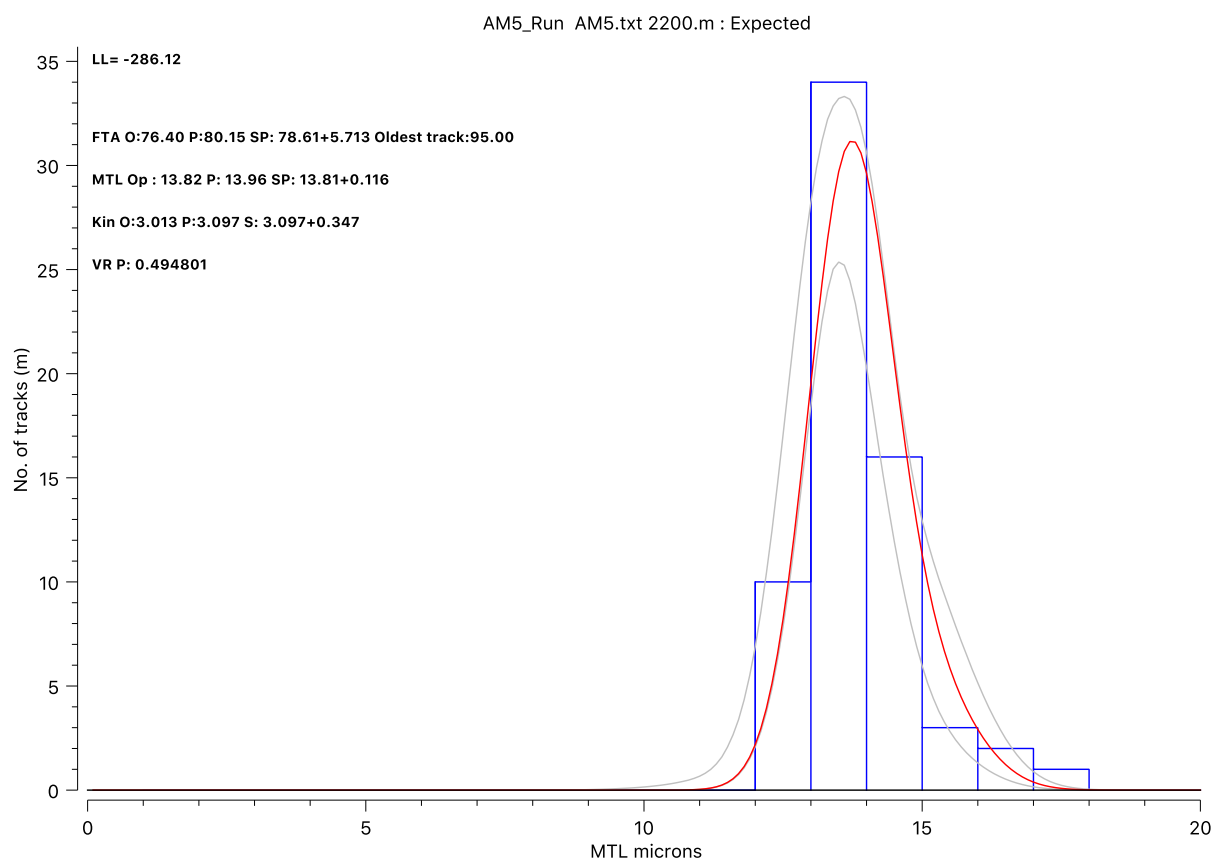
AM5

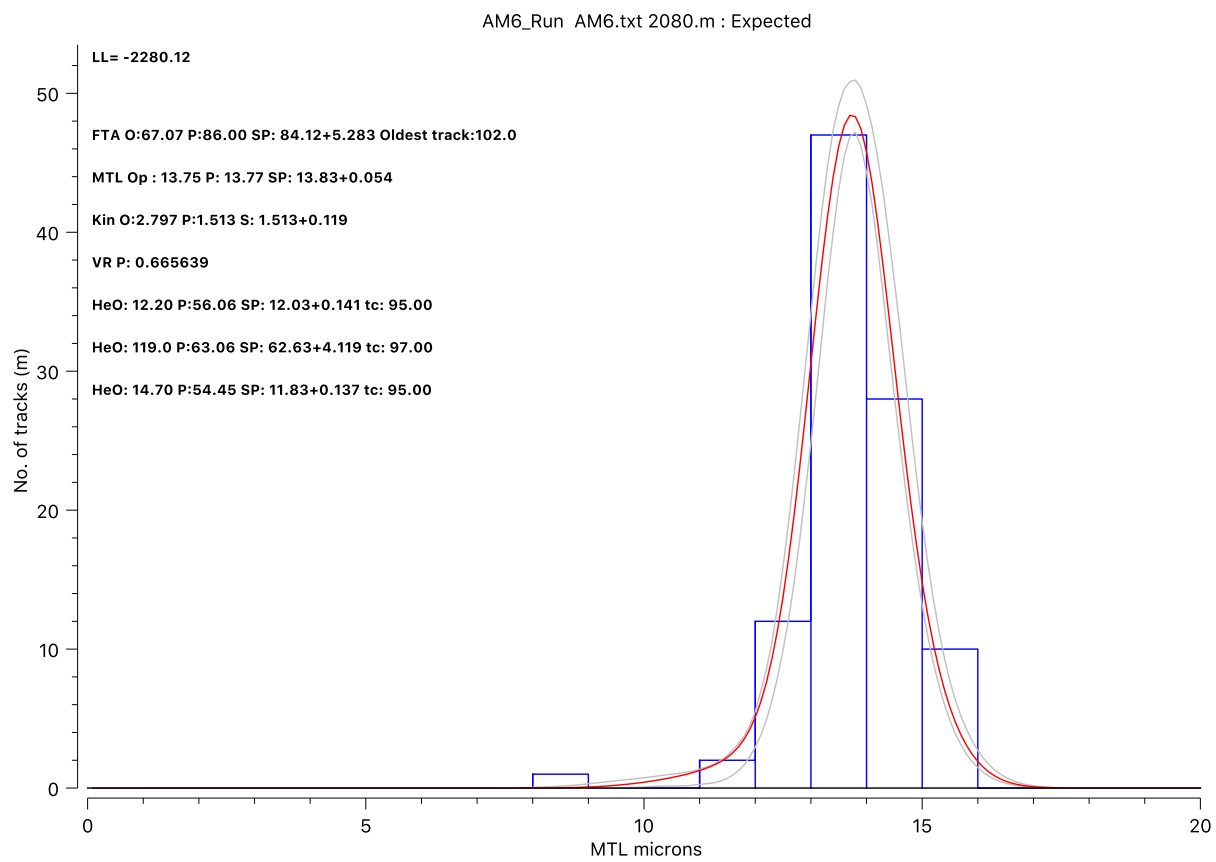
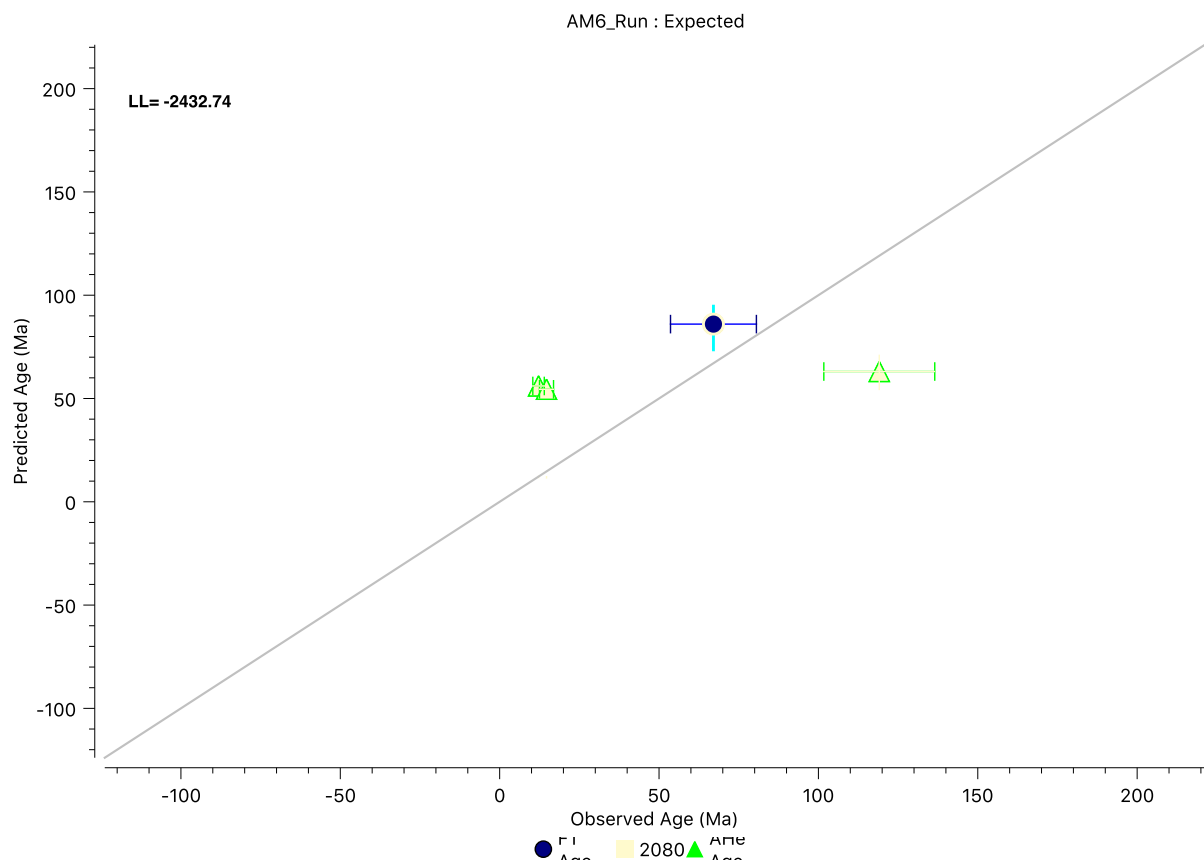
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AM5_Run : Expected

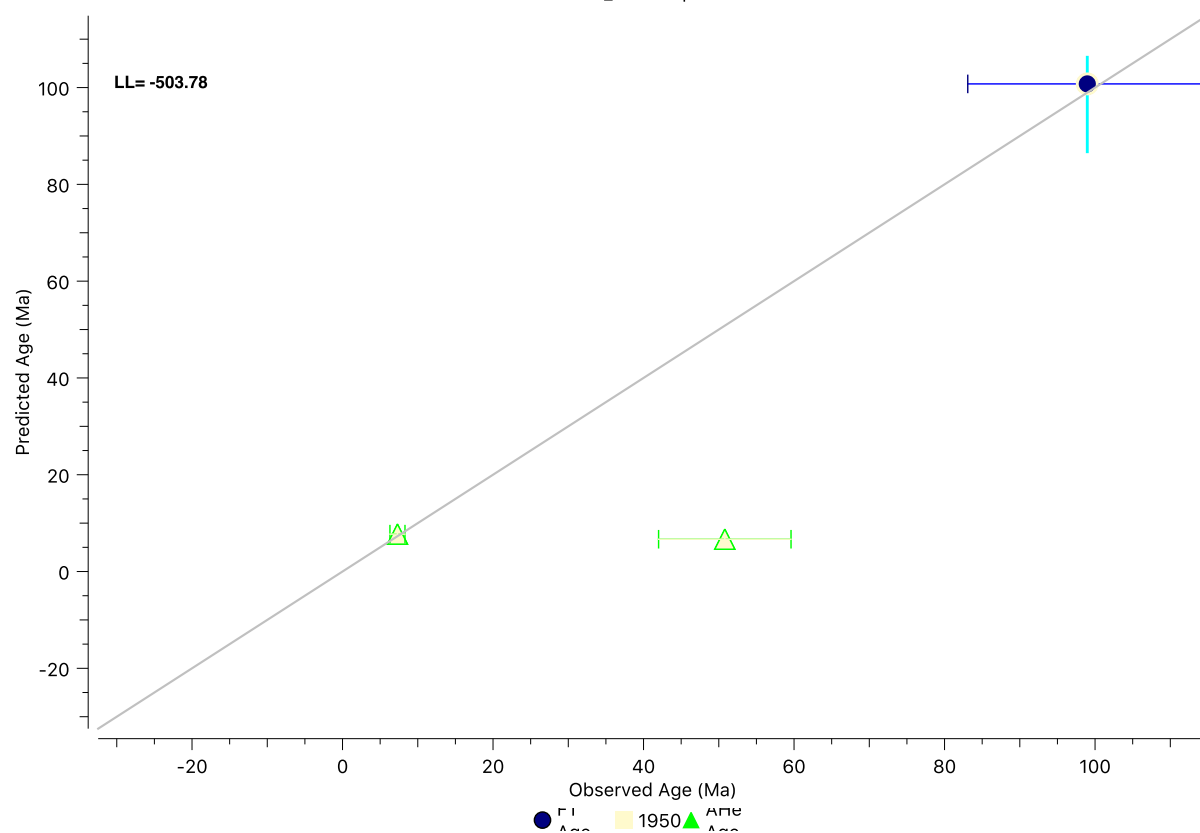
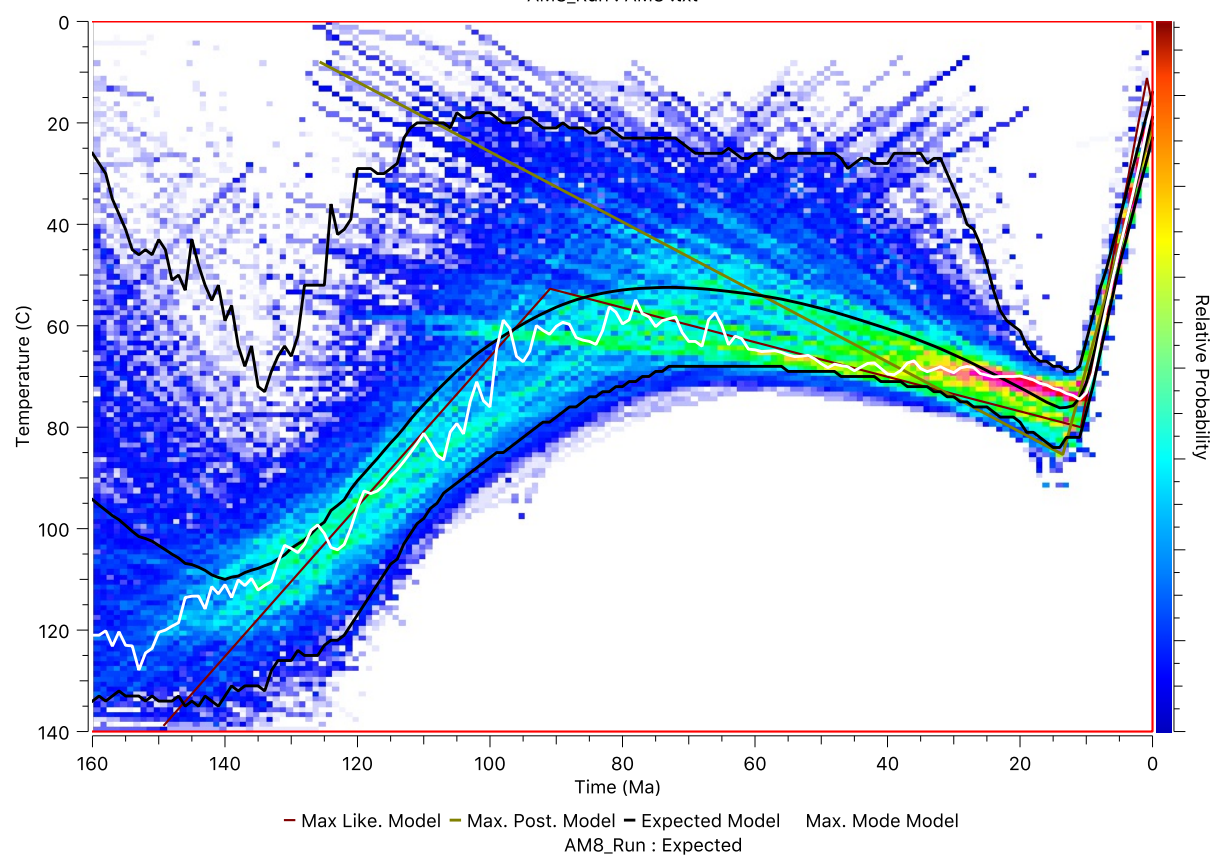


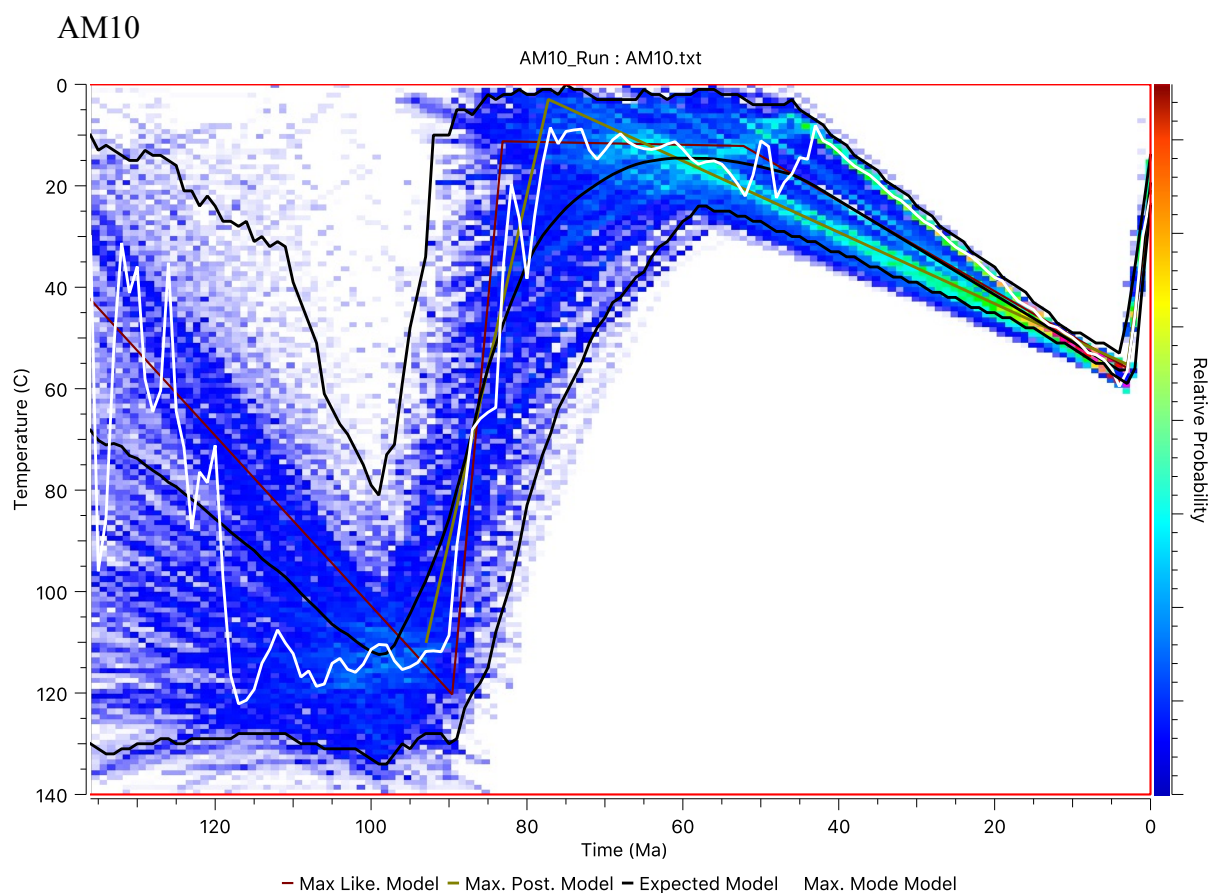
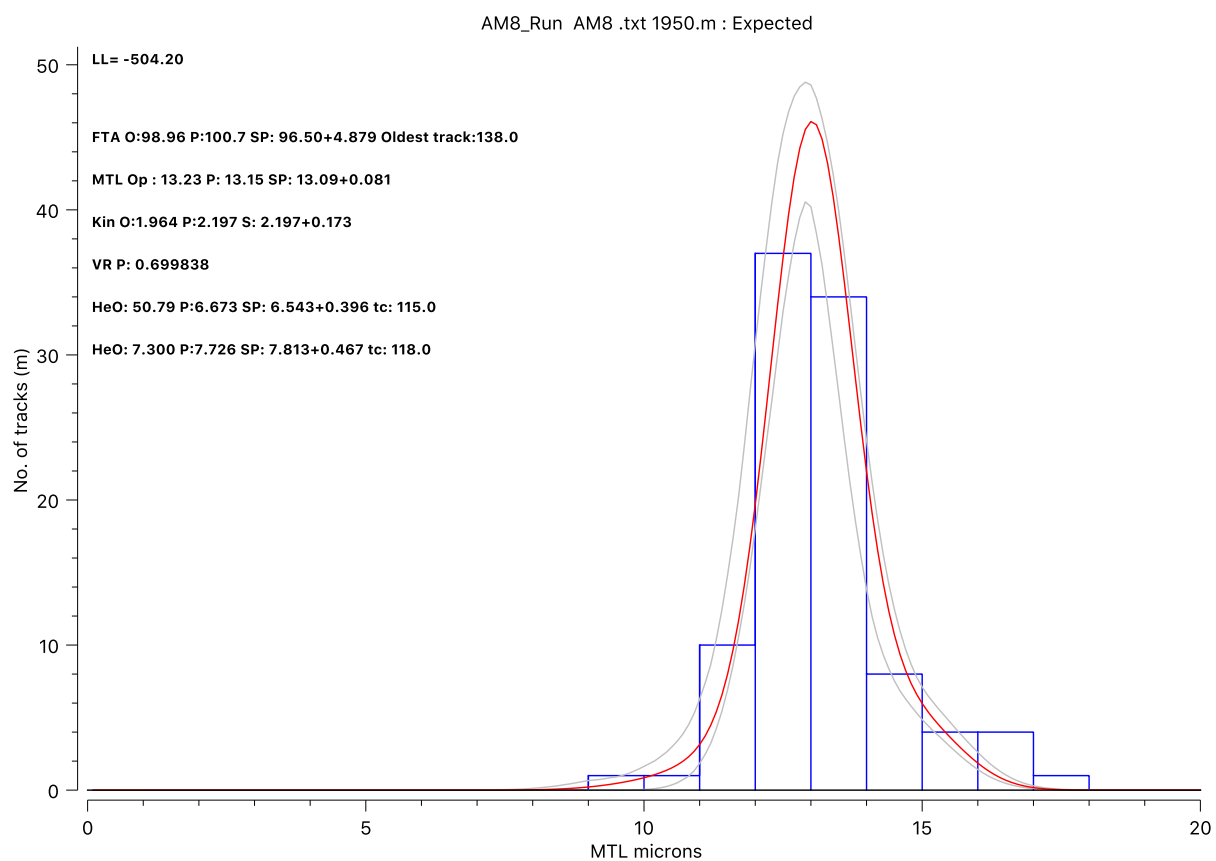


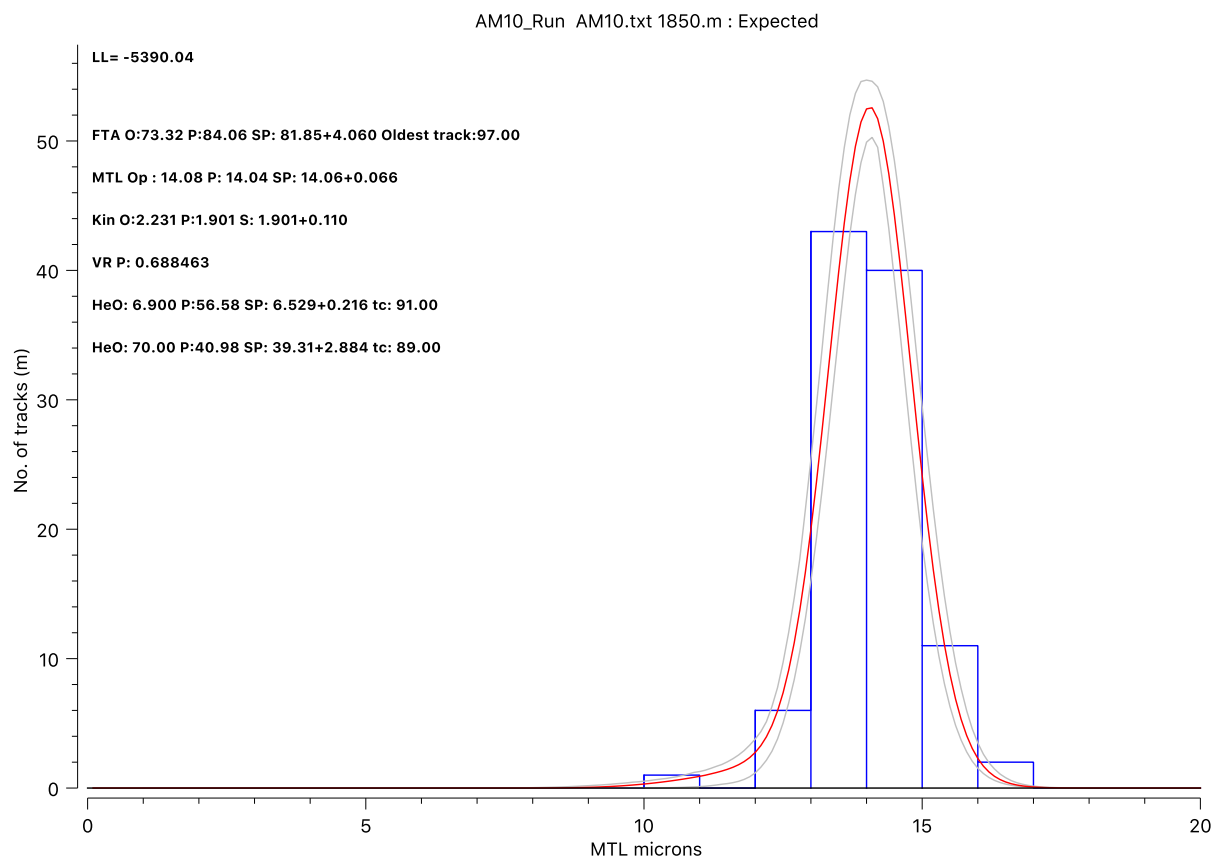
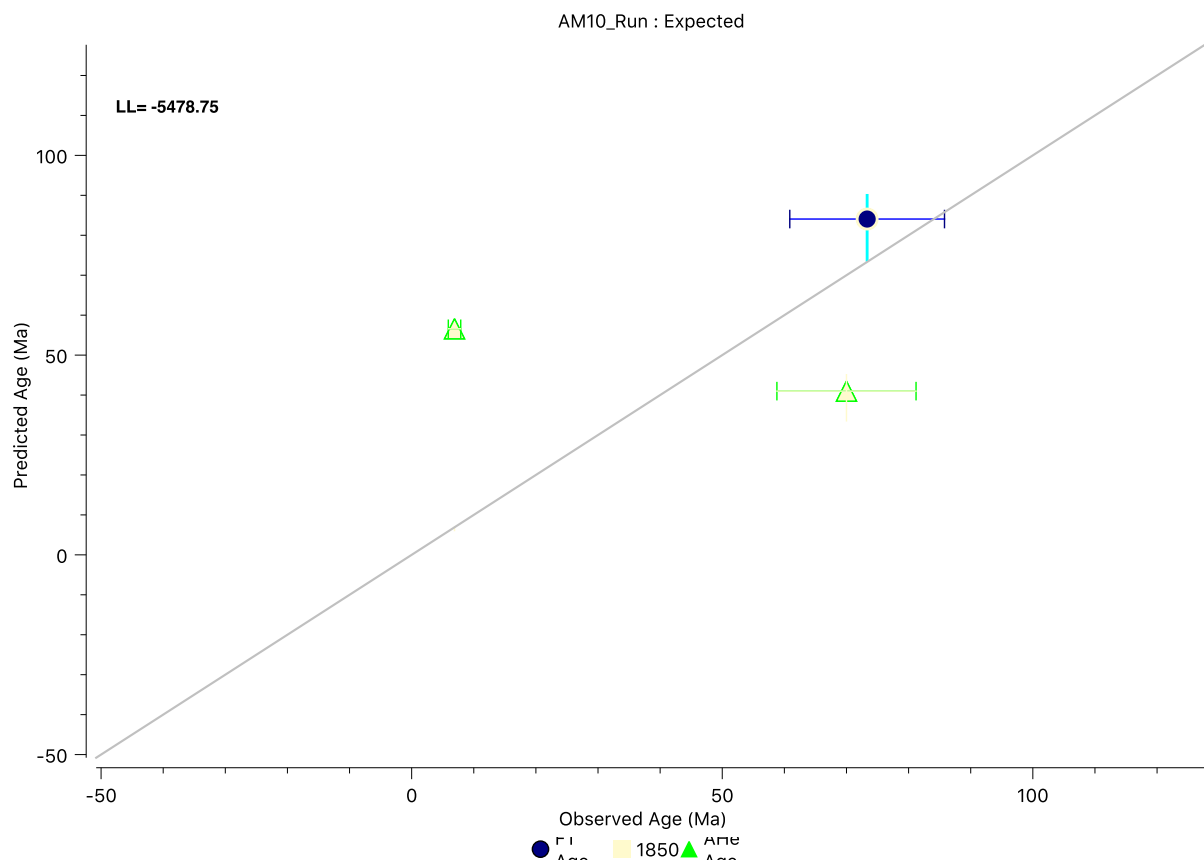


AM8

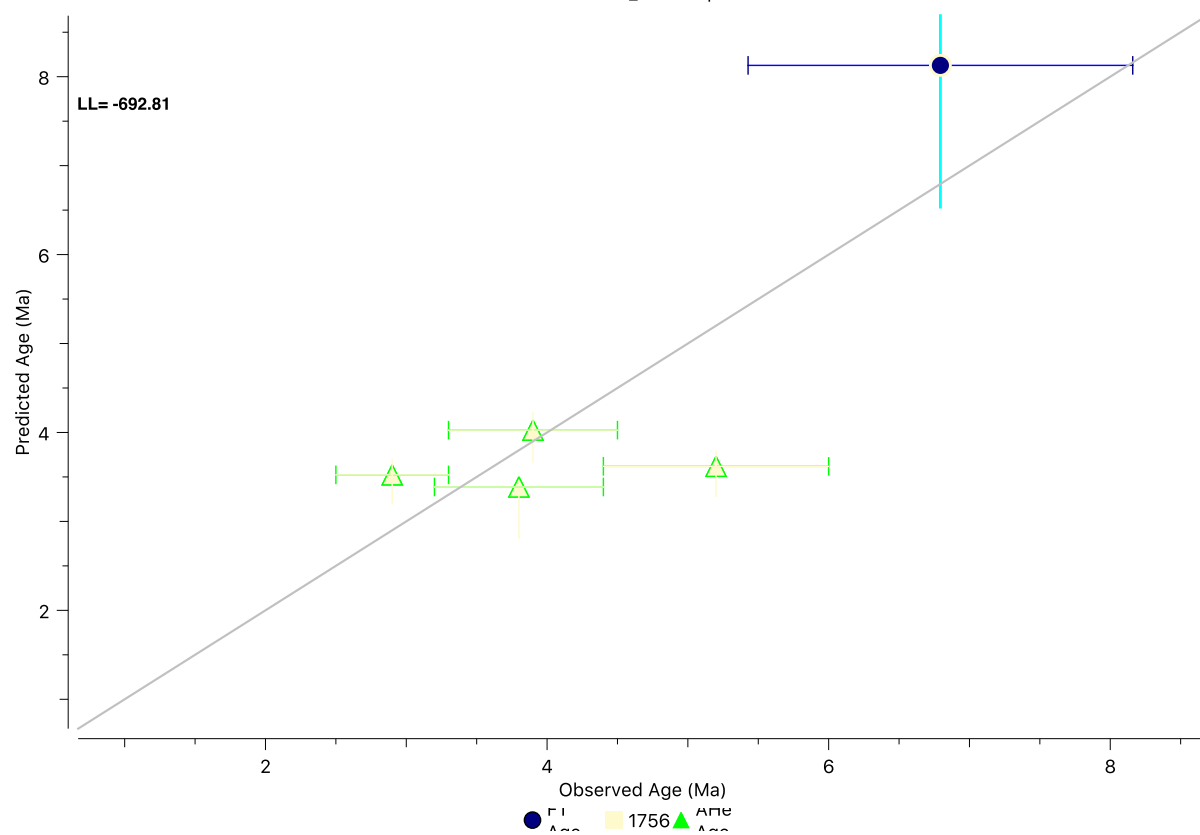
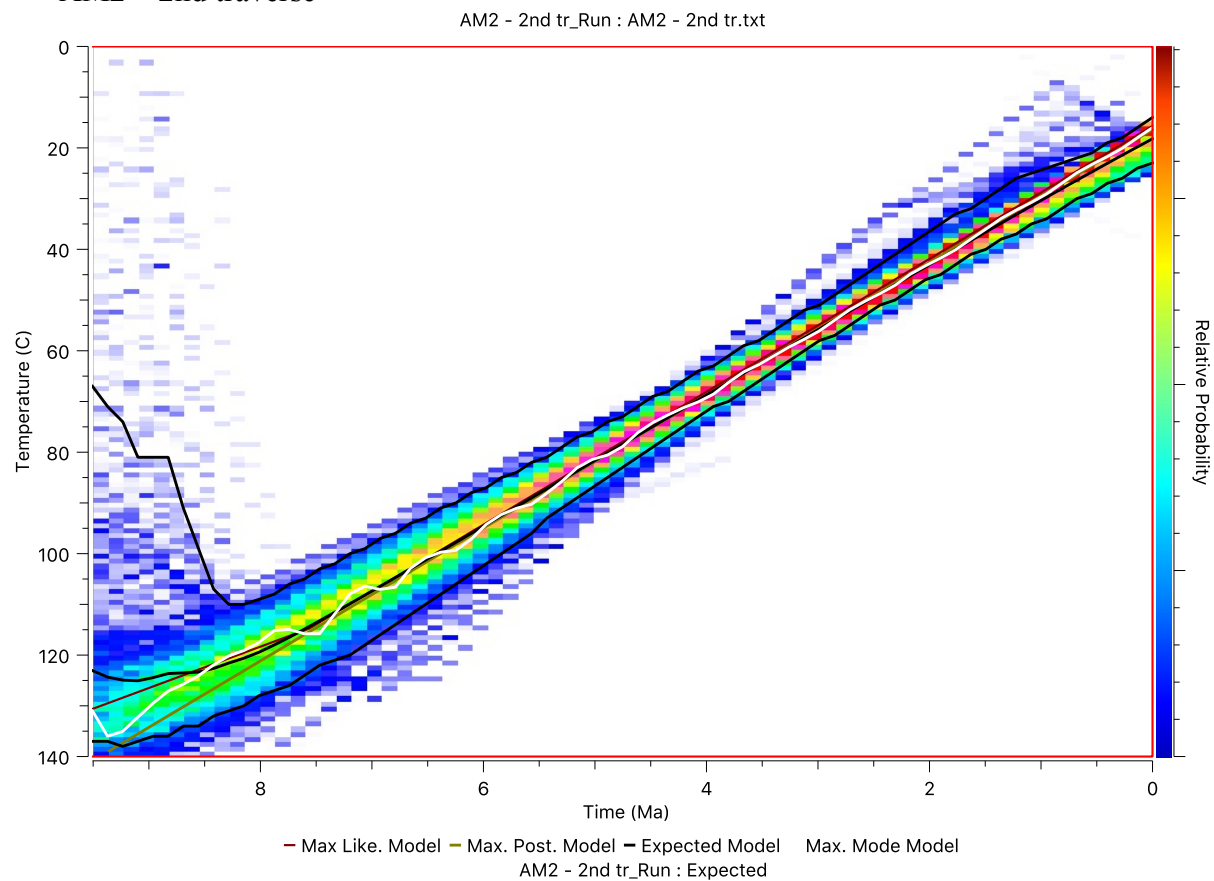
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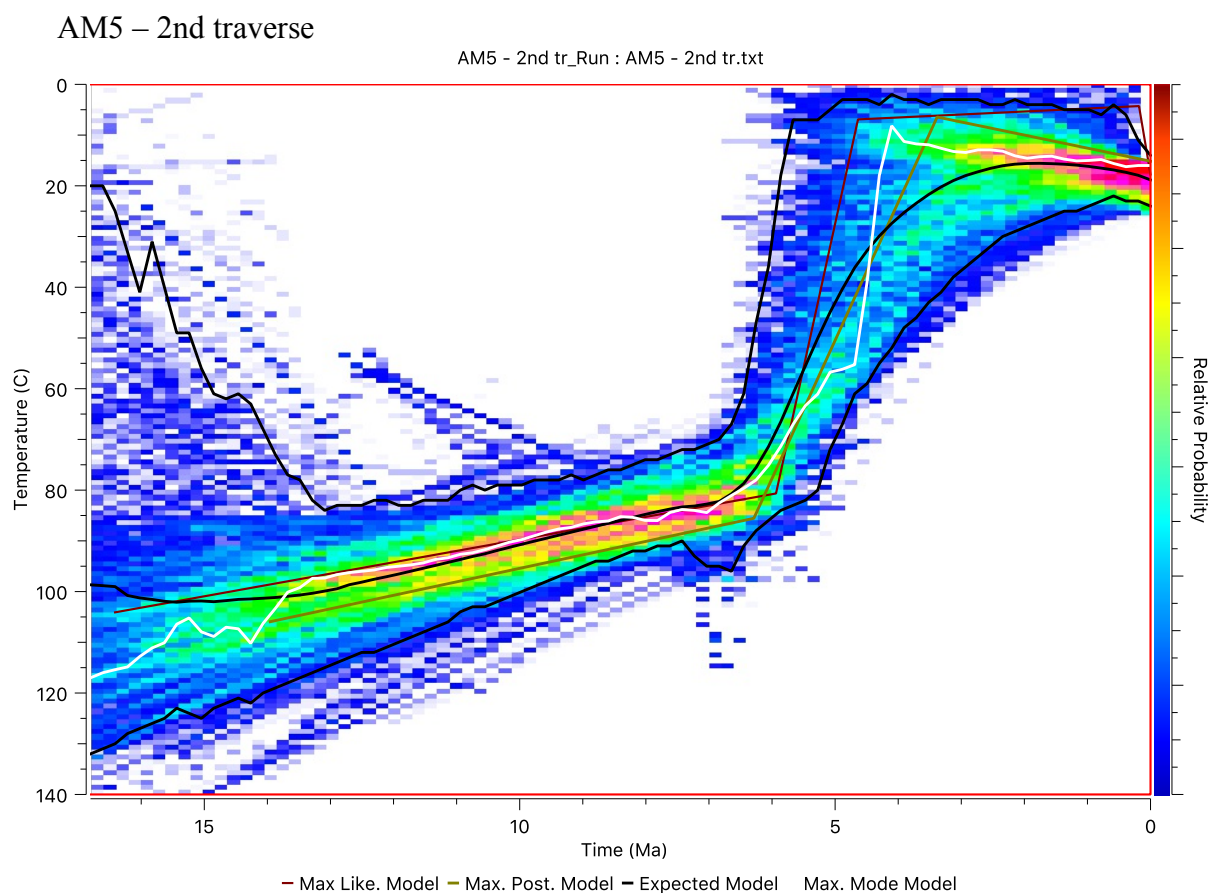
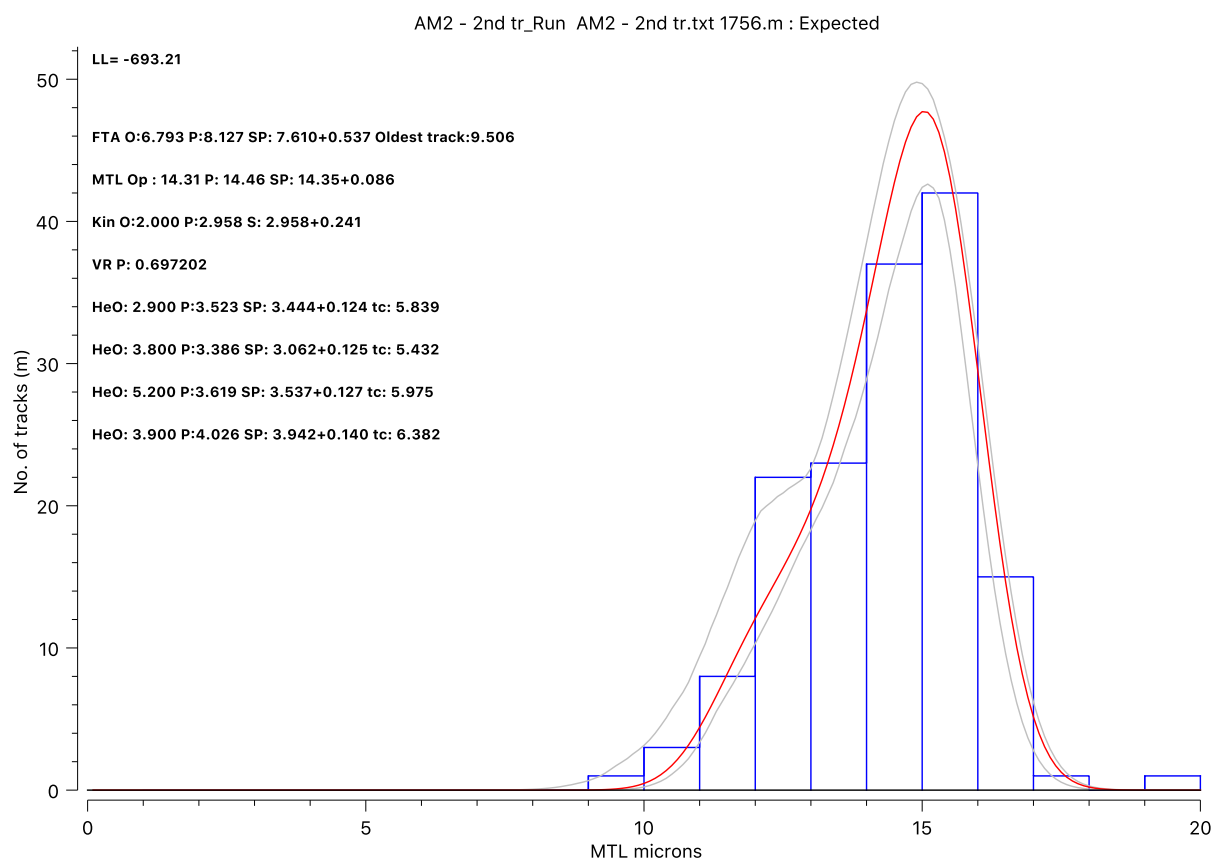


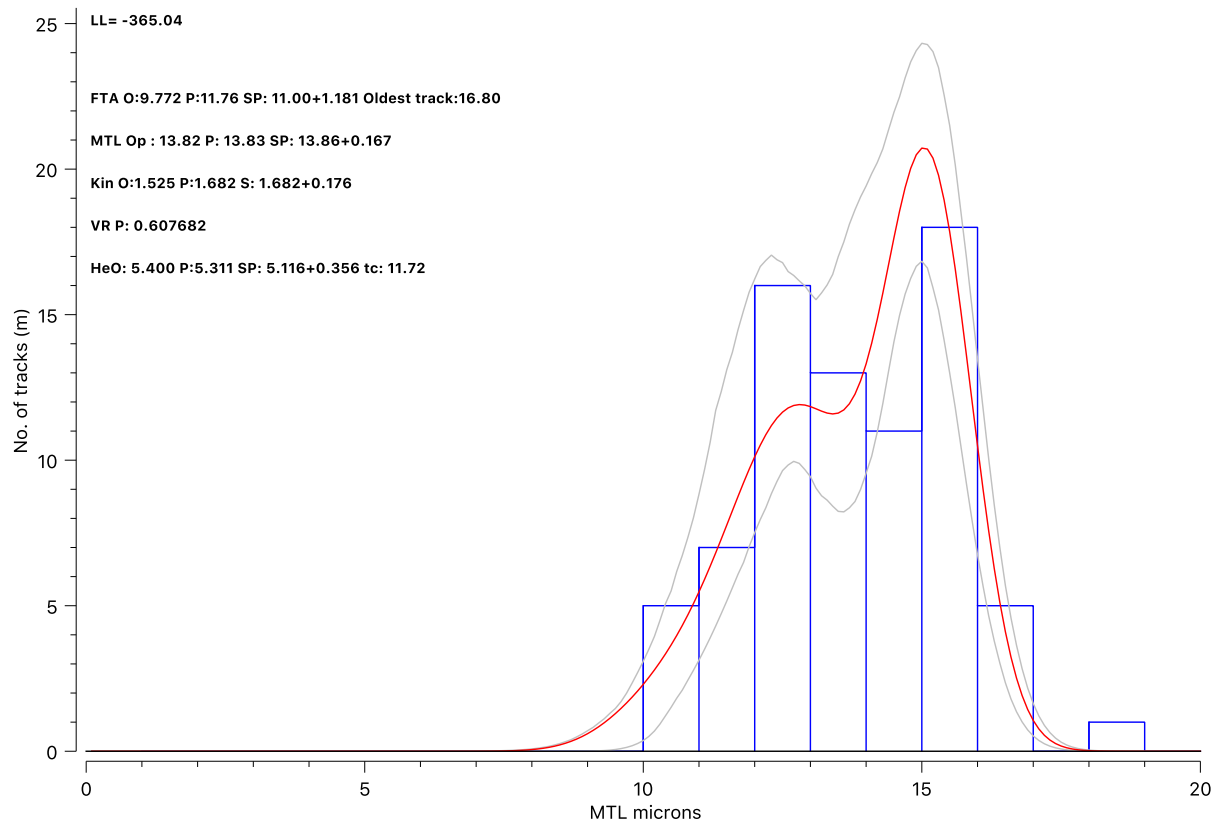
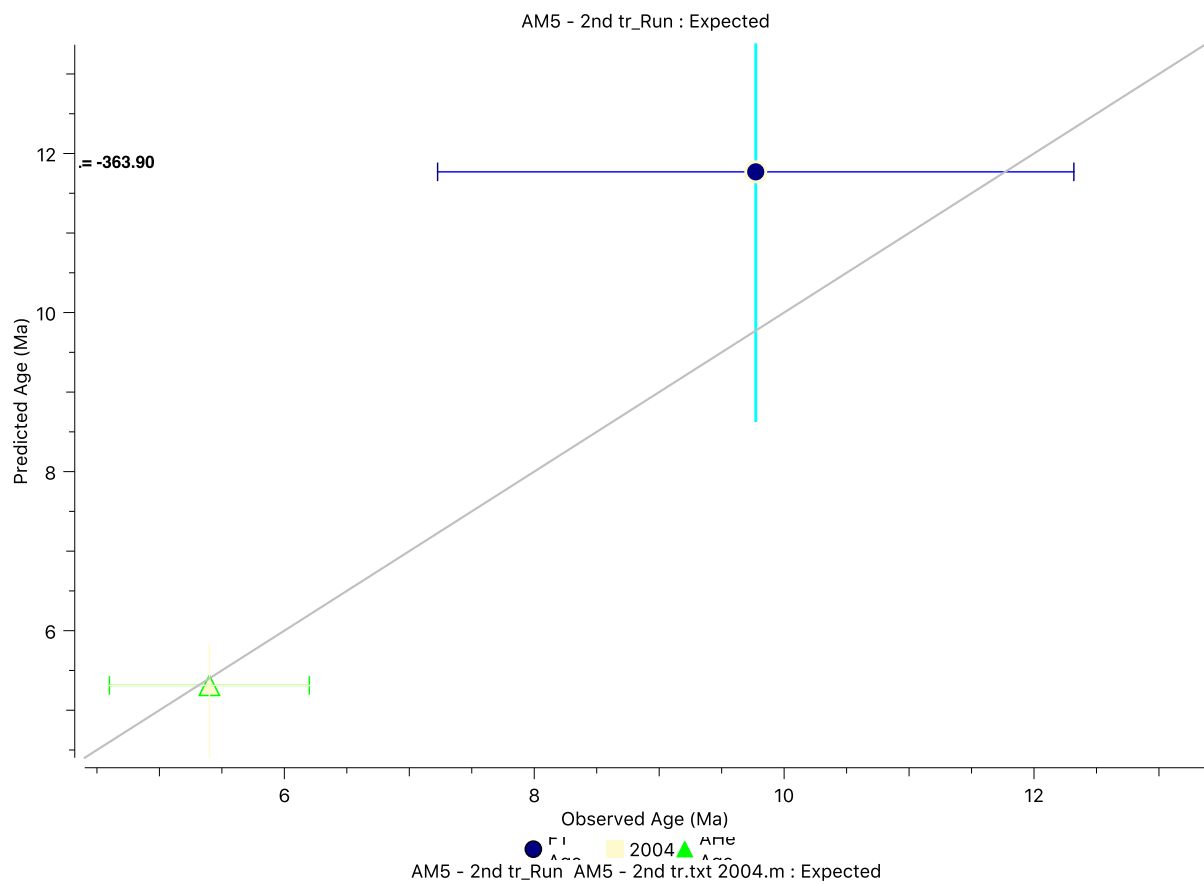




AM2 – 2nd traverse

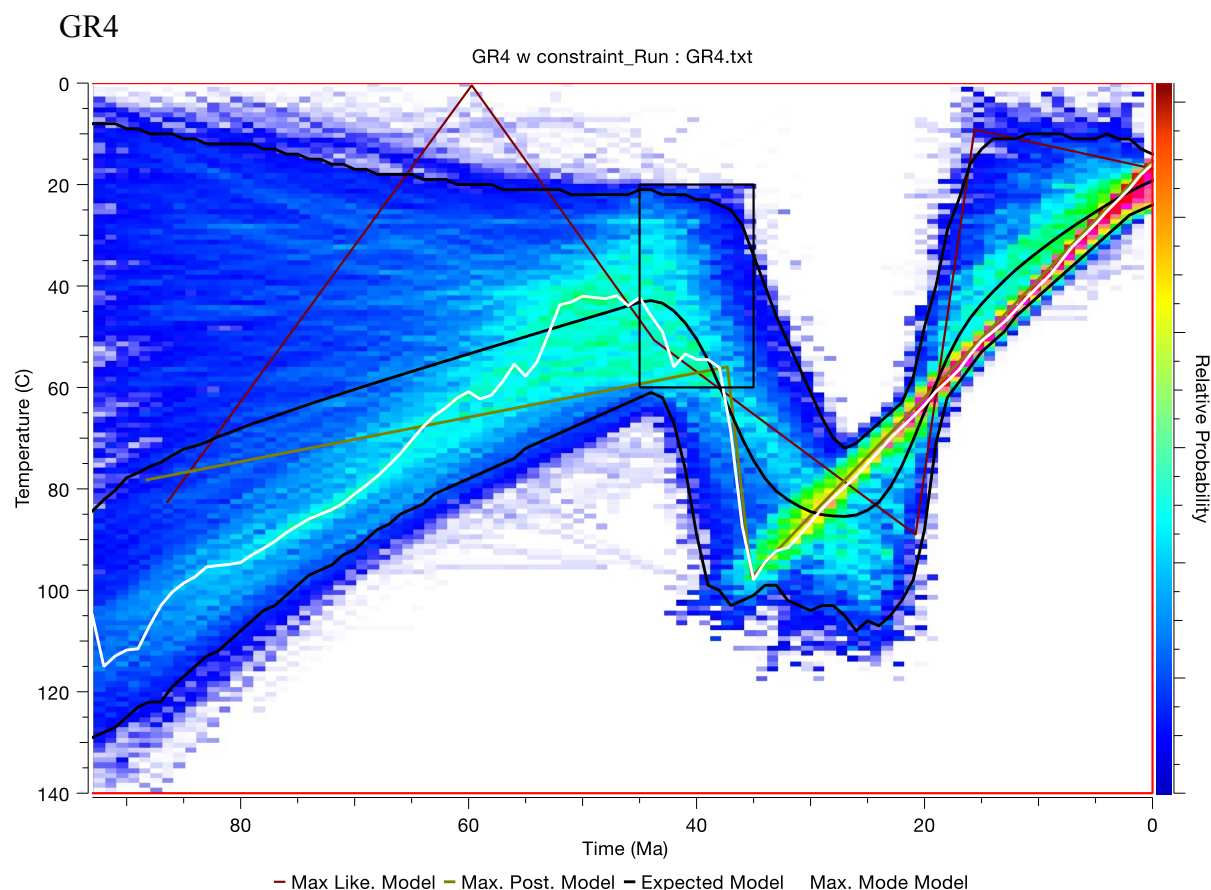


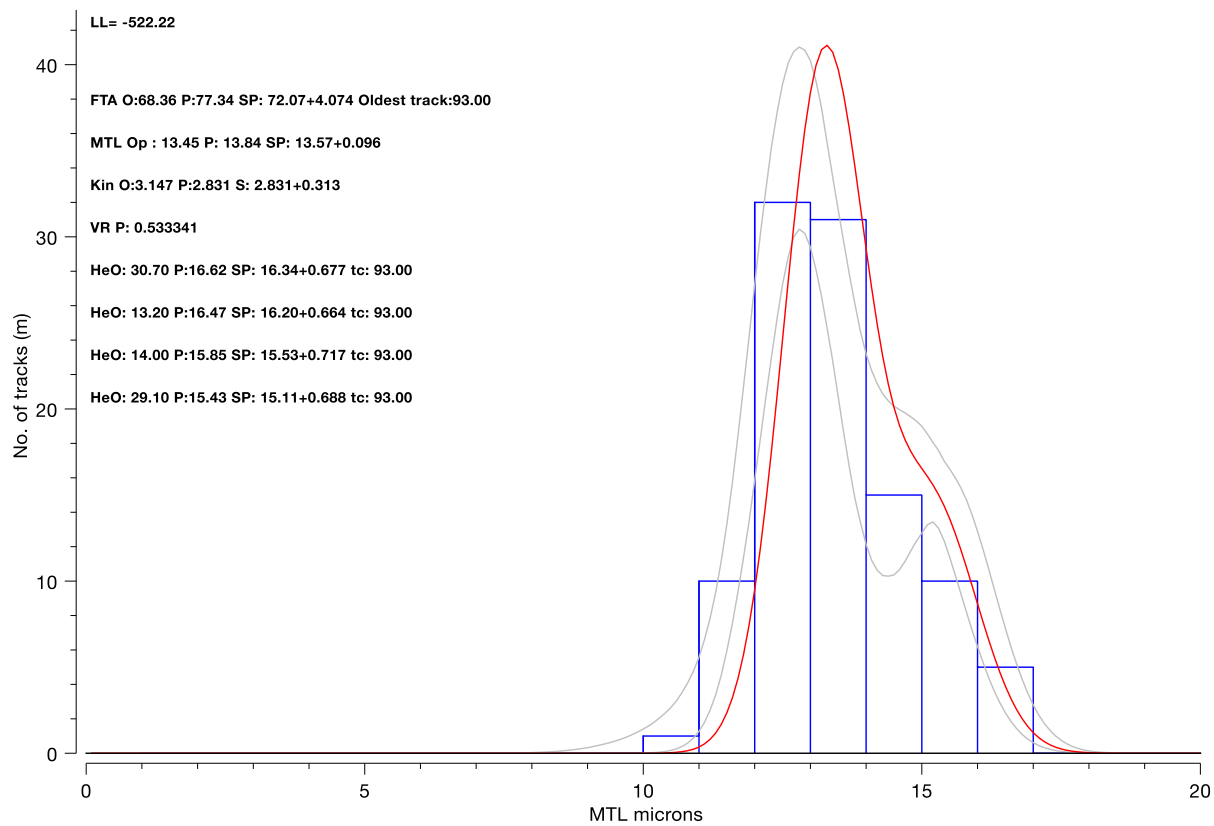
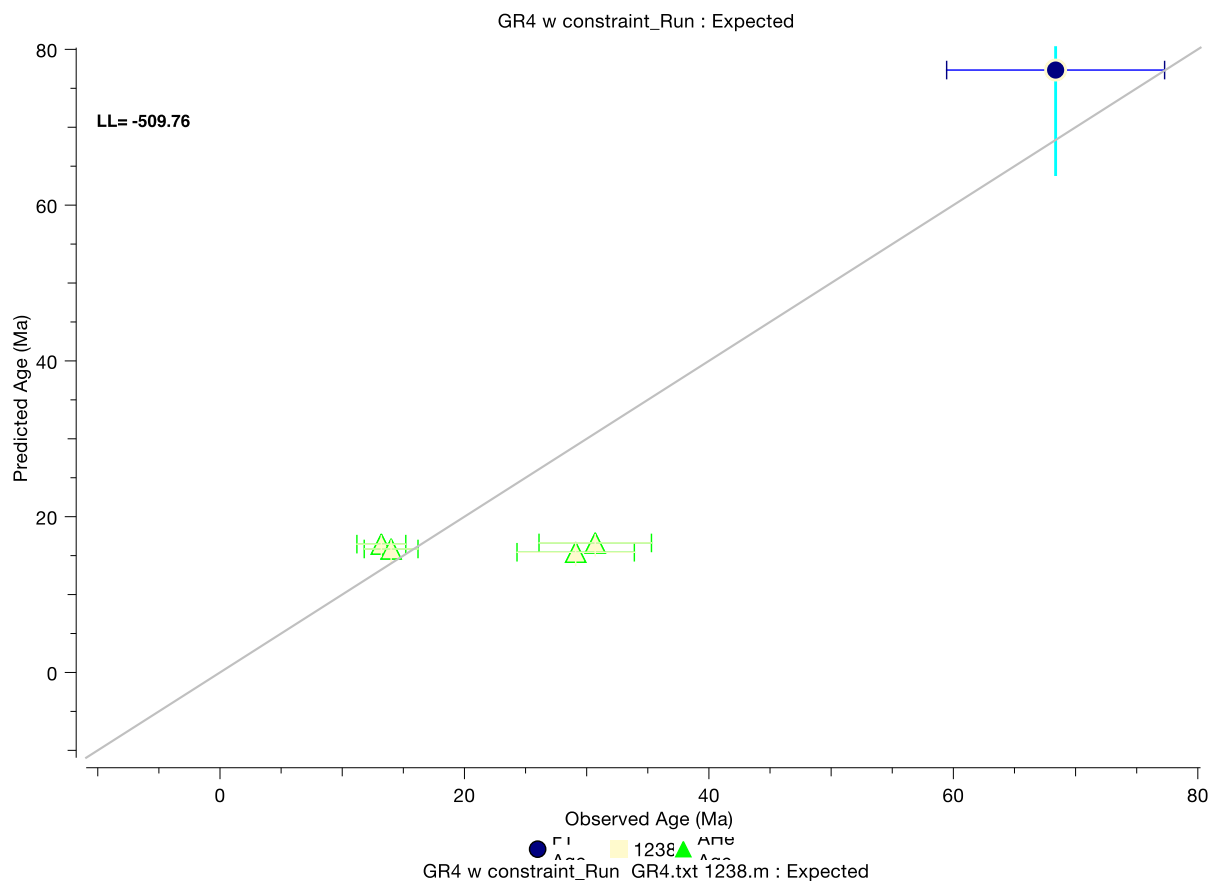




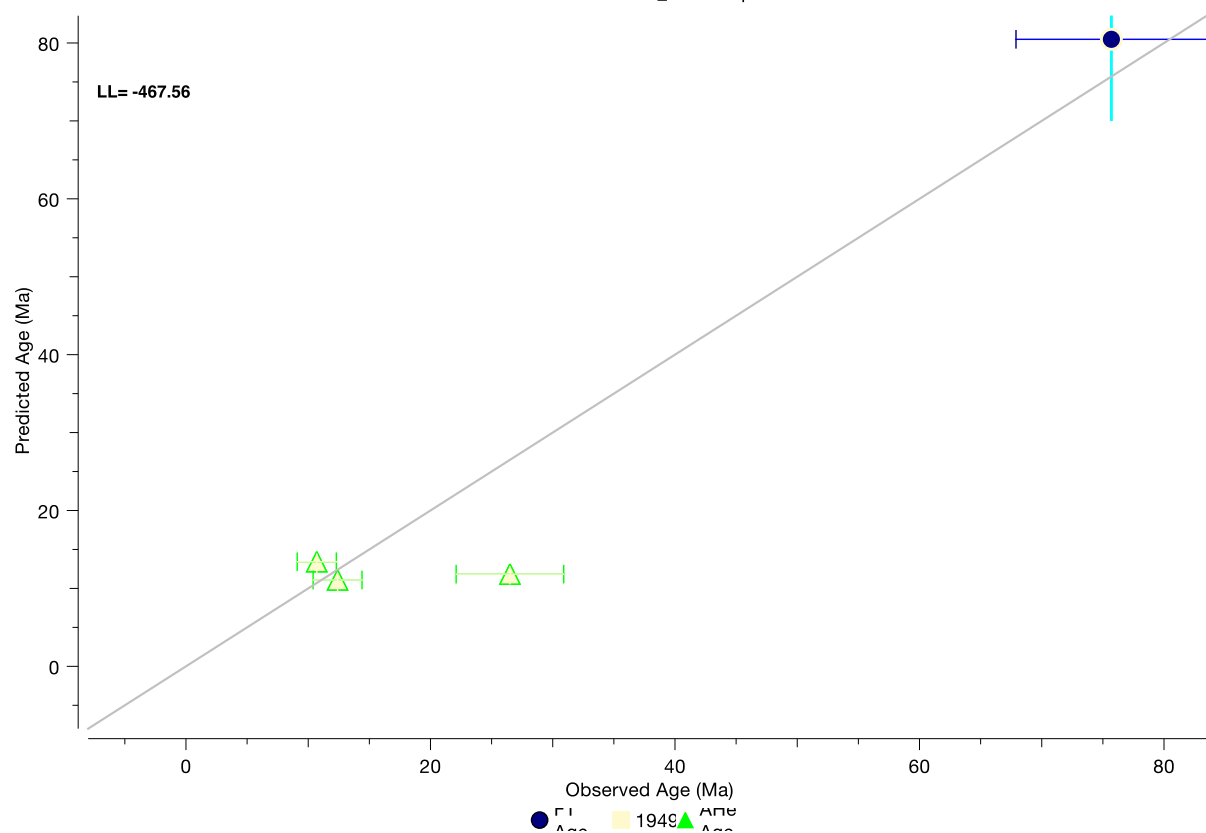
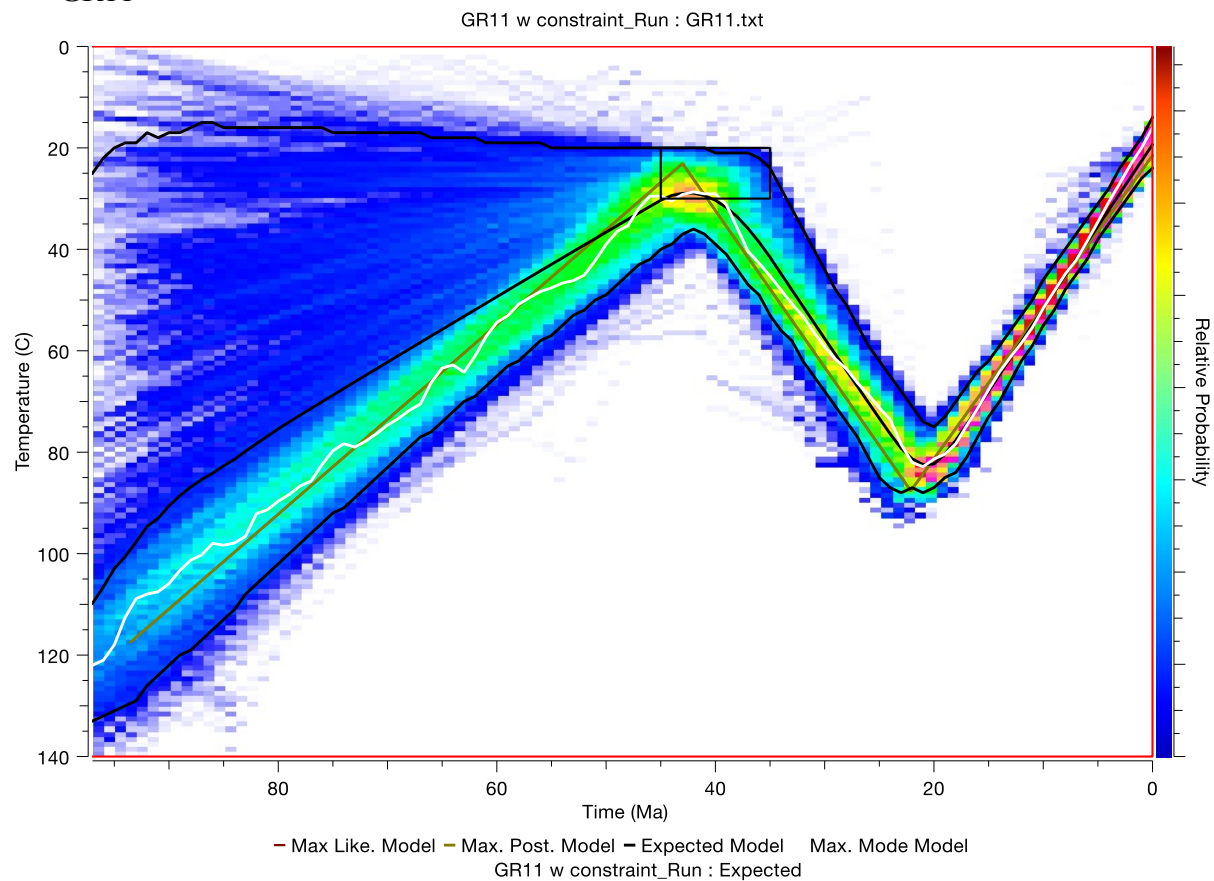
Supplementary Data B: Individual sample thermal history models with t-T constraints

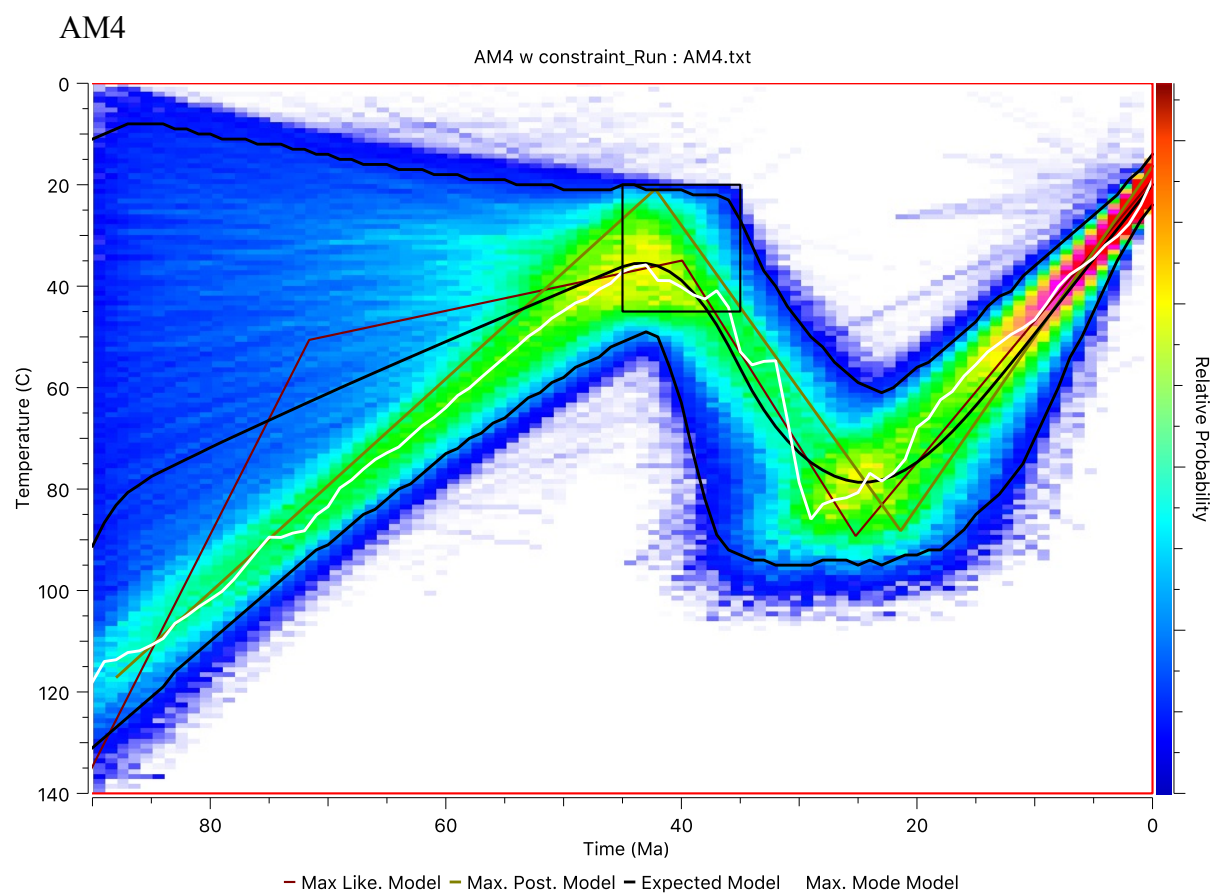
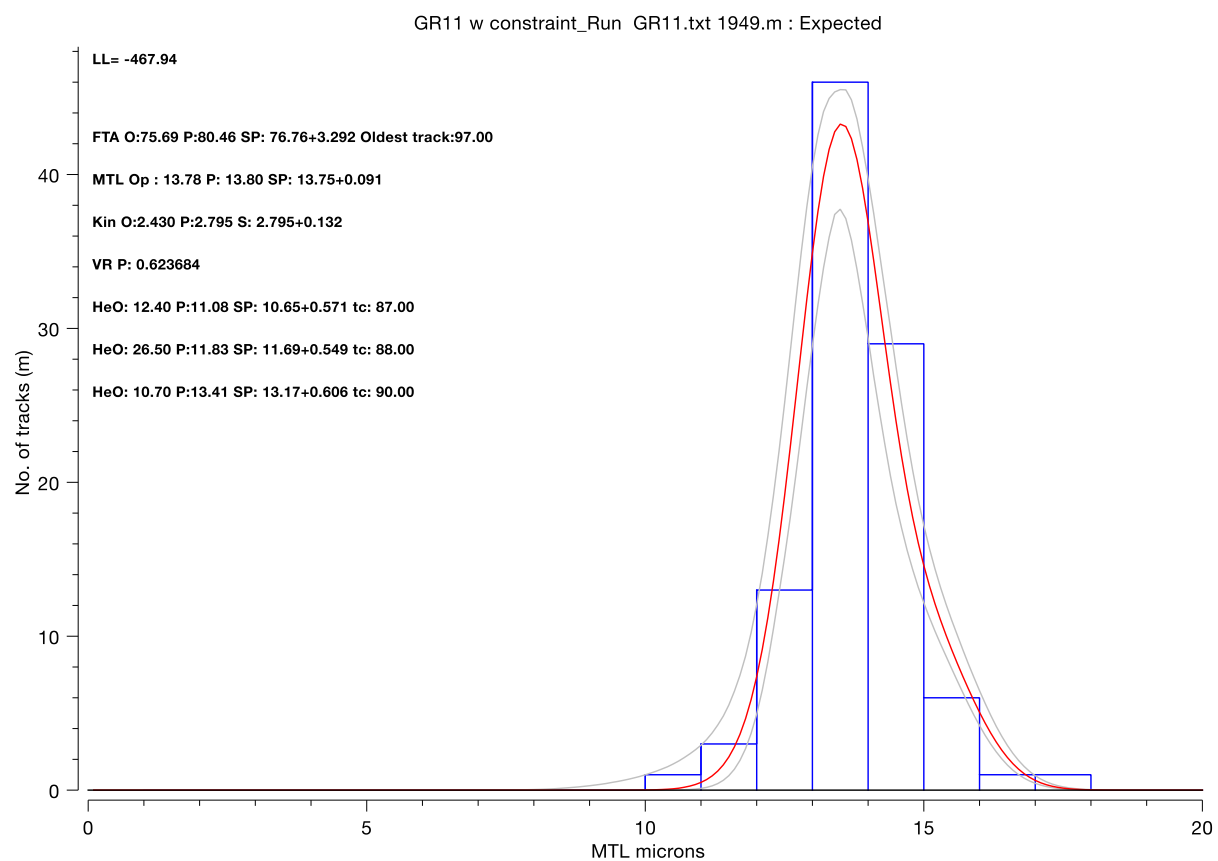
Thermal history models displayed below were generated in the software *QTQt* using AFT and AHe data (when available), following the procedure outlined in section 4.6. Only samples with a statistically robust number of confined track length measurements (> 60) were modelled. A first time-temperature constraint (black box in thermal history models) corresponds to the emplacement of the Eocene Amaro Gamo Basalts (45-35 Ma; *Ebinger et al.*, 1993) overlying Precambrian basement rocks in the Beto Basin and Amaro Horst regions. The temperature range associated with this t-T constraint varied between samples, depending on the structural depth of the basement sample in relation to the overlying base Amaro Gamo Basalts and a conservative paleogeothermal gradient of $30\text{ }^{\circ}\text{C}/\text{km}$. This initial t-T constraint was not applied to Amaro Horst 2nd traverse samples, as any once-overlying Amaro Gamo Basalts have been completely eroded away. Thermal history models of individual Amaro Horst 2nd traverse samples without the Eocene t-T constraint can be found in Supplementary Data A. A second t-T constraint requires samples to be at near-surface temperatures ($20 \pm 10\text{ }^{\circ}\text{C}$) by the present-day. Plots of observed versus predicted AFT ages and confined track distributions are displayed below each thermal history model. All plots were generated by *QTQt*. For descriptions of the various plots, see the caption for Figure 4.7.

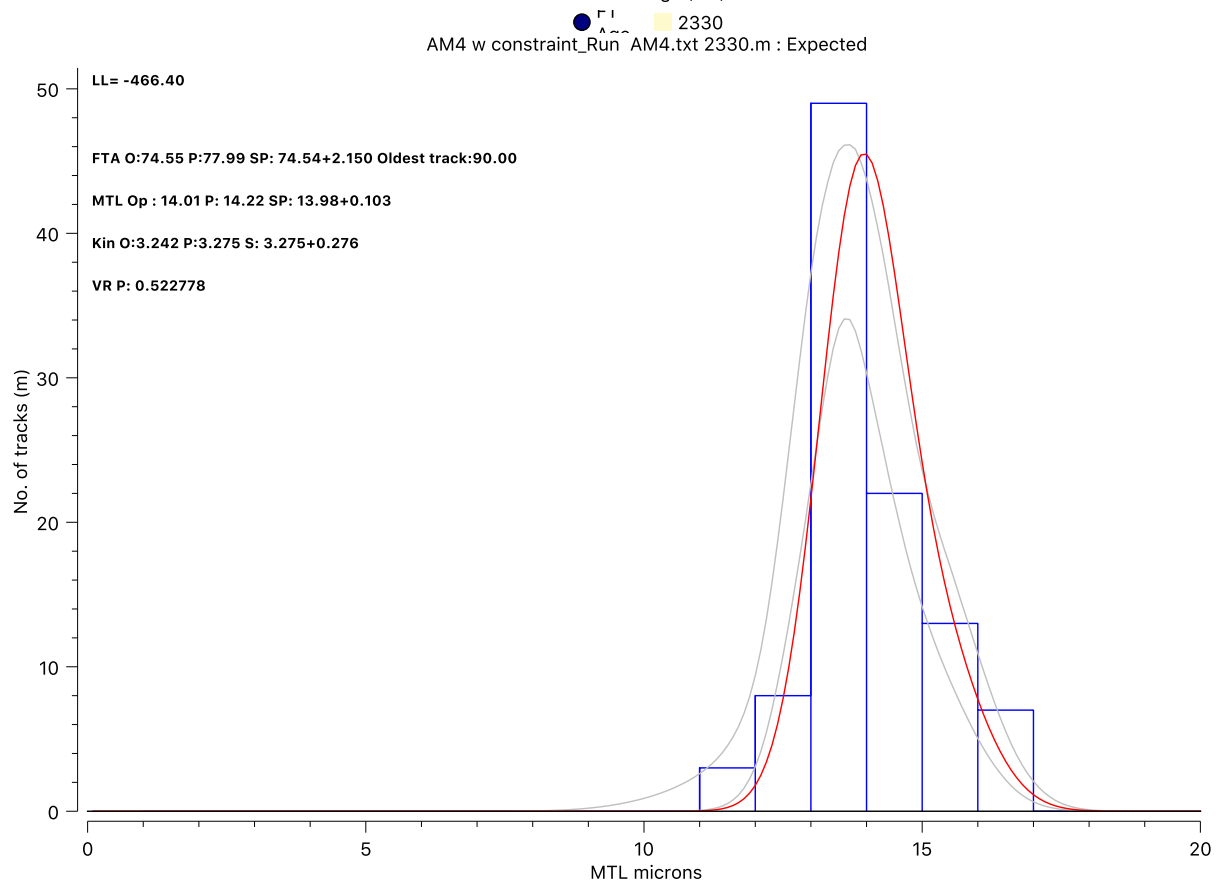
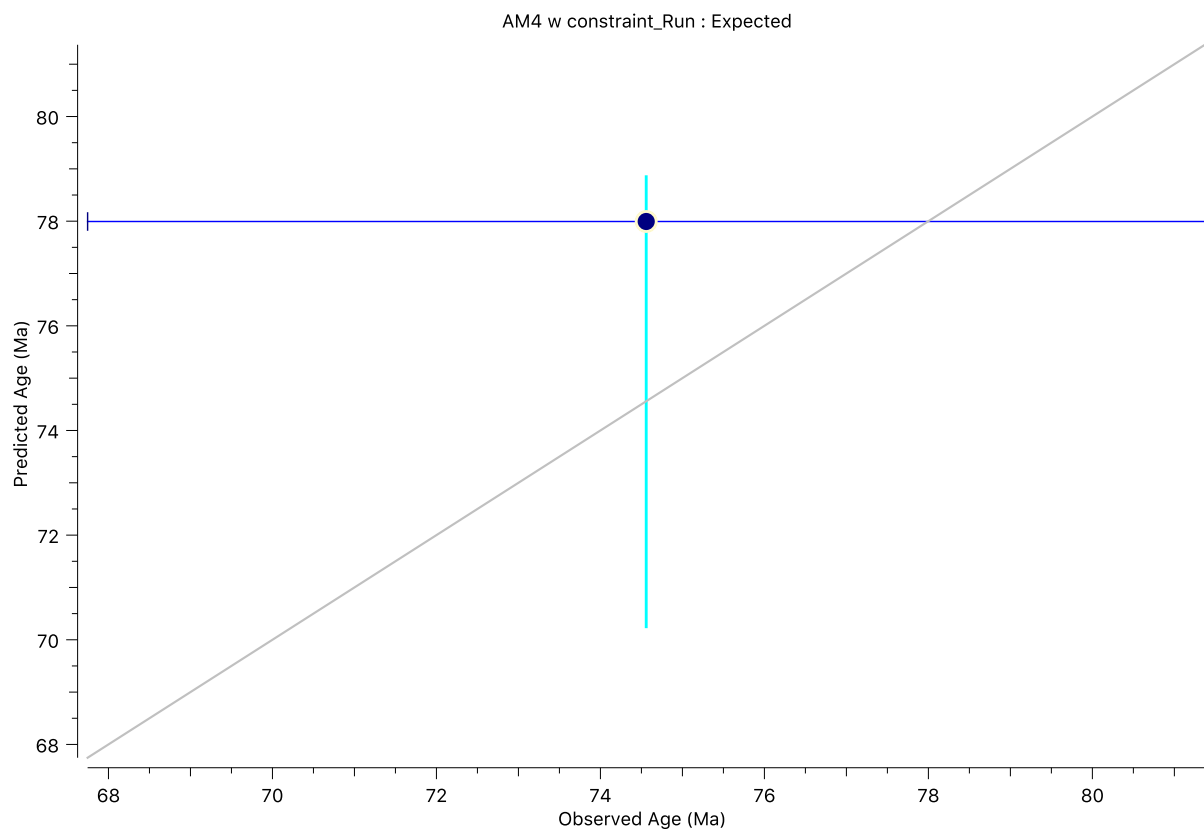




GR11

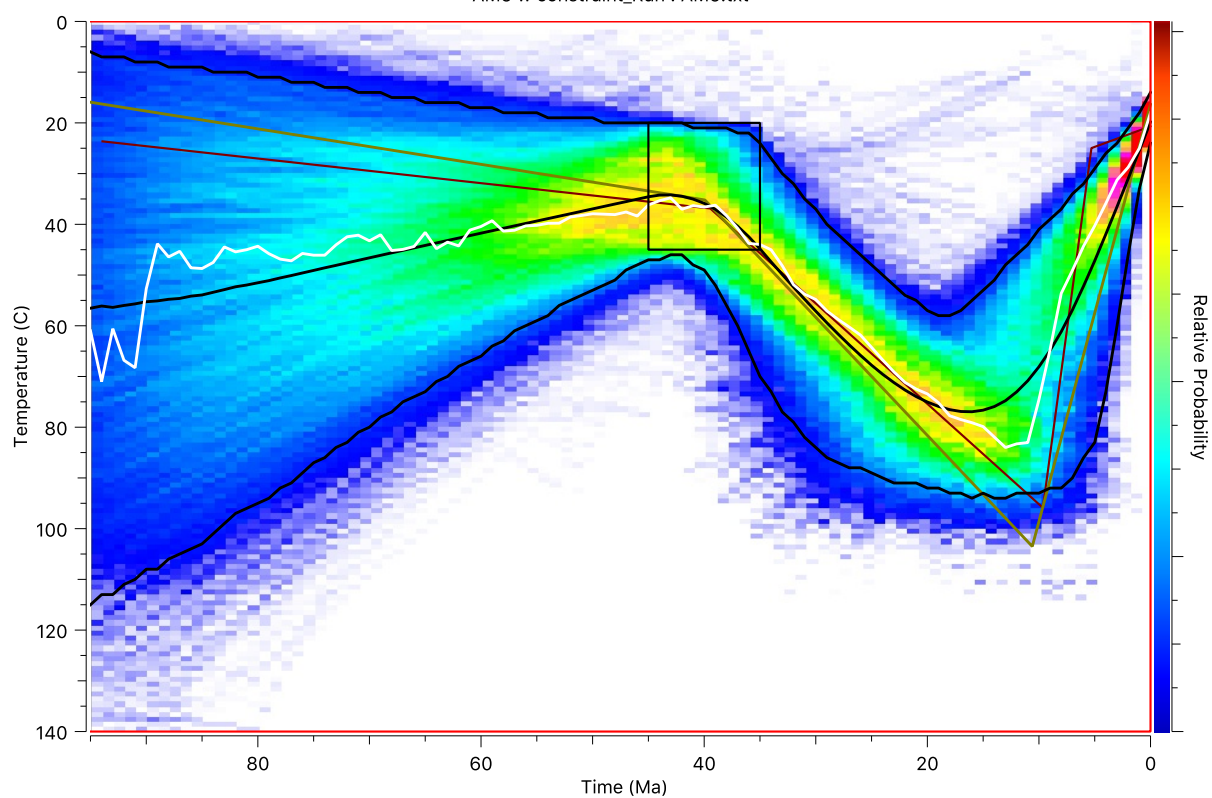




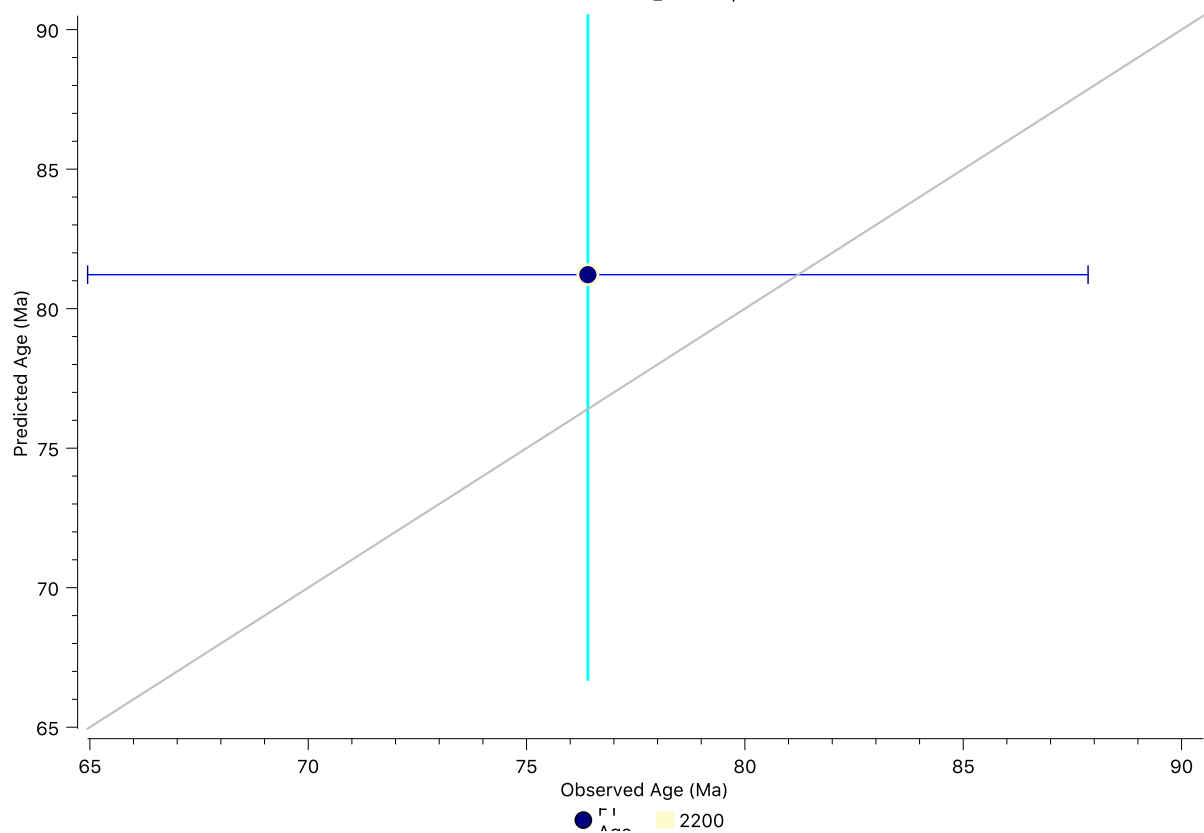


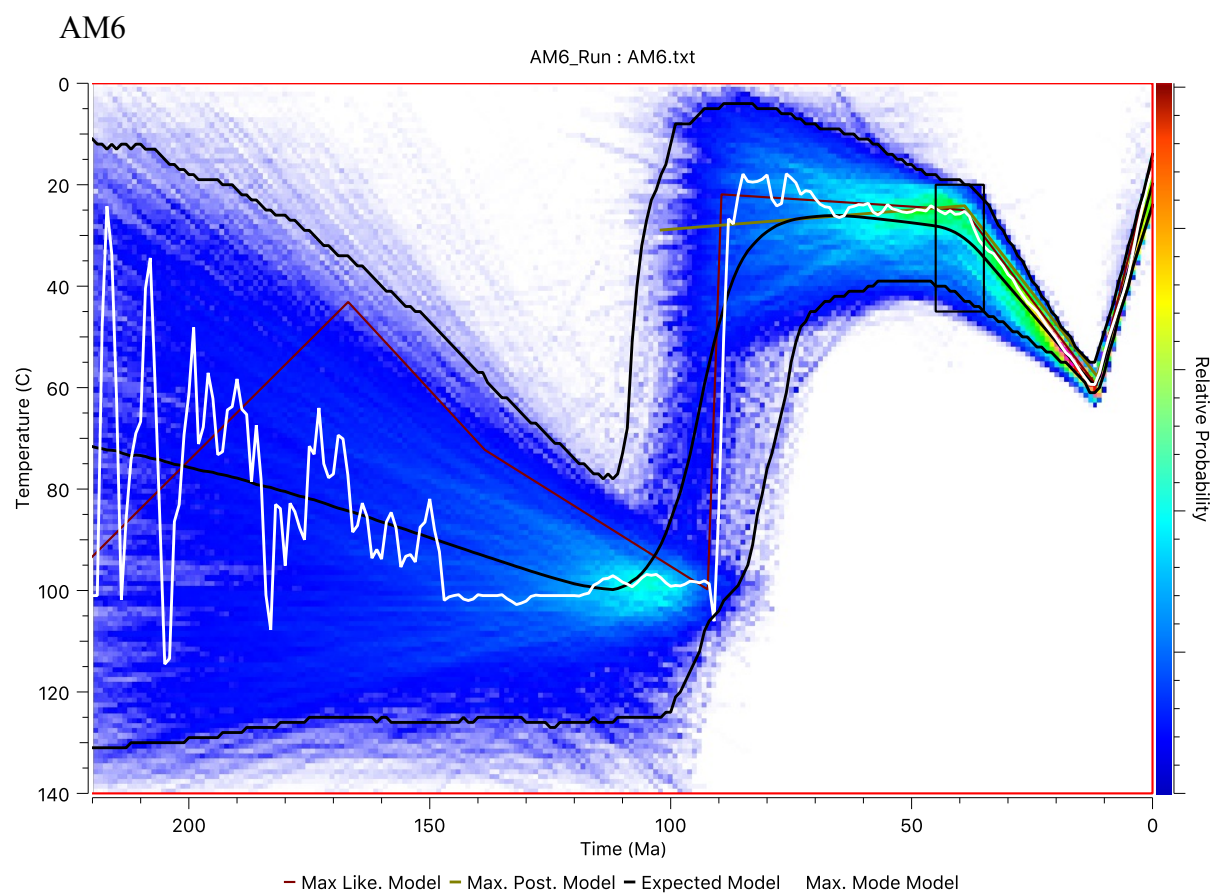
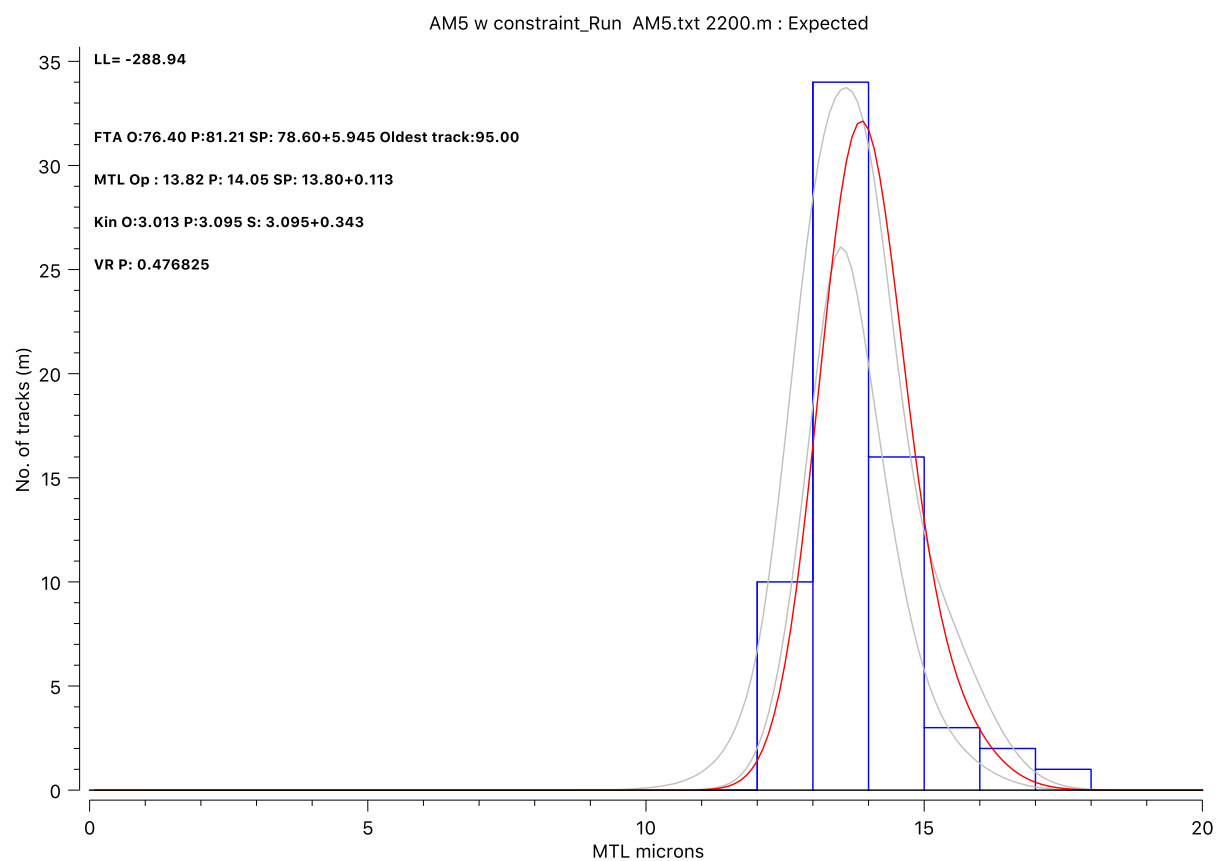
AM5

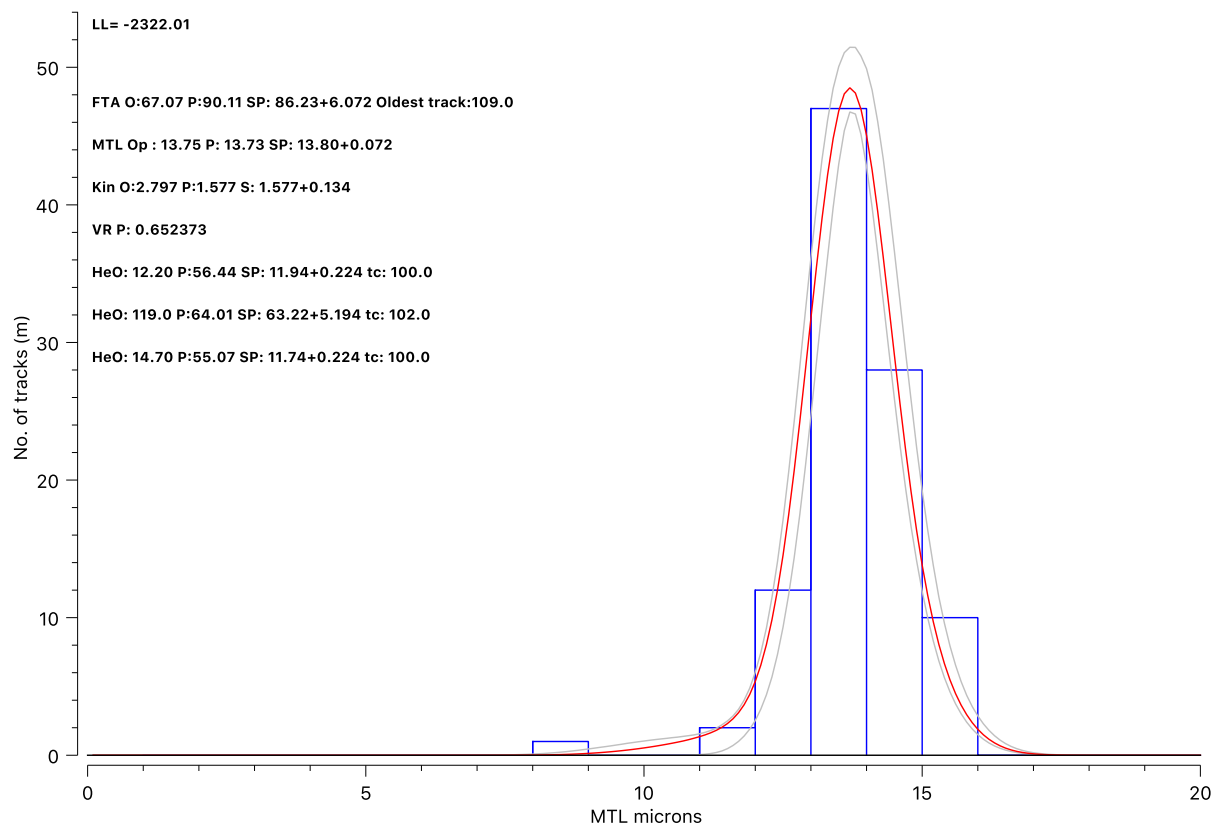
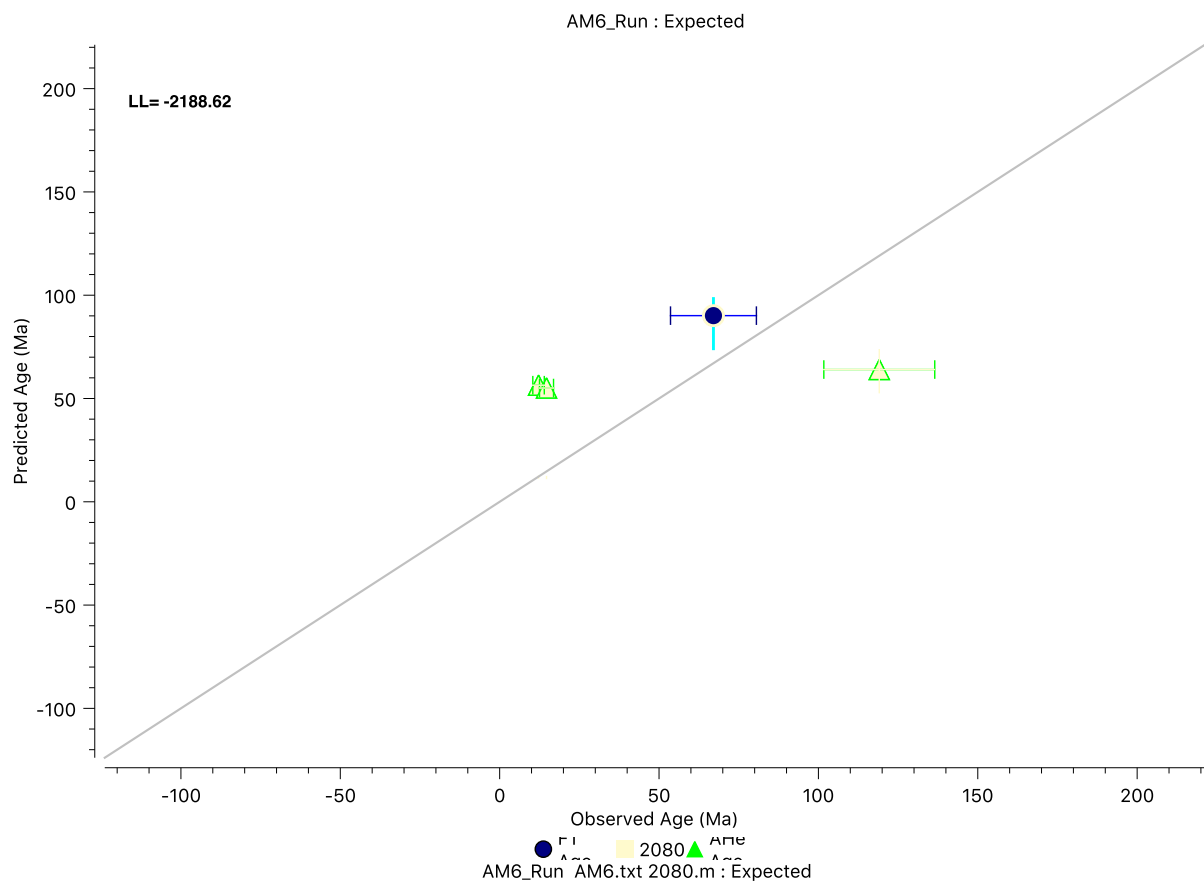
AM5 w constraint_Run : AM5.txt



AM5 w constraint_Run : Expected

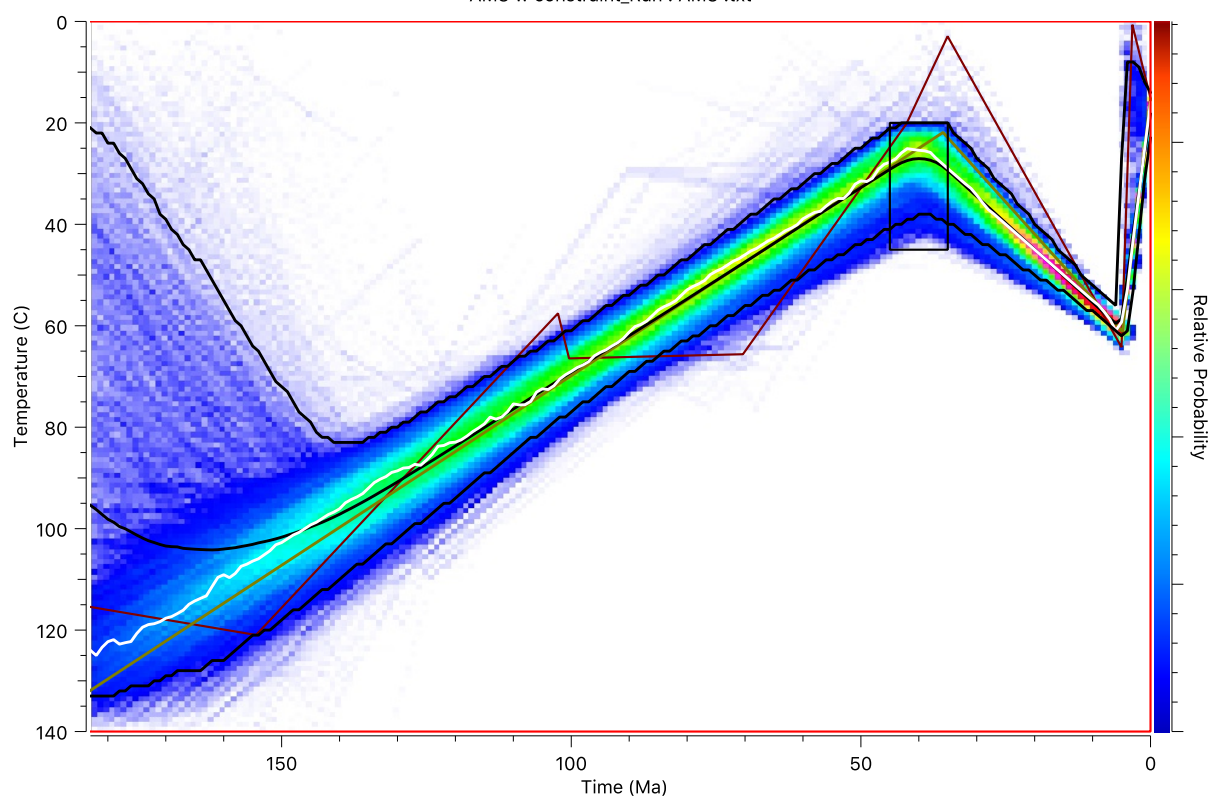




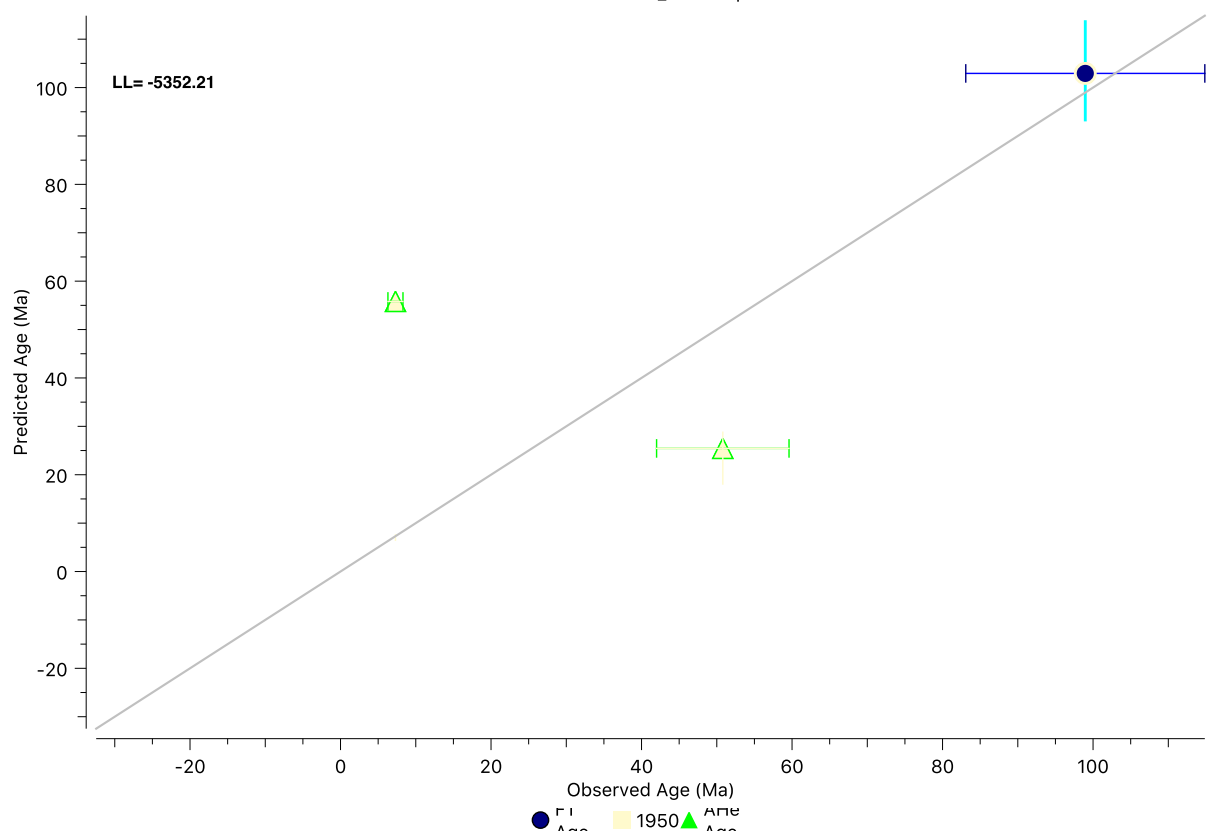


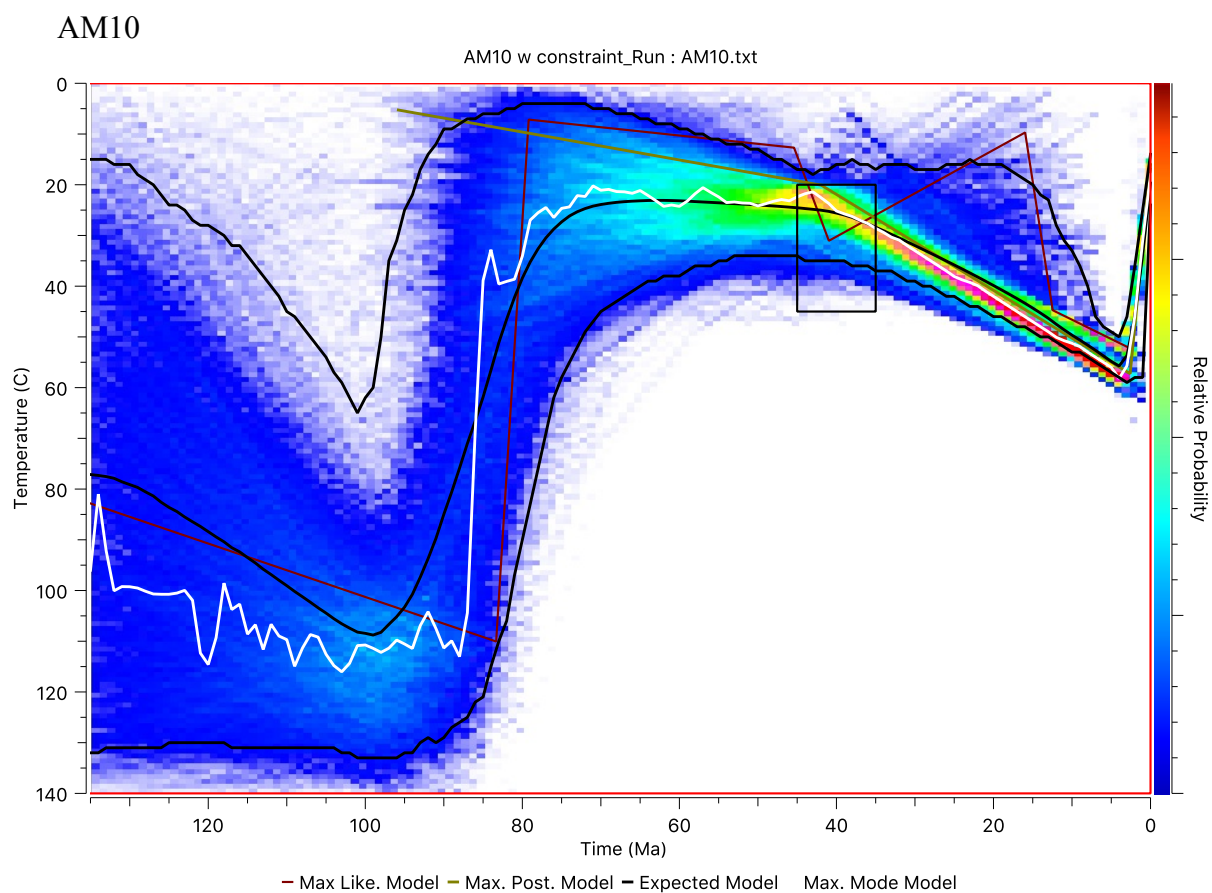
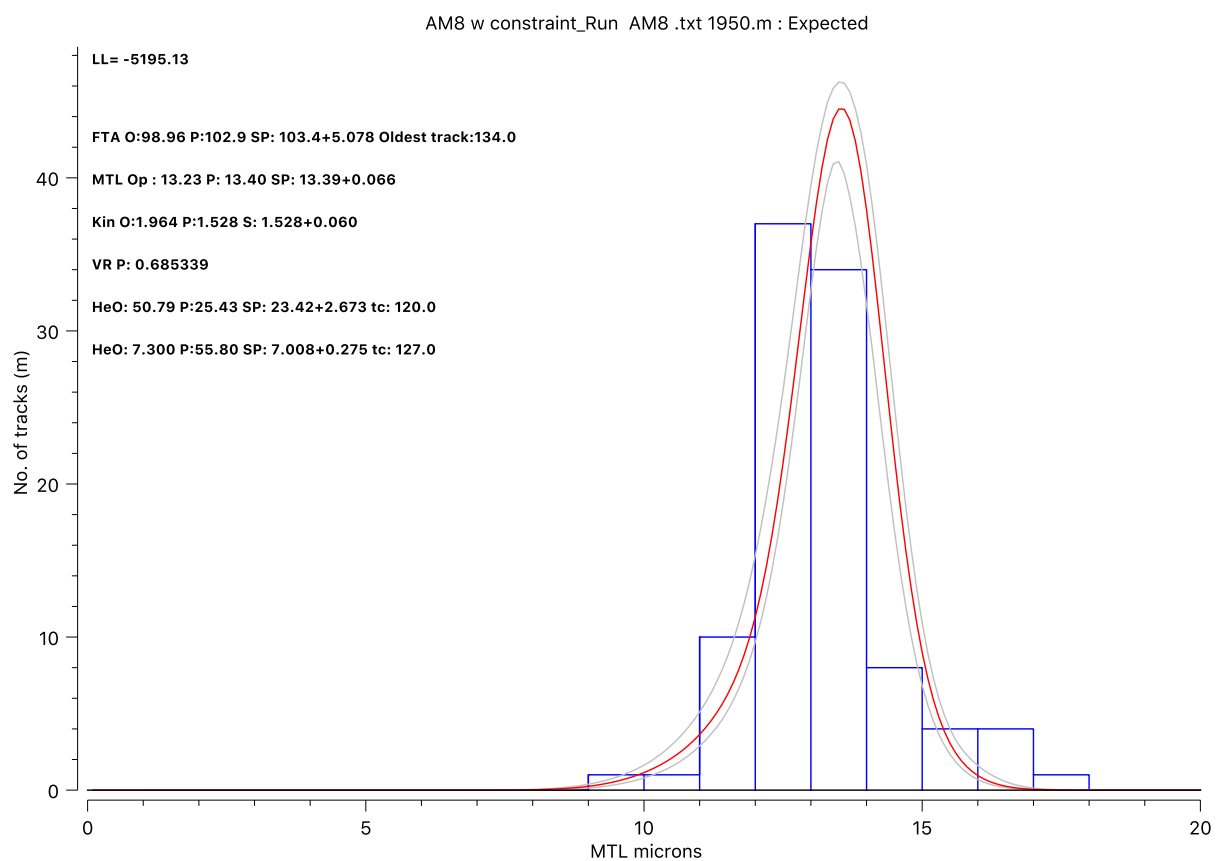
AM8

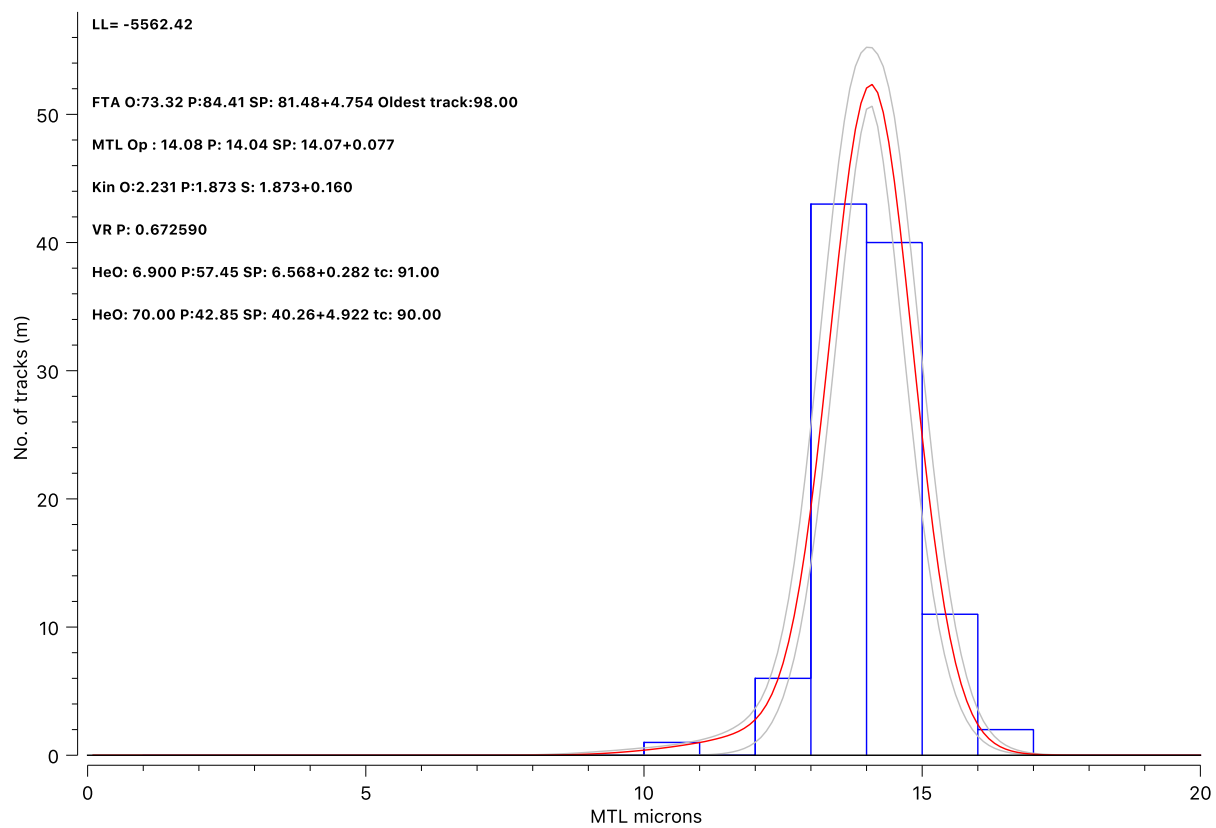
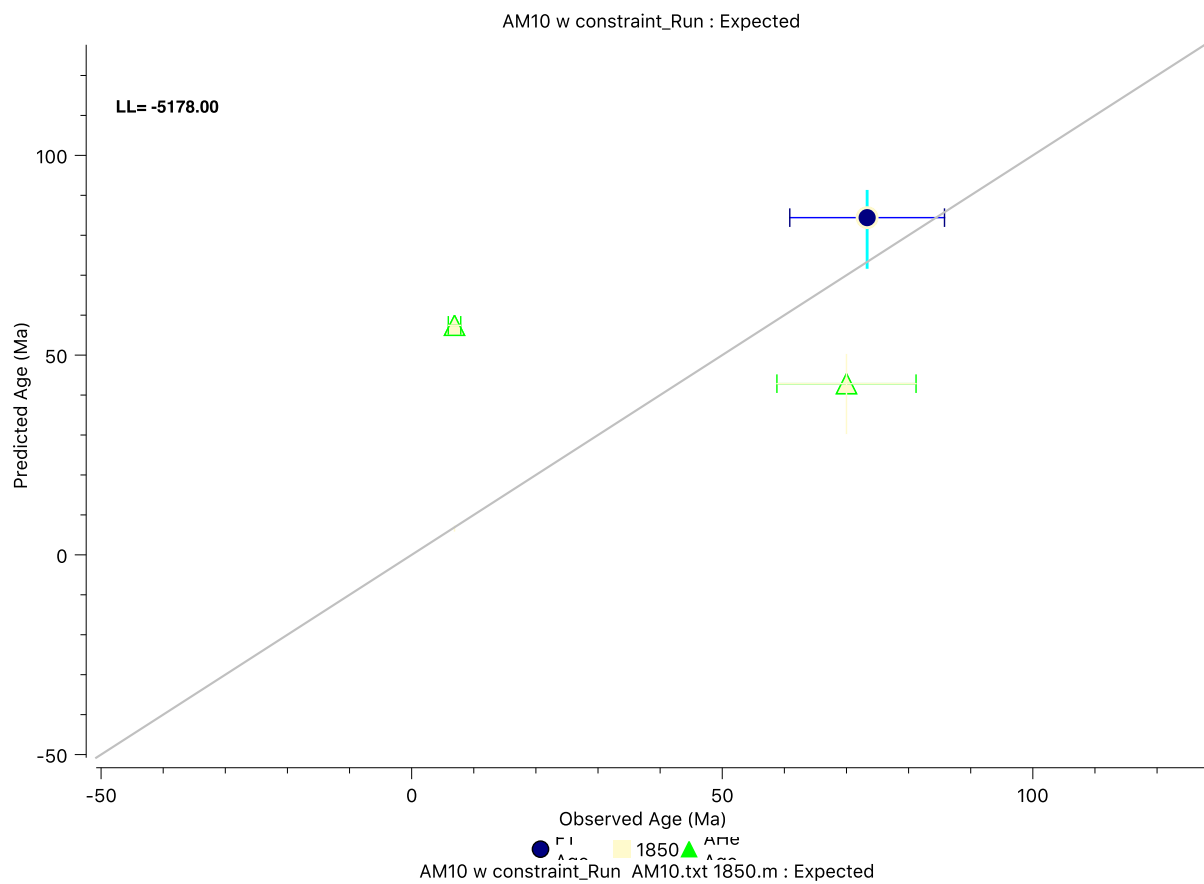
AM8 w constraint_Run : AM8 .txt



AM8 w constraint_Run : Expected



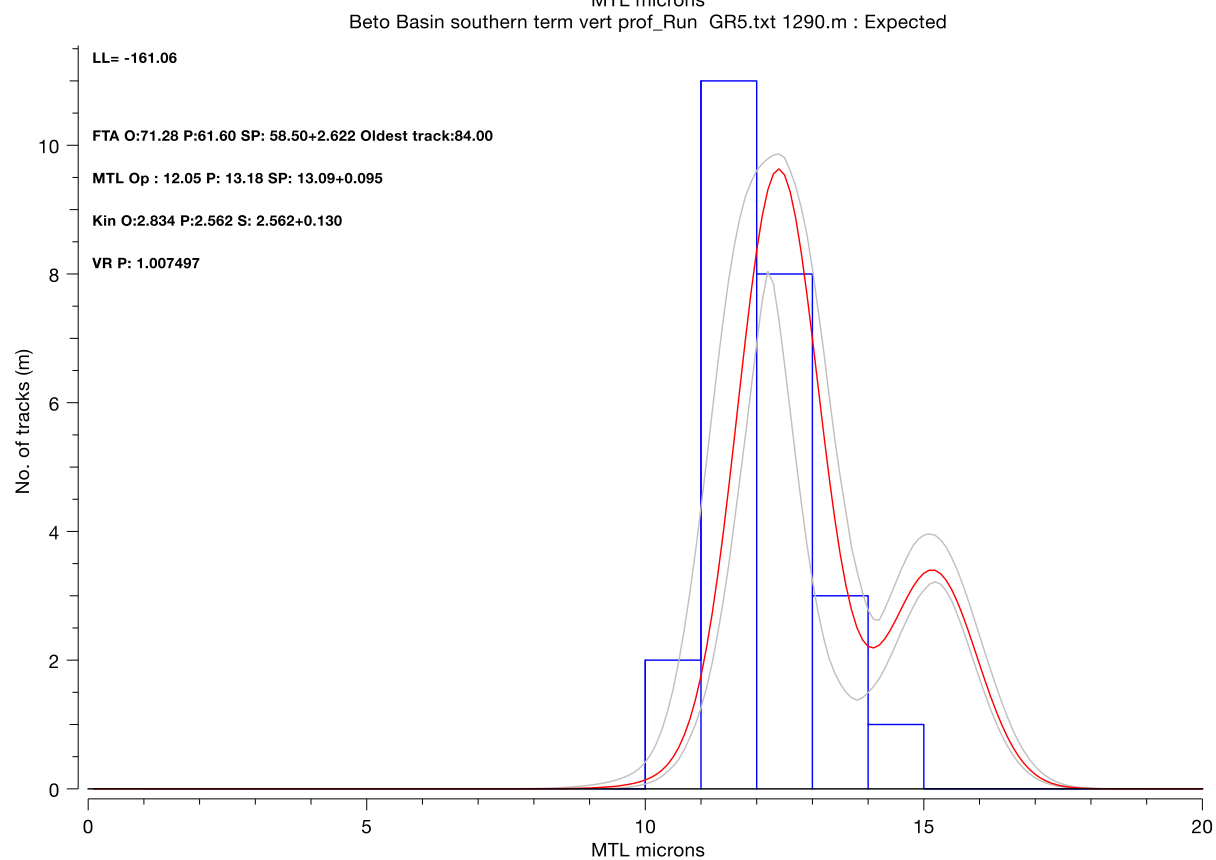
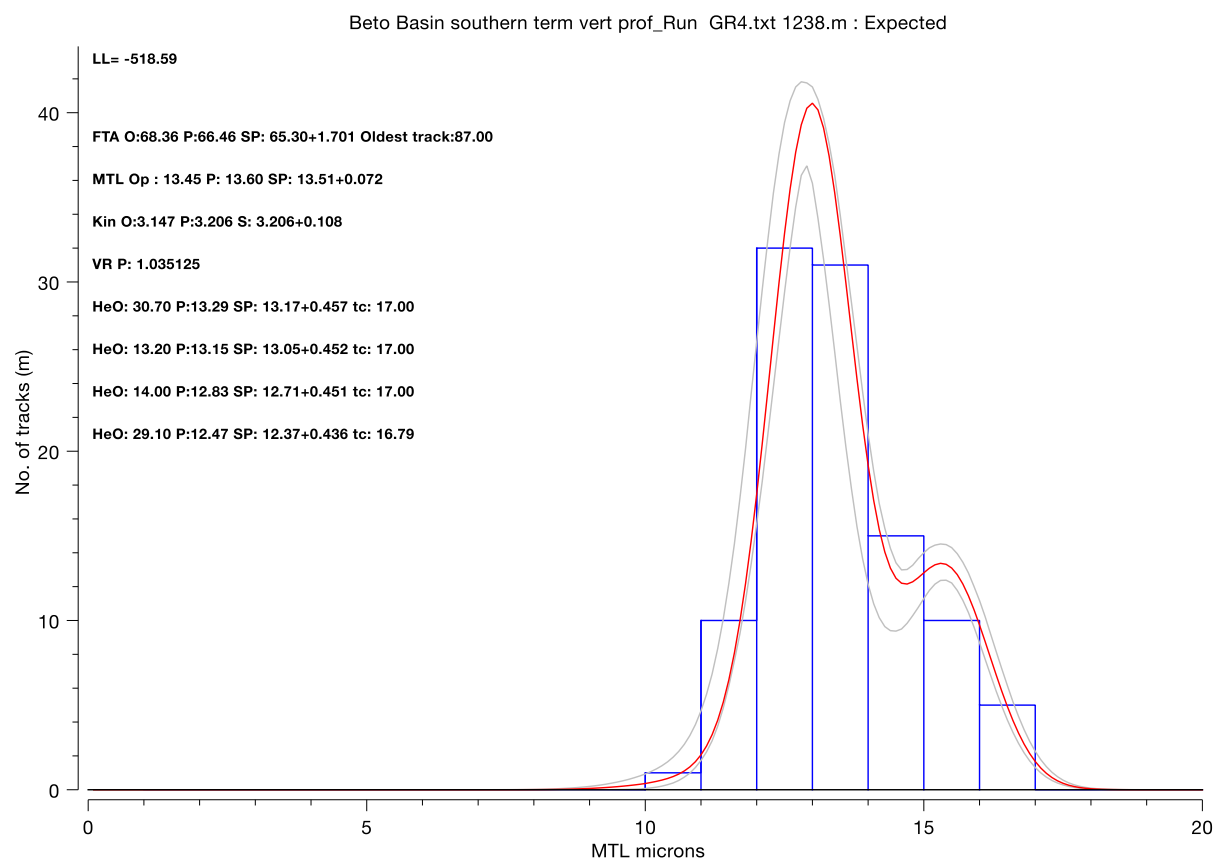




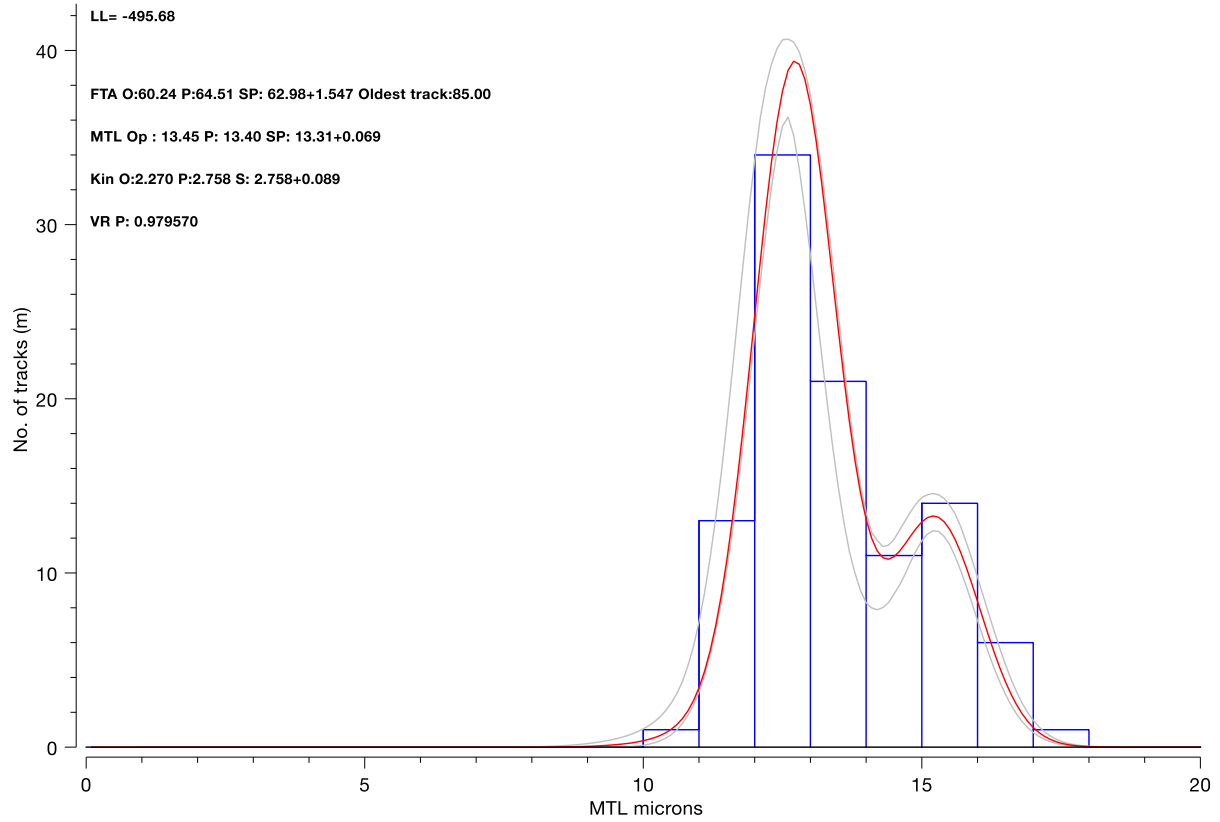
Supplementary Data C: Observed versus predicted confined track distributions for vertical profile thermal history models

Plots of observed confined fission track distributions overlain by the predicted distribution (red curve) and its corresponding 95% confidence interval (grey curves) for individual samples included in the vertical profile thermal history models displayed in Figure 4.7. All samples, including those with a less robust number of confined track length measurements (> 60), were modelled for each vertical profile. Samples with no confined track data (i.e. GR8) were omitted during thermal history modelling. A first time-temperature constraint (black box in thermal history models) corresponds to the emplacement of the Eocene Amaro Gamo Basalts (45-35 Ma; *Ebinger et al.*, 1993) overlying Precambrian basement rocks in the Beto Basin and Amaro Horst regions. The temperature range associated with this t-T constraint varied between samples, depending on the structural depth of the basement sample in relation to the overlying base Amaro Gamo Basalts and a conservative paleogeothermal gradient of $30\text{ }^{\circ}\text{C /km}$. This initial t-T constraint was not applied to Amaro Horst 2nd traverse samples, as any once-overlying Amaro Gamo Basalts have been completely eroded away. A second t-T constraint requires samples to be at near-surface temperatures ($20 \pm 10\text{ }^{\circ}\text{C}$) by the present-day. Thermal history models of individual samples included in these vertical profiles are presented in Supplementary Data A (without the Eocene t-T constraint) and Supplementary Data B (with the Eocene t-T constraint). A second t-T constraint requires samples to be at near-surface temperatures ($20 \pm 10\text{ }^{\circ}\text{C}$) by the present-day. All plots were generated by *QTQt*. See Section 4.6 for a more detailed description of the thermal history modelling protocol.

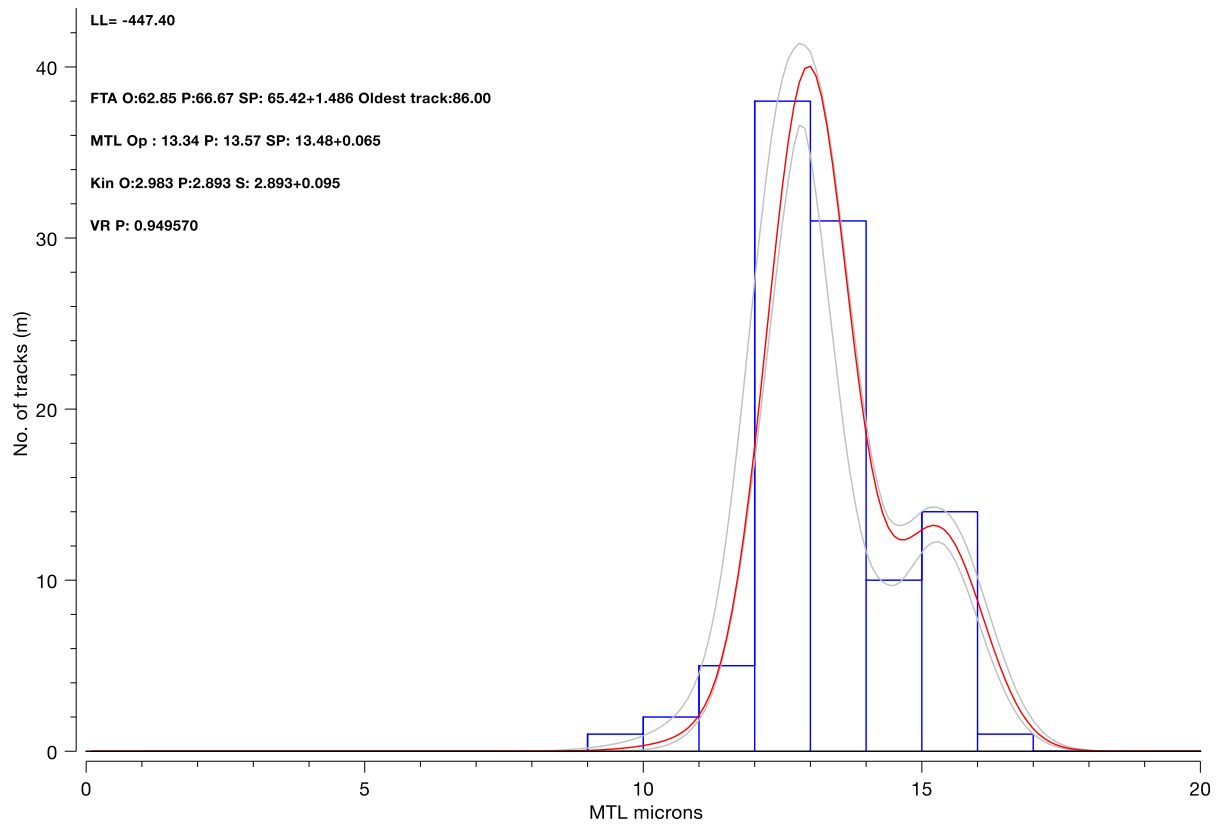
Beto Basin Southwest Margin: Samples GR4, 5, 6, 7, 8, 9, 10, and 11



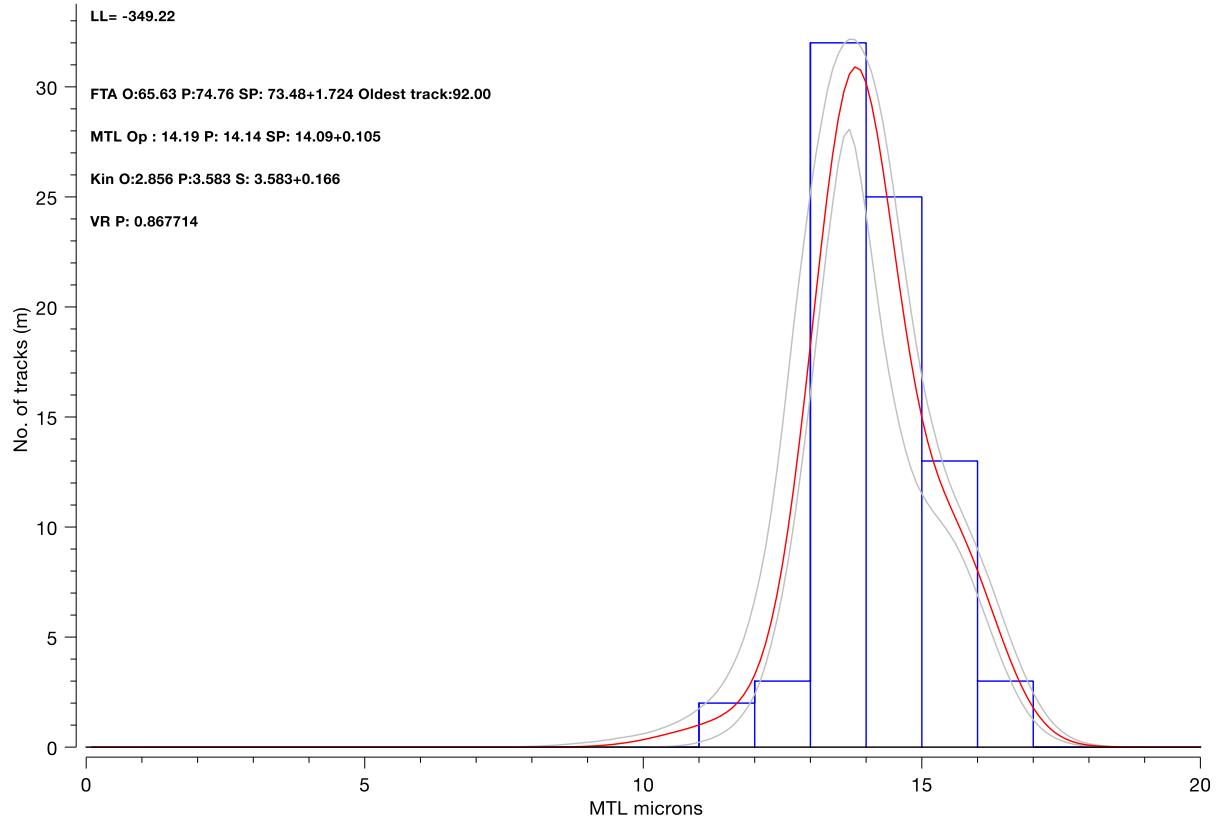
Beto Basin southern term vert prof_Run GR6.txt 1345.m : Expected



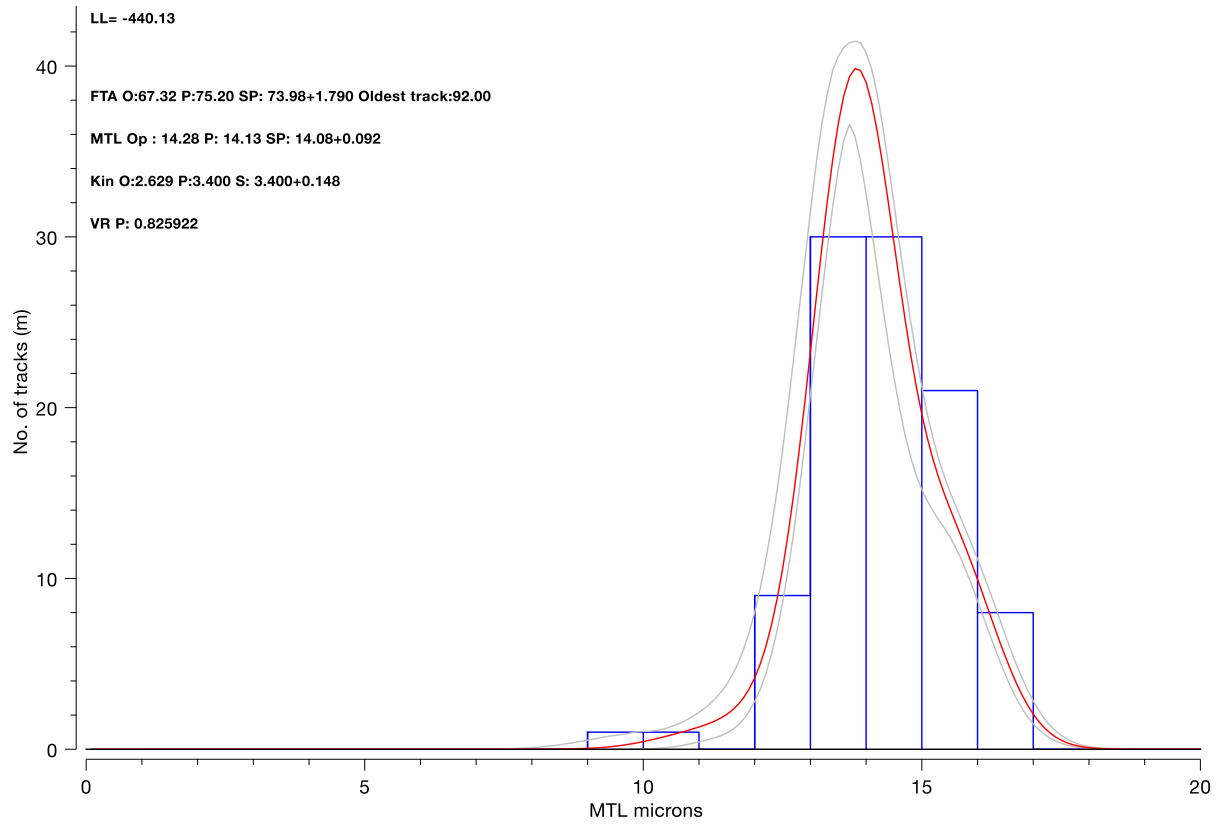
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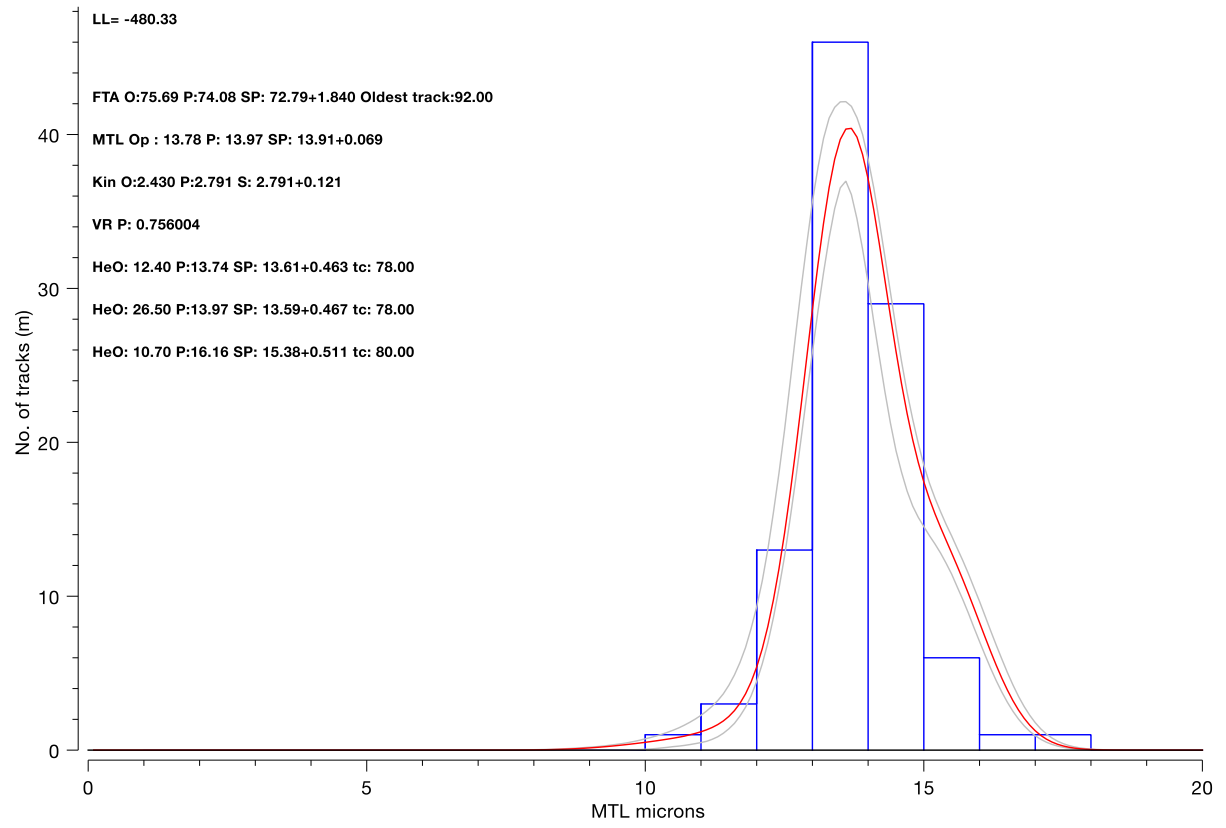
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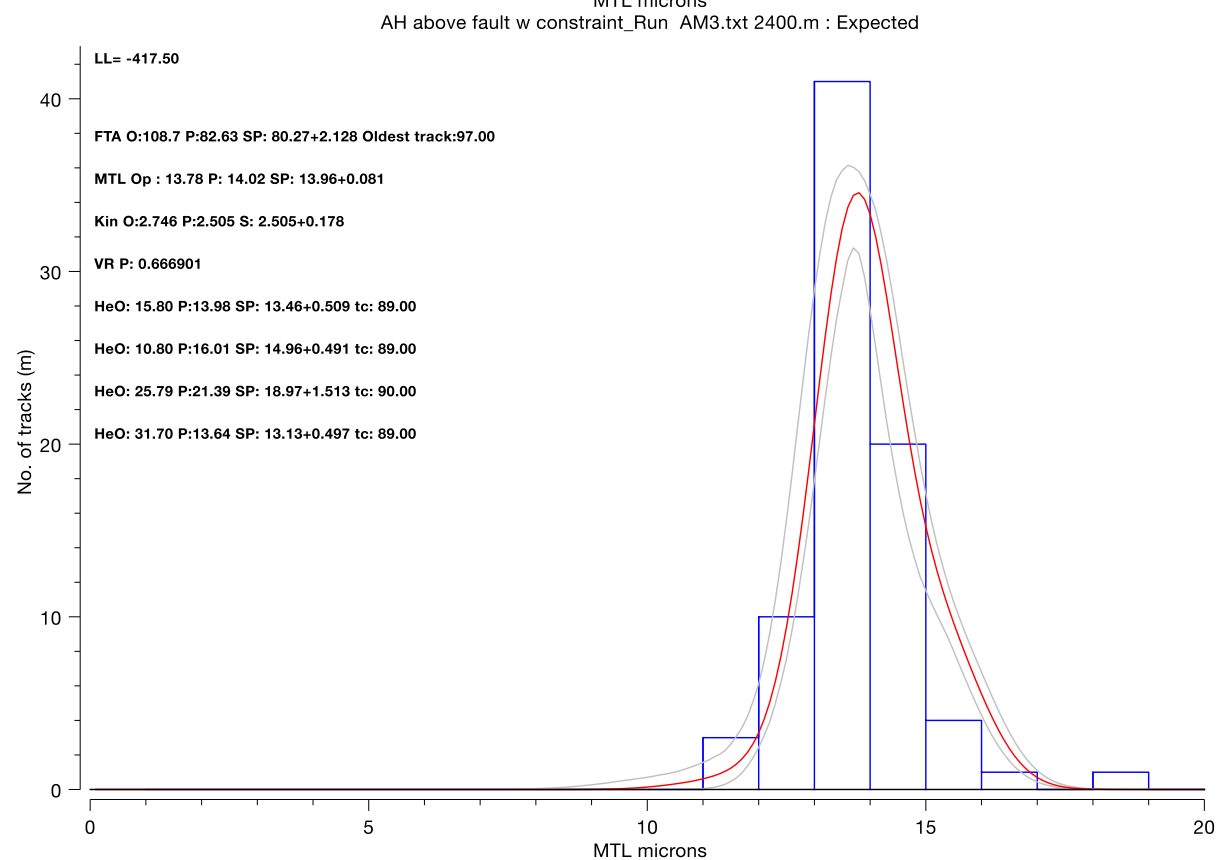
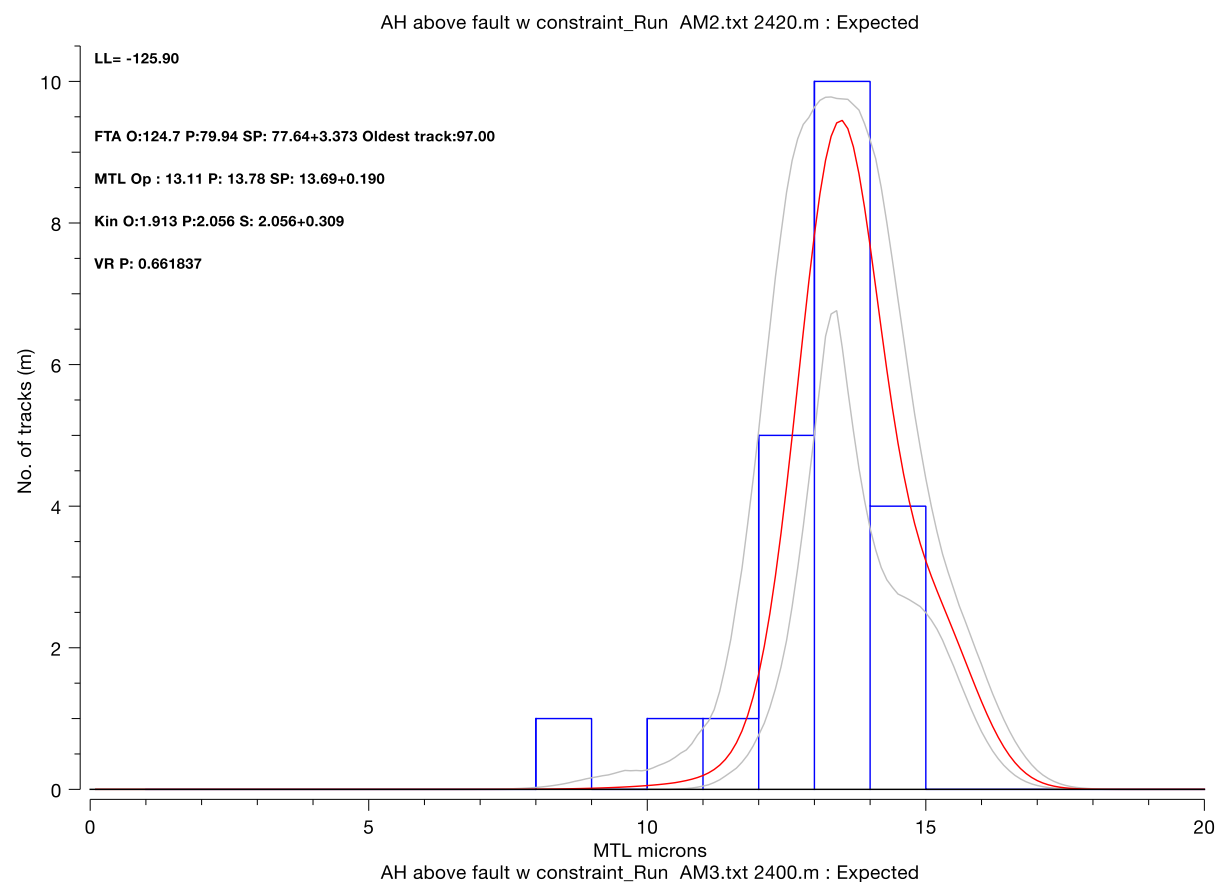
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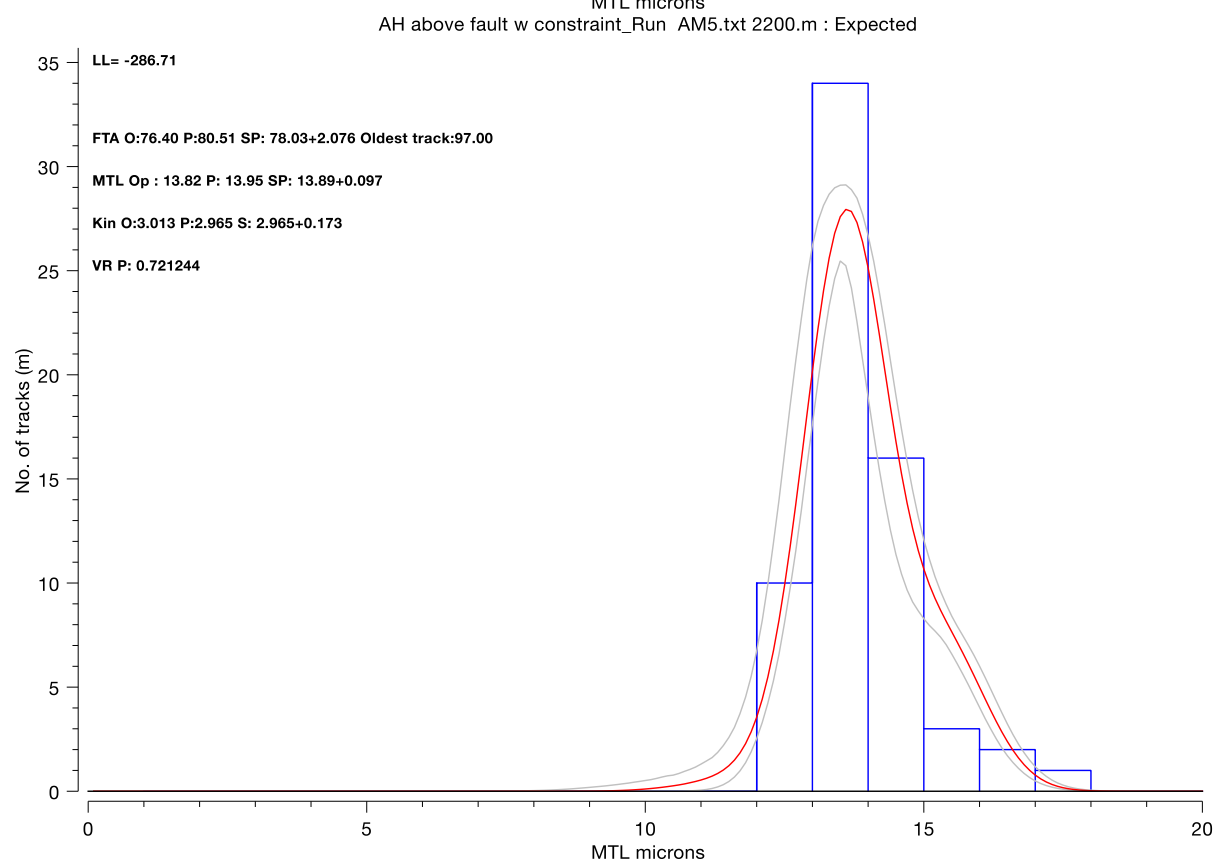
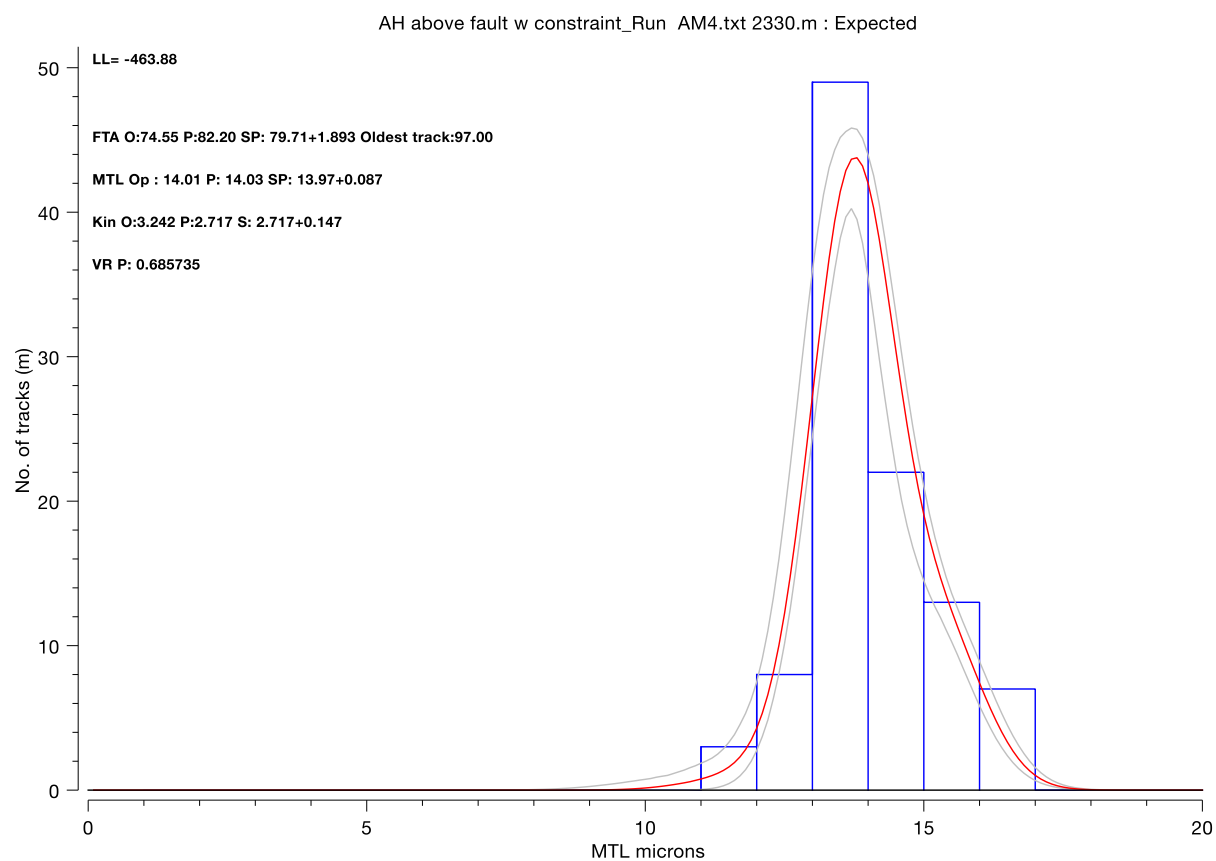


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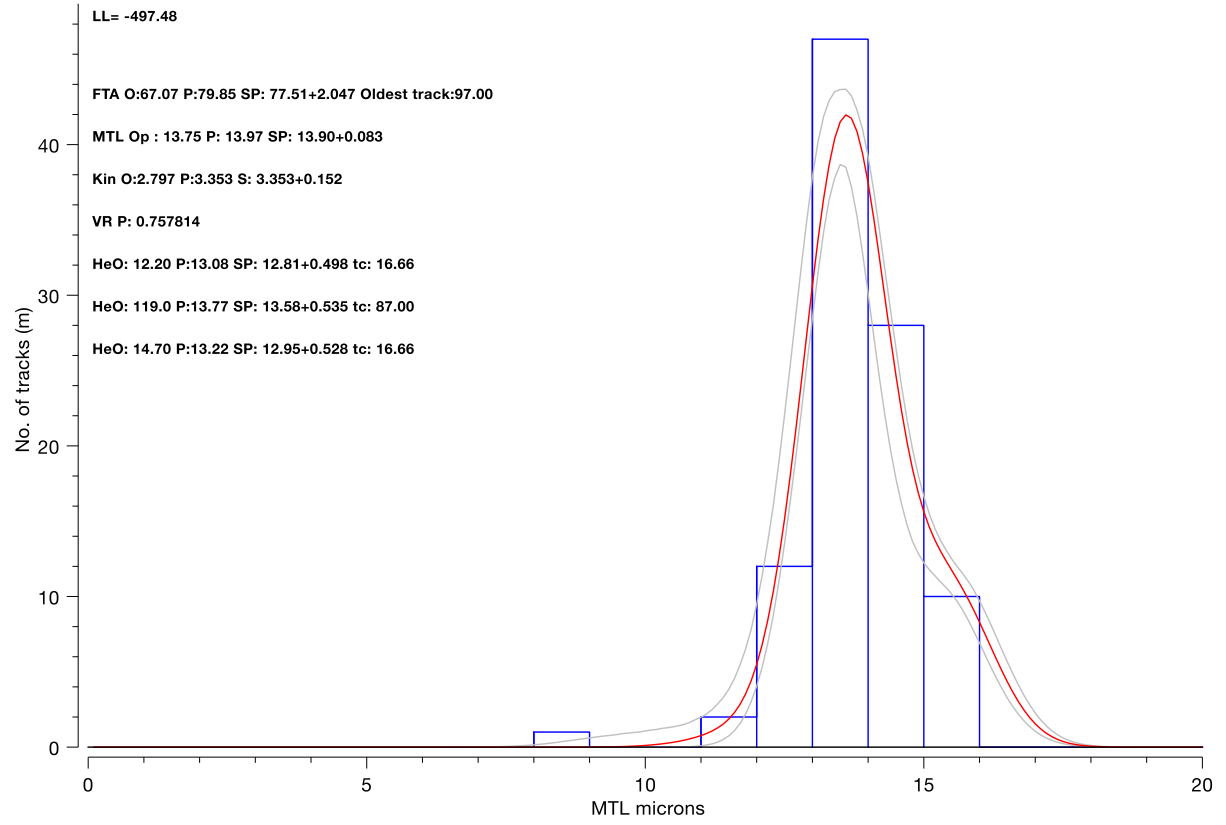


Amaro Horst 1st traverse, above secondary fault: Samples AM2-1st tr, AM3-1st tr, AM4-1st tr, AM5-1st tr, and AM6-1st tr

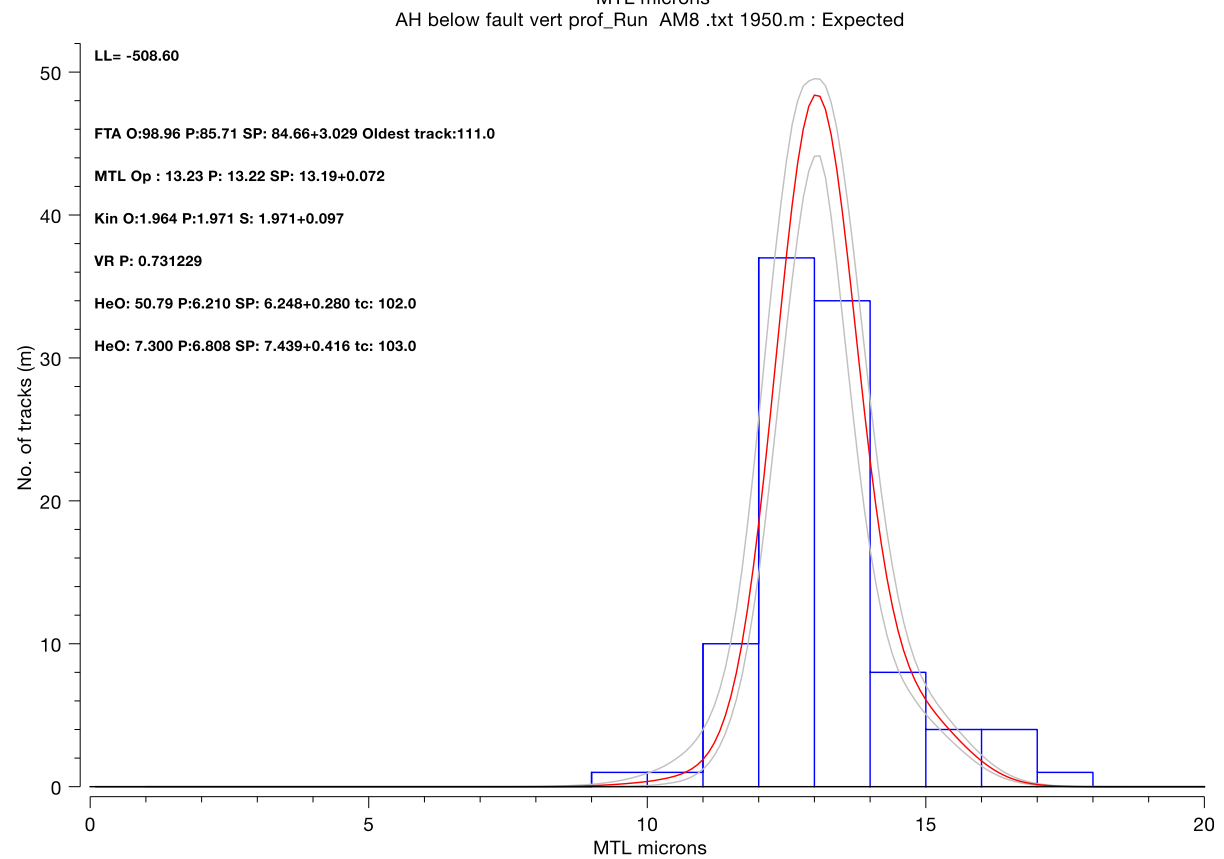
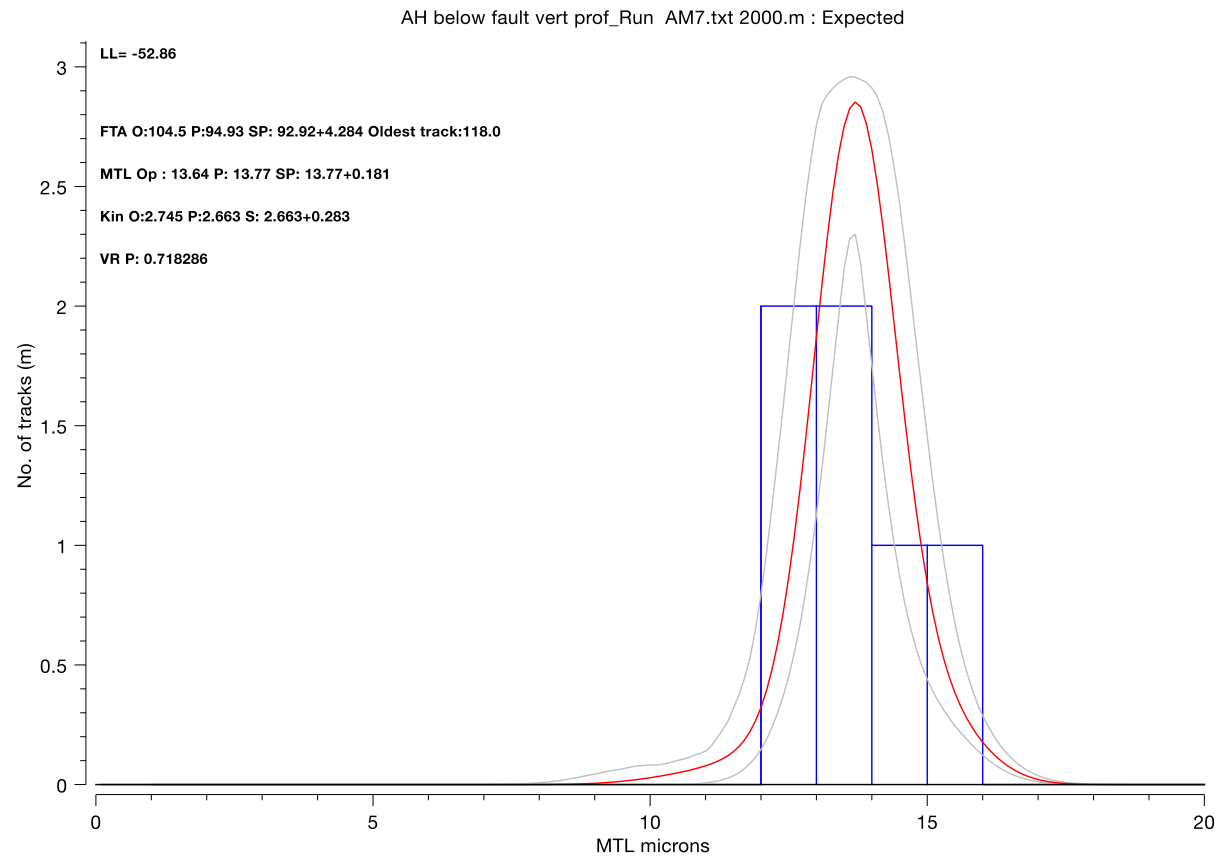


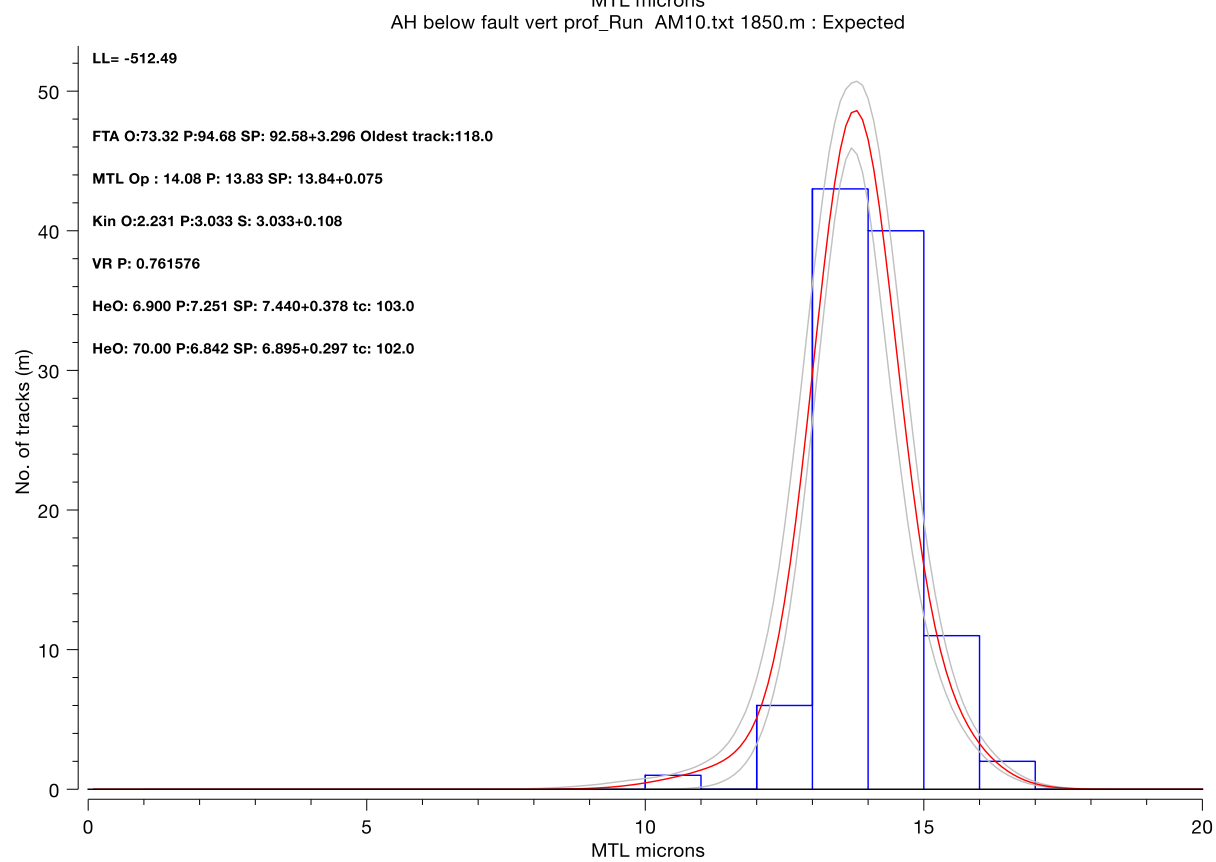
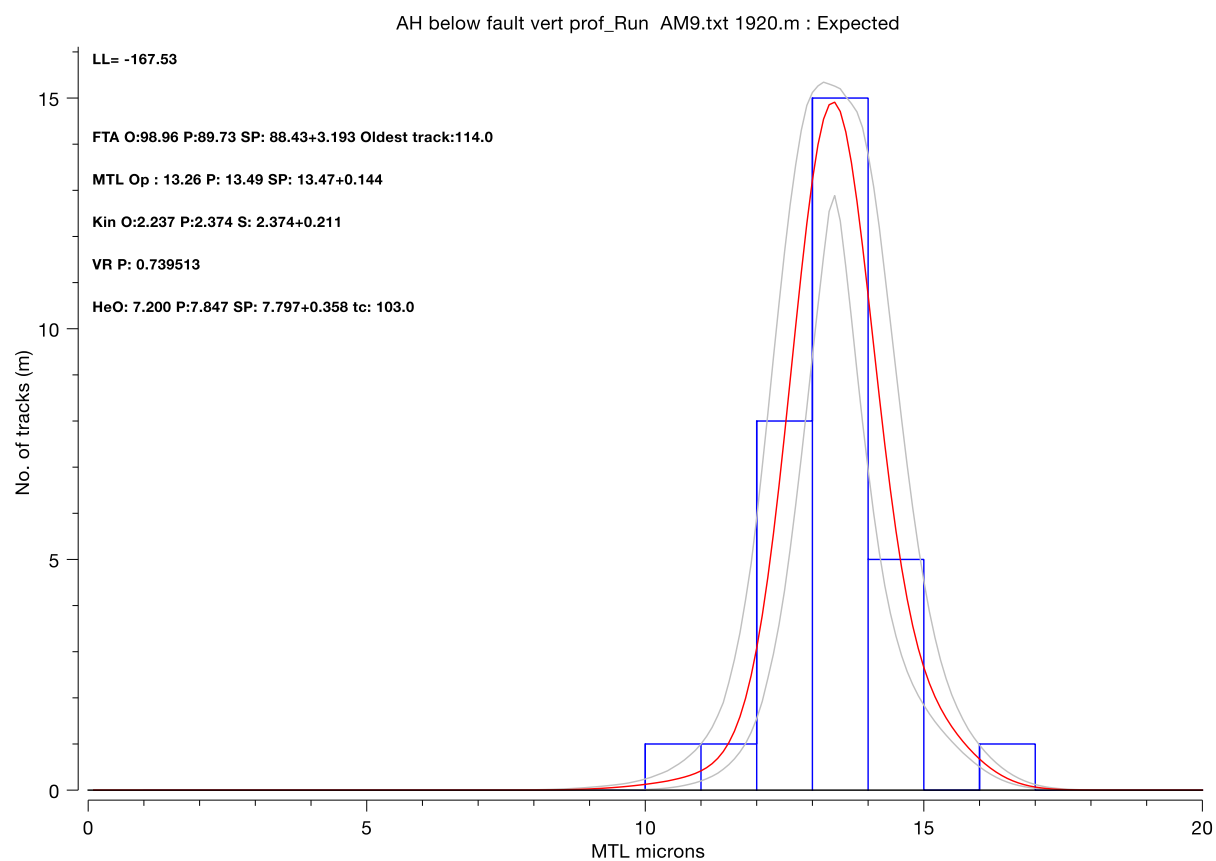


AH above fault w constraint_Run AM6.txt 2080.m : Expected

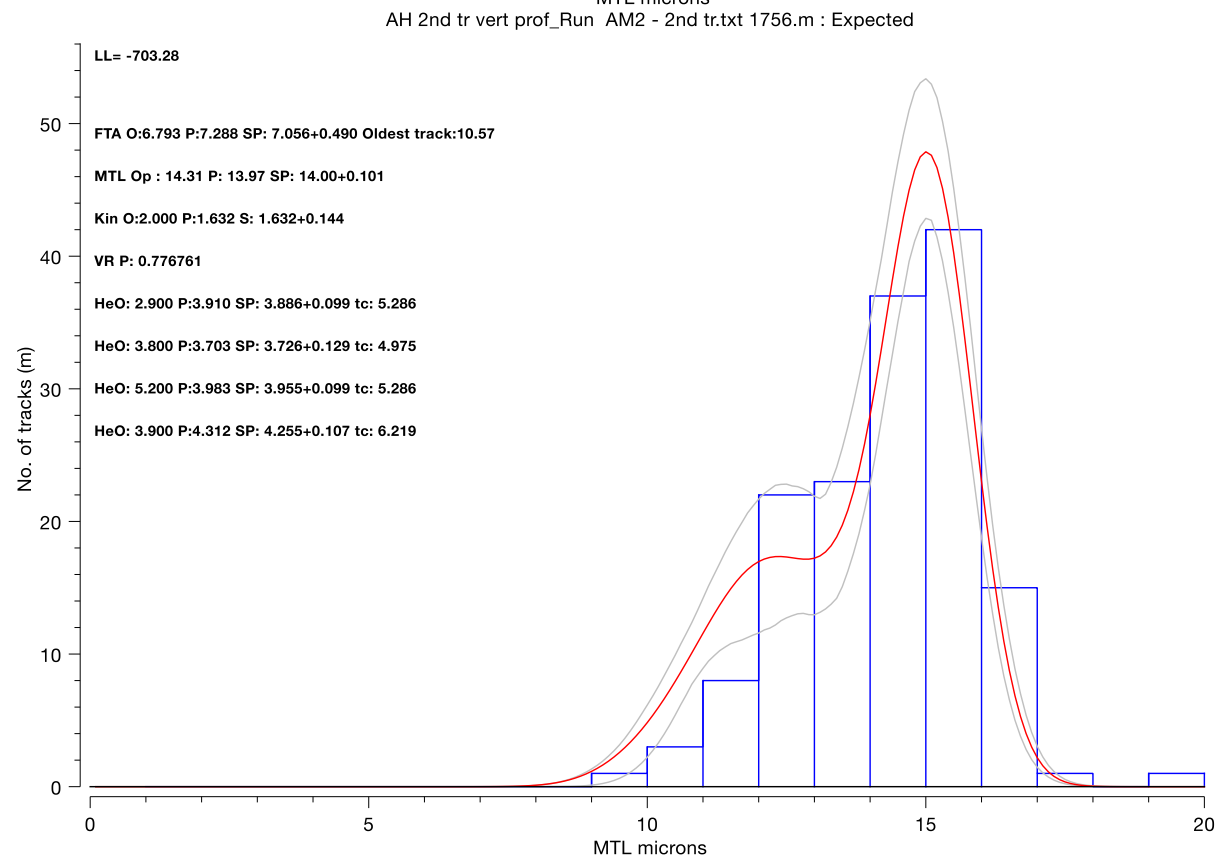
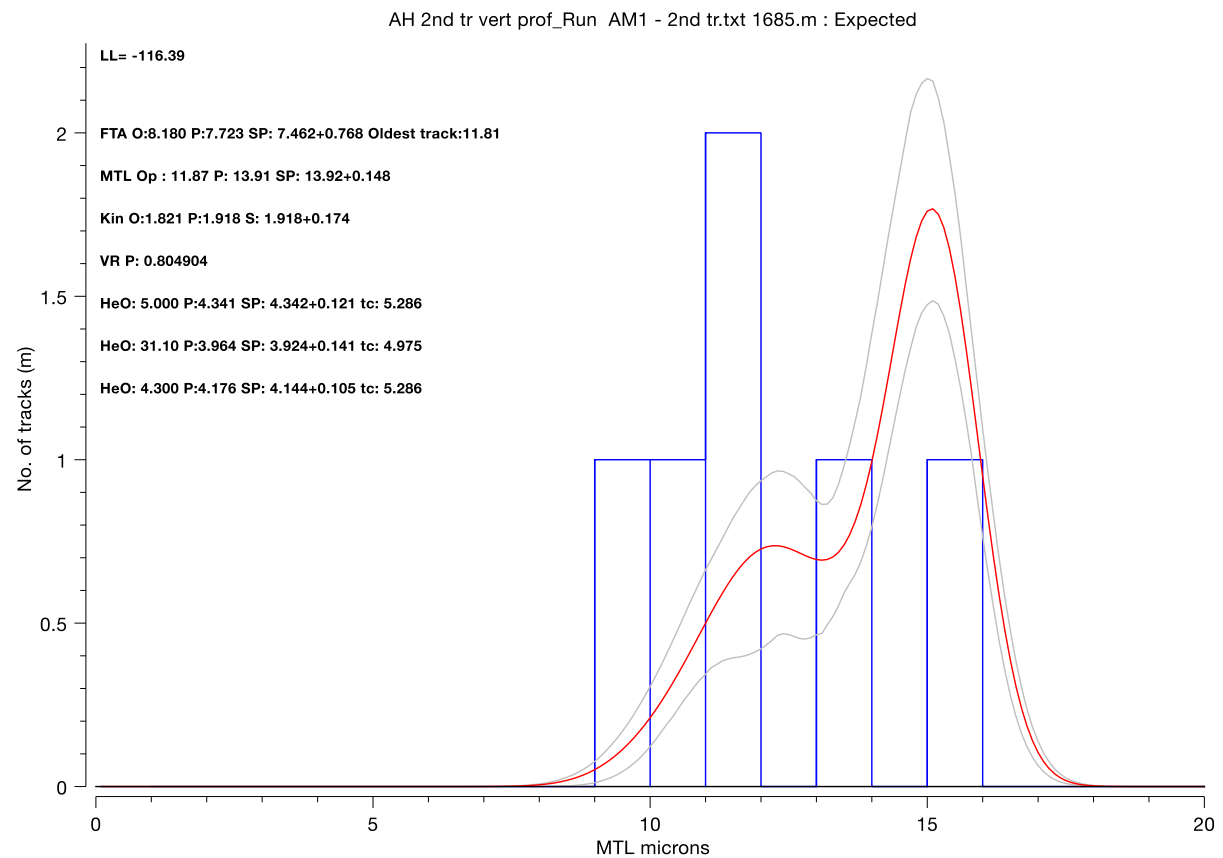


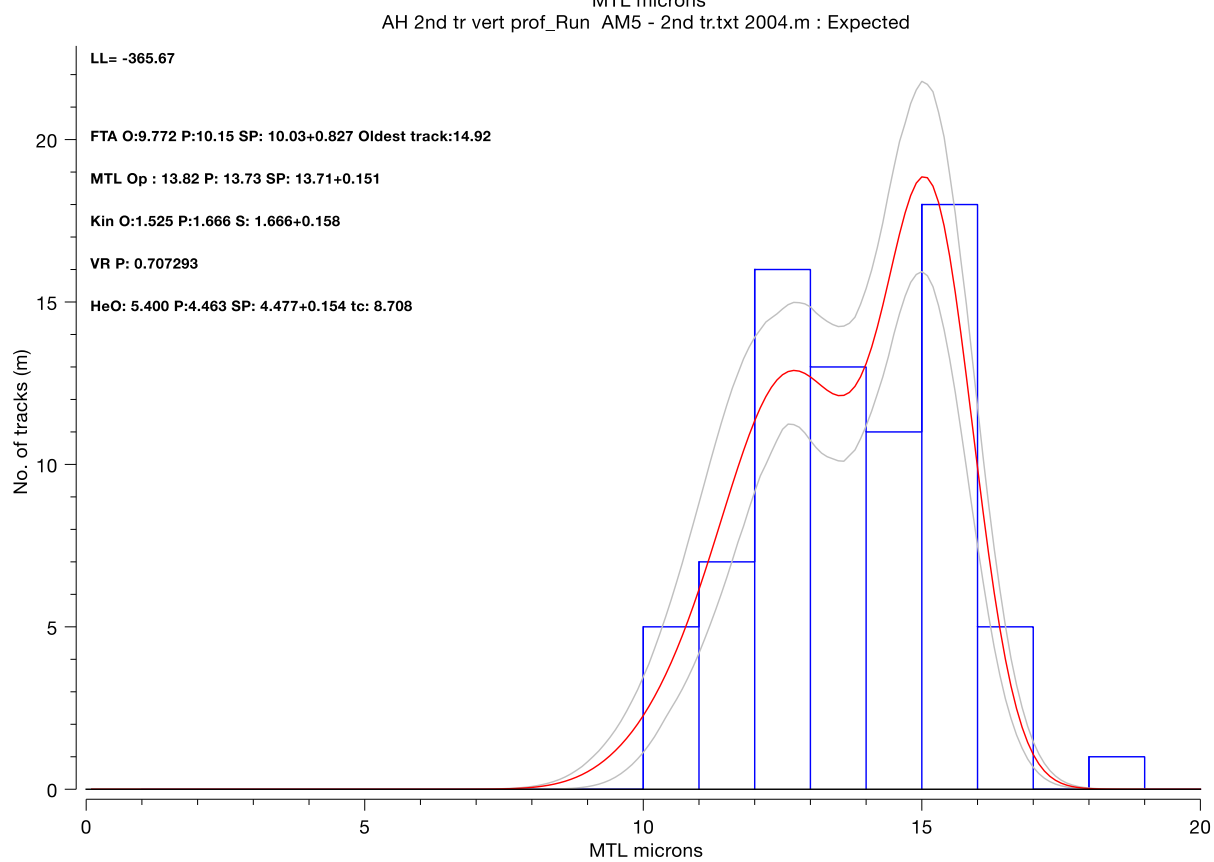
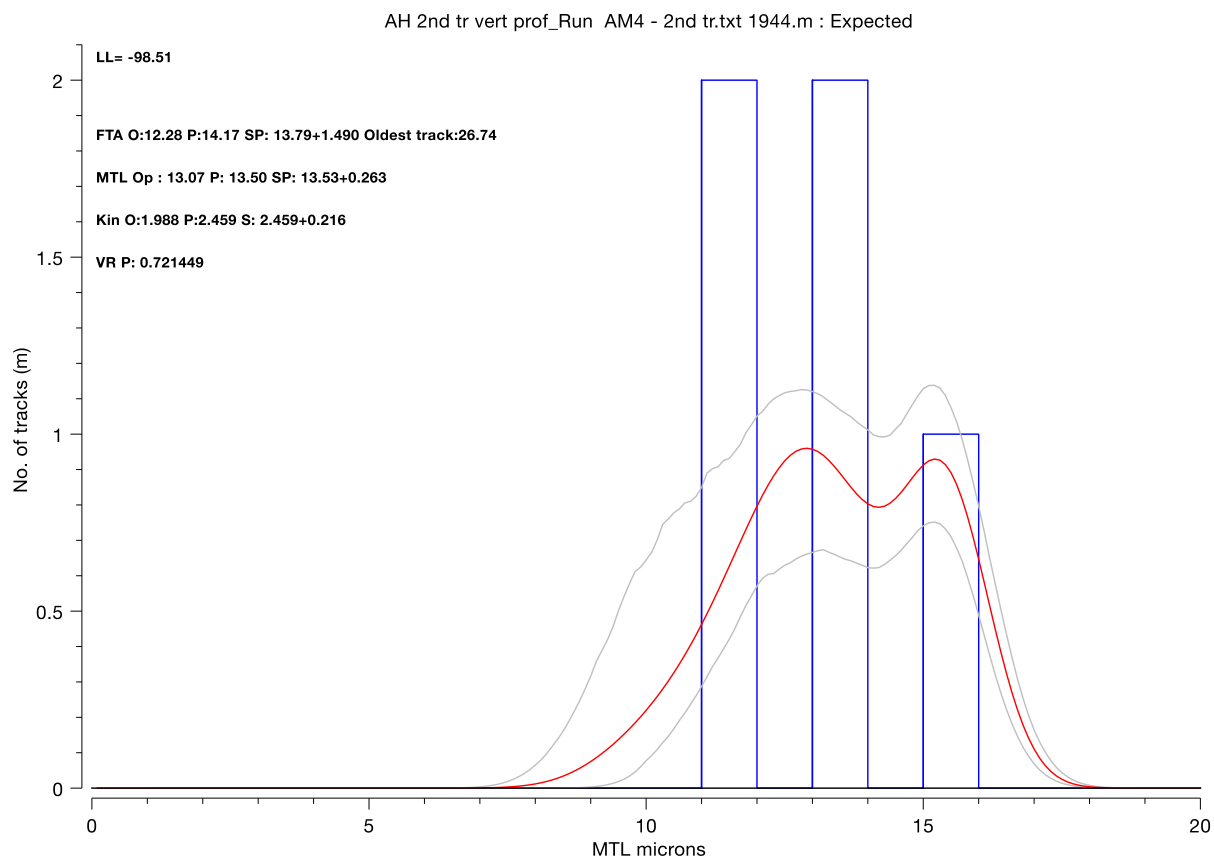
Amaro Horst 1st traverse, below secondary fault: Samples AM7-1st tr, AM8-1st tr, AM9-1st tr, and AM10-1st tr





Amaro Horst 2nd traverse: Samples AM1-2nd tr, AM2-2nd tr, AM4-2nd tr, and AM5-2nd tr





Influence of rift superposition on lithospheric response to East African Rift System extension: Lapur Range, Turkana, Kenya

Abstract

The Turkana Depression of northern Kenya, lies at the intersection of the NW-SE trending late Mesozoic-early Paleogene South Sudan and Anza rifts and the N-S trending late Paleogene-Recent East African Rift System (EARS). A low-temperature thermochronology study in the Lapur Range reveals a complex tectonothermal evolution related to multiple periods of regional and local tectonism. Zircon (U-Th)/He data from Precambrian basement record rapid Early Cretaceous denudational cooling. Coeval subsidence in the adjacent Anza and South Sudan rifts suggests that the northern Turkana region acted as a basement high separating the grabens, as well as an axial source of sediment. Between ~95-90 Ma, a period of reheating commenced with burial of Lapur basement beneath ~500 m of Turonian-Eocene Lapur Sandstone and ~1.5-3.5 km of latest Eocene-early Miocene Turkana Volcanics. Apatite fission track (AFT) and apatite (U-Th-Sm)/He data record a transition to rapid denudational cooling in the mid Miocene (~14 Ma) in response to EARS-related extension in the northern Turkana Basin. Thermal history models indicate the Lapur Range experienced ~90-100 °C of mid-Miocene to Plio-Pleistocene cooling, yielding the first Neogene AFT ages reported from Kenya related to EARS exhumation. We attribute the larger magnitude of cooling in the Lapur Range compared to other regions of the EARS to the attenuated crustal thickness and elevated heat flow of the Turkana Depression, crustal properties inherited from earlier Cretaceous-Paleogene rifting. The resulting low effective elastic thickness of the Turkana lithosphere allowed for increased isostatic footwall uplift in response to EARS extension.

5.1 Introduction

The East African Rift System (EARS) is the foremost modern example of a continental rift zone on Earth and has been pivotal in developing our understanding of the early stages of rifting and continental breakup. Extending ~3,500 km from the Red Sea coast of northern Ethiopia to Mozambique, the EARS is considered to be an intra-cratonic prelude to oceanic opening (Chorowicz, 2005; Morley, Ngenoh, & Ego, 1999) (Figure 5.1). Rifting in the EARS is localized within two extensive regions of crustal uplift, the northern Ethiopian Dome and the southern East African Dome, averaging 1,500 and 1,200 m elevation respectively. These two plateaux are separated by the Turkana Depression, a broad region of relatively subdued topography (average ~600 m) running NW-SE from the plains of South Sudan, through southern Ethiopia and northern Kenya, to the Kenyan coast. Here, at the junction of the Ethiopian and Kenyan Rifts, the Turkana Depression displays a complex history of rifting, rift-sedimentation and crustal thinning that is in stark contrast to the rest of the EARS.

Evolution of the Turkana Depression commenced at least as far back as the Late Jurassic during a regional extensional phase that resulted in the development of the large NW-SE trending South Sudan and Anza Rifts, the latter of which is traditionally thought to extend from the Kenyan coast to the eastern shores of Lake Turkana (Figure 5.1) (Ebinger et al., 2000; Morley, Bosworth, et al., 1999; Schull, 1988). A proposed subsurface link between these late Mesozoic-Paleogene rifts is believed to lie near the Kenyan-Ethiopian border, west of Lake Turkana (Bosworth, 1992; Ebinger, Yemane, Woldegabriel, Aronson, & Walter, 1993; Ibrahim, Ebinger, & Fairhead, 1991). During the Paleogene, a series of extensional basins developed in the Turkana region as part of a larger, poorly understood NNW-SSE rifting trend stretching from Sudan, along the South Sudanese-Ethiopian border and down into eastern Kenya (Bosworth, 1992; Hendrie, Kusznir, Morley, & Ebinger, 1994).

Oligocene?- Miocene to Recent EARS rifting was superimposed again, forming approximately N-S trending basins sub-parallel to those of the neighbouring Ethiopian and Kenyan Rifts, reactivating some pre-existing structures and abandoning others (Hendrie et al., 1994; Morley, Stone, Harper, & Wigger, 1999). Consequently, compared to the Ethiopian and Kenyan Rifts, the present-day Turkana segment exhibits anomalously thin crust (~20 km) and elevated heat flow (Benoit et al., 2006; Sippel et al. 2017).

Despite over a century of work investigating continental rifting in East Africa, significant uncertainty remains concerning the spatiotemporal evolution of extension in the long-lived Turkana portion of the EARS. This is largely due to poor surface exposure of older rift deposits in Turkana, which are often buried beneath thick volcanic sequences or by Lake Turkana itself (Morley, Stone, et al., 1999).

Low-temperature thermochronology, such as apatite fission track (AFT), apatite (U-Th-Sm)/He (AHe) and zircon (U-Th)/He (ZHe) analyses, together with thermal history modelling provide a powerful approach to investigate the tectono-thermal evolution of the uppermost ~10 km of the crust. These methods have often been utilized in rift settings to constrain footwall exhumation histories and the propagation of extension over time (e.g. Brown et al., 1990; Sullivan et al., 2000; Lisker et al., 2003; Seiler et al., 2011; Feinstein et al., 2013). Although several low-temperature thermochronology studies have previously investigated the timing and magnitude of faulting in other parts of the EARS (e.g. Foster and Gleadow, 1992, 1993, 1996; Wagner et al., 1992; Noble et al., 1997; Spiegel et al., 2007; Pik et al., 2008; Bauer et al., 2012, 2013; Torres Acosta et al., 2015), very few data are available from the Turkana Depression. A more precise understanding of the thermal evolution of the Turkana crust may reveal clues about how rifting has propagated over time in Turkana and how its lithospheric response to extension has been affected by multi-phase rift superposition.

Here, we present AFT, AHe and ZHe data for 10 samples from basement and overlying sedimentary rocks collected from the Lapur Range escarpment and along the ~E-W Lokitaung Gorge that incises the range. The Lapur Range is an exposed basement block in the footwall of the ~N-S Murua Rith-Lapur (MRL) master fault that bounds the northern Turkana Basin. Its location precisely at the intersection of the Cretaceous-early Paleogene Anza-South Sudan Rifts and the Neogene-Recent EARS presents a unique opportunity to investigate the effects of rift superposition on the region's upper crustal thermal evolution. Combined thermal history modelling reveals a comprehensive late Mesozoic-Recent tectono-thermal evolution of the Lapur block, providing important spatiotemporal constraints on the propagation of rifting in the Turkana Depression.

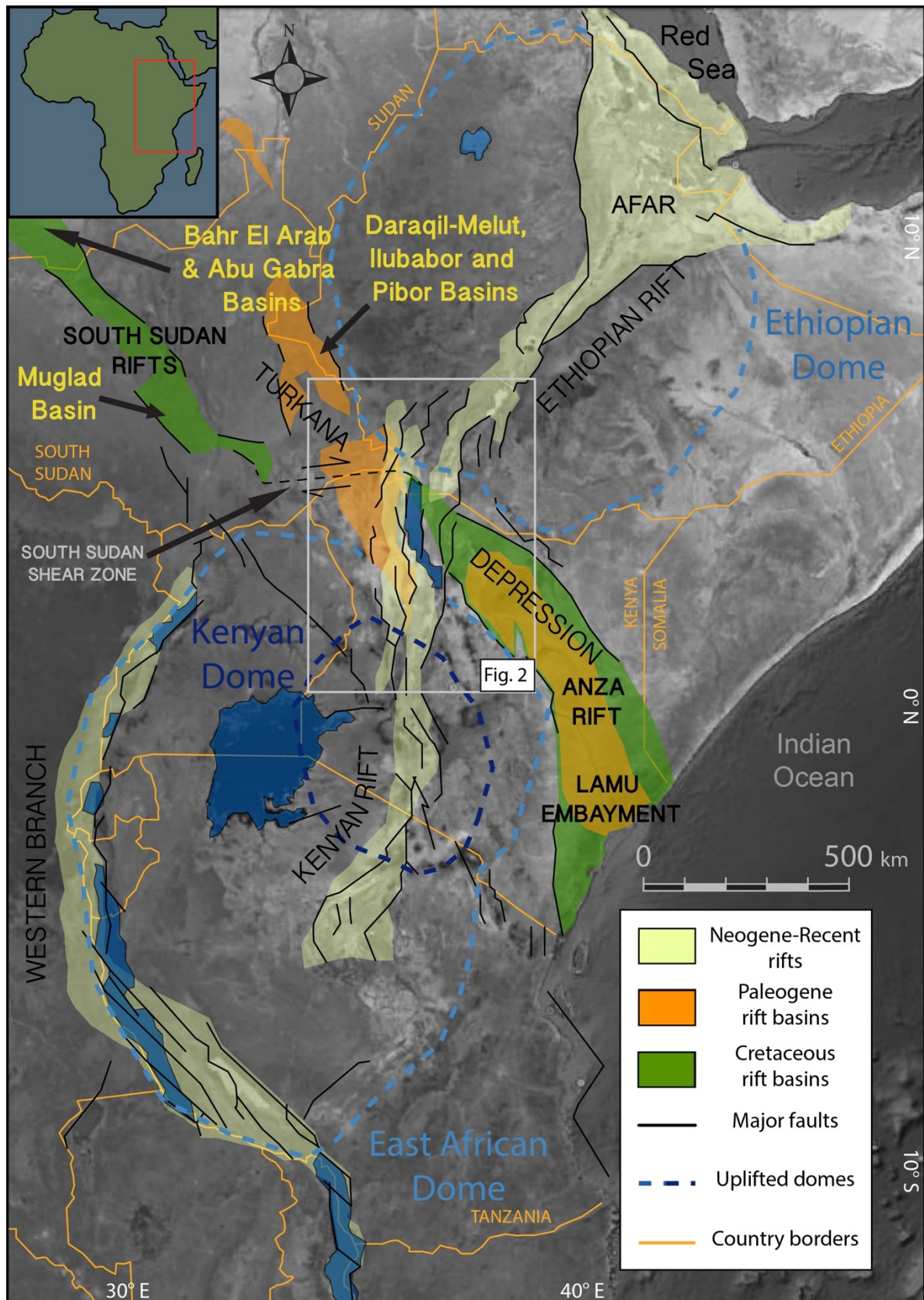


Figure 5.1. Regional map of East Africa showing the location of the Turkana Depression in relation to the distribution of Cretaceous, Paleogene and Neogene-Recent rift basins. The broad, uplifted Ethiopian and East African Domes are outlined in light blue. The Kenyan Dome, a subsidiary of the East African Dome, is outlined in dark blue. (after Hendrie et al., 1994).

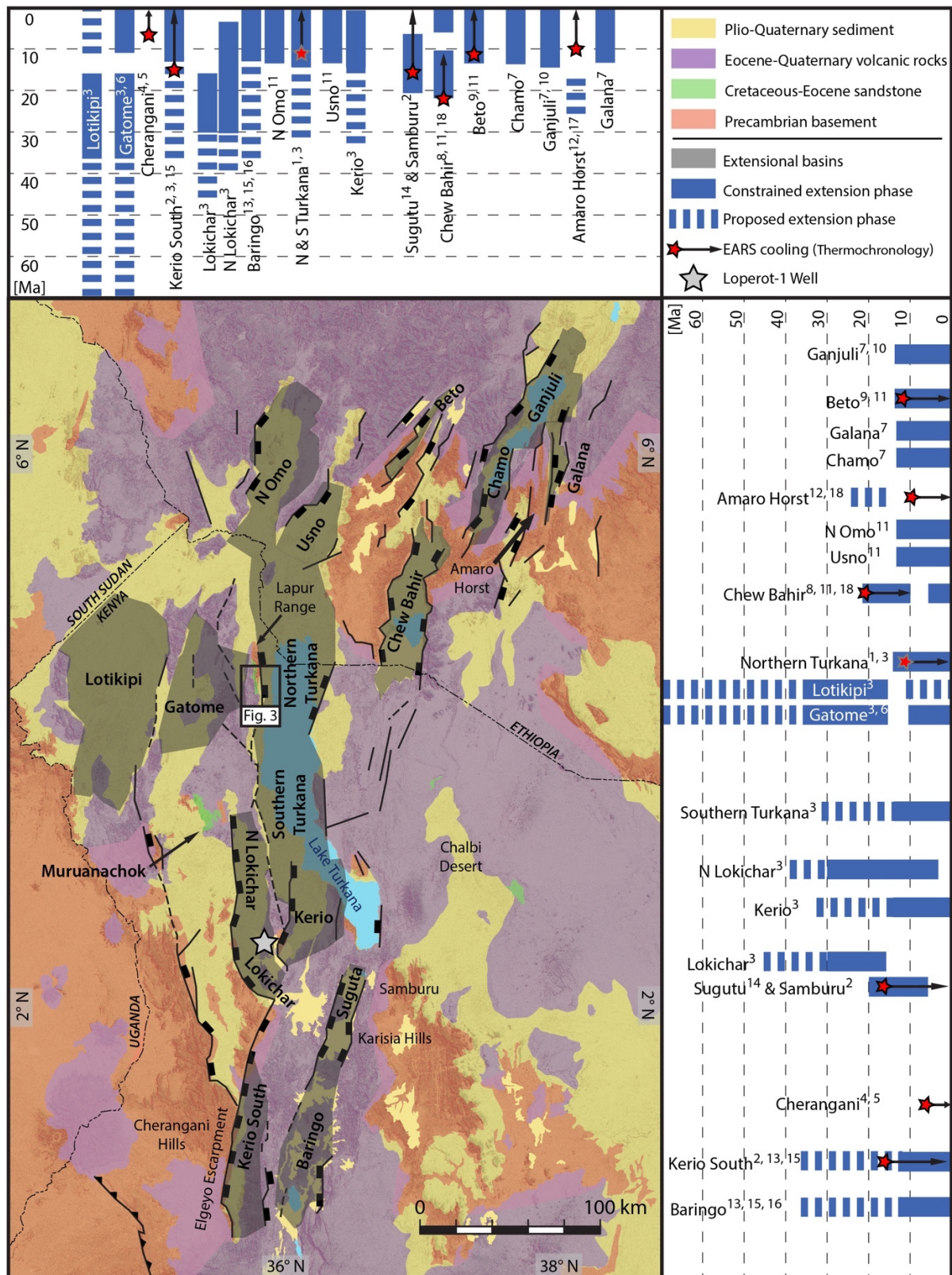


Figure 5.2 (Previous page). Simplified geological map of the Turkana Depression showing the distribution of latest Cretaceous?-Paleogene to Recent extensional basins (adapted from Kazmin, 1972; GMRD-Sudan, 1981; BEICIP, 1987; Lehto et al., 2014; Philippon et al., 2014; Balestrieri et al., 2016). Chronograms at the top and right sides of the geological map depict the spatial-temporal relationship of Cenozoic extension (after Torres Acosta et al., 2015). Solid blue bars show documented periods of extension from previous sedimentary and structural geology studies, while possible periods of extension that are poorly constrained by data are shown as hashed blue bars. Periods of rift-related rapid cooling from low-temperature thermochronology studies are shown with red stars with black outlines. A red star with a grey outline marks cooling observed in this study. Information was compiled from the following sources: (1) this study, (2) Torres Acosta et al. (2015), (3) Morley et al. (1999b), (4) Foster and Gleadow (1996), (5) Spiegel et al. (2007), (6) Wescott et al. (1999), (7) Ebinger et al. (1993), (8) Pik et al. (2008), (9) Philippon et al. (2014), (10) Woldegabriel et al. (1990), (11) Ebinger et al. (2000), (12) Zanettin et al. (1983), (13) Mugisha et al. (1997), (14) Morley et al. (1999c), (15) Lippard (1972), (16) Hautot et al. (2000), (17) Balestrieri et al. (2016) and (18) Bonini et al. (2005).

5.1.1 Tectonic and Geological Setting

Africa has been a centre of continental accretion since the late Proterozoic, with its modern terrane configuration having been established during the late Neoproterozoic-Early Paleozoic Pan-African Orogeny. The tectonic, magmatic and metamorphic activity of the Pan-African Orogeny culminated in the amalgamation of the supercontinent Gondwana (Kröner & Stern, 2005). The related ocean closure, arc and microcontinent accretion and the suturing of continental fragments resulted in a series of orogenic and mobile belts throughout Gondwana, including the Mozambique Belt that runs through modern-day Kenya and consists of Neoproterozoic passive margin sediments intercalated with Archean gneisses (Begg et al., 2009).

The early Paleozoic was marked by pervasive postorogenic magmatism, shearing, major tectonic uplift and deep seated exhumation of mobile belts throughout Gondwana (Unrug, 1997). Breakup of Gondwana began in earnest during late Paleozoic-Jurassic Karoo rifting as evidenced by the formation of widespread intracratonic rift basins over much of southern and eastern Africa (Kreuser, 1995).

5.1.1.1 Jurassic-Paleogene Anza and South Sudan Rifts

The NW-SE trending Anza Rift encompasses an area of extended continental crust stretching ~600 km from the Kenyan coast to the eastern shores of Lake Turkana (Figure 5.1) (Reeves, Karanja, & MacLeod, 1987). The eastern portion of the trough, in the Lamu Embayment area, is thought to have initially formed as an aulacogen during the break-up of Madagascar

from East Africa in the Early-Middle Jurassic (Greene, Richards, & Johnson, 1991) and was later reactivated during Cretaceous-Paleogene rifting. The sedimentary record of the Anza Rift is dominated by Upper Cretaceous-Paleogene lacustrine shales and fluvio-deltaic sandstones, showing an overall trend of younger rift activity towards the southeast (Morley, Bosworth, et al., 1999). Unlike later EARS rifting, Cretaceous-Paleogene extension was not accompanied by significant igneous activity (Morley, Ngenoh, et al., 1999).

The Cretaceous-Paleogene Abu Gabra, Muglad and Bahr el Arab rifts, collectively known as the South Sudan Rifts (Figure 5.1), formed in relation to the opening of the South and Equatorial Atlantic (Guiraud & Maurin, 1992). In the southeast the basins narrow towards the South Sudanese-Kenyan border, however due to a data gap in eastern South Sudan and superposition of later Paleogene-Recent EARS rifting in northern Kenya, the geometry of this termination remains uncertain (Guiraud & Bosworth, 1997). The compatible architecture, coeval timing and similar gravity signal suggest that the Anza Rift may be linked to the South Sudan Rifts in northern Turkana (Ibrahim et al., 1991; Schull, 1988). A half-graben basin of possible Cretaceous-Paleogene age detected by seismic imaging beneath the Lotikipi Plains of western Turkana (Figure 5.2) (Wescott, Wigger, Stone, & Morley, 1999) and the presence of Upper Cretaceous sediment elsewhere in Turkana suggests the Anza and South Sudan Rifts could be linked via a series of smaller rift basins that have since been masked by Neogene rifting in northern Kenya. Alternatively, the two rift systems may be connected via the proposed dextral South Sudan Shear Zone (Figure 5.1) (e.g. Browne and Fairhead, 1983; Bosworth, 1992).

5.1.1.2 Paleogene-Recent East African Rift System (EARS)

The EARS comprises two rifts trends, an eastern and western branch, manifested in two narrow (50-150 km) corridors of asymmetric half-grabens developed above listric faults with rollover anticlines in the hanging wall, which follow a trend of Precambrian mobile belts (Figure 5.1) (Morley, Ngenoh, et al., 1999). In general, volcanism in the EARS started earlier than extension, with the earliest manifestation being tholeiitic flood basalts emplaced between 45 and 33 Ma in the northern Turkana Depression (Bellieni et al., 1981; Ebinger et al., 1993; R. M. George & Rogers, 2002; Wescott et al., 1999). The volcanically-rich eastern branch consists of the Ethiopian and Kenyan Rifts, where EARS-related extension began in the early Miocene (Morley, Ngenoh, et al., 1999). However, there is some evidence for earlier rift activity in the Turkana Depression during the Paleogene (Bosworth & Morley,

1994; Ebinger et al., 1993; Hendrie et al., 1994; Morley, Stone, et al., 1999; Wescott et al., 1999).

5.1.1.3 Turkana Depression

The pattern of rifting in the Turkana Depression contrasts with the rest of East Africa in terms of both duration and deformation style (Figure 5.2). Anza rifting likely commenced in the Early Cretaceous on the east side of Lake Turkana, but by Paleogene time extensional tectonics had migrated to the west where it linked up with the South Sudan rifts (Bosworth, 1992; Hendrie et al., 1994; Wescott et al., 1999). EARS-related rifting began in western the Turkana Depression by the late Oligocene and migrated eastward during the Miocene, initiating the rift basins underlying Lake Turkana and their onshore equivalents by the late Miocene (Baker & Wohlenberg, 1971; Morley, Stone, et al., 1999). During the latest Pliocene-Pleistocene (< 3 Ma), extension in the Turkana Depression localised into a narrower zone of igneous intrusions and minor faults east of Lake Turkana, in alignment with coevally developed dike and fault swarms observed in the Ethiopian and Kenyan rifts (Morley, Ngenoh, et al., 1999). As a result, the modern Turkana Depression consists of a series of N-S striking half-grabens that are bound by steep ($\sim 65^\circ$) east-dipping listric normal faults, which are thought to sole into a shallow crustal ductile detachment at ~ 10 - 12 km depth (Morley, Stone, et al., 1999). These master faults have several kilometres of vertical throw, providing the accommodation space for up to ~ 7 km of late Cenozoic sediments (Chorowicz, 2005). The Lokichar Basin (Figure 5.2) alone experienced over 10 km of extension (Morley et al., 1992).

The prolonged duration of rifting, superposition of multiple rift phases and longitudinal propagation of extension resulted in the Turkana Depression being the widest part of the EARS (~ 150 km, consisting of 3-4 major half-grabens) (Morley, Stone, et al., 1999), as well as the area of thinnest continental crust south of the Afar (18-20 km under western Turkana, compared to 35-40 km under the Kenyan Rift) (Henry et al., 1990; Maguire, Swain, Masotti, & Khan, 1994; Mechie et al., 1994).

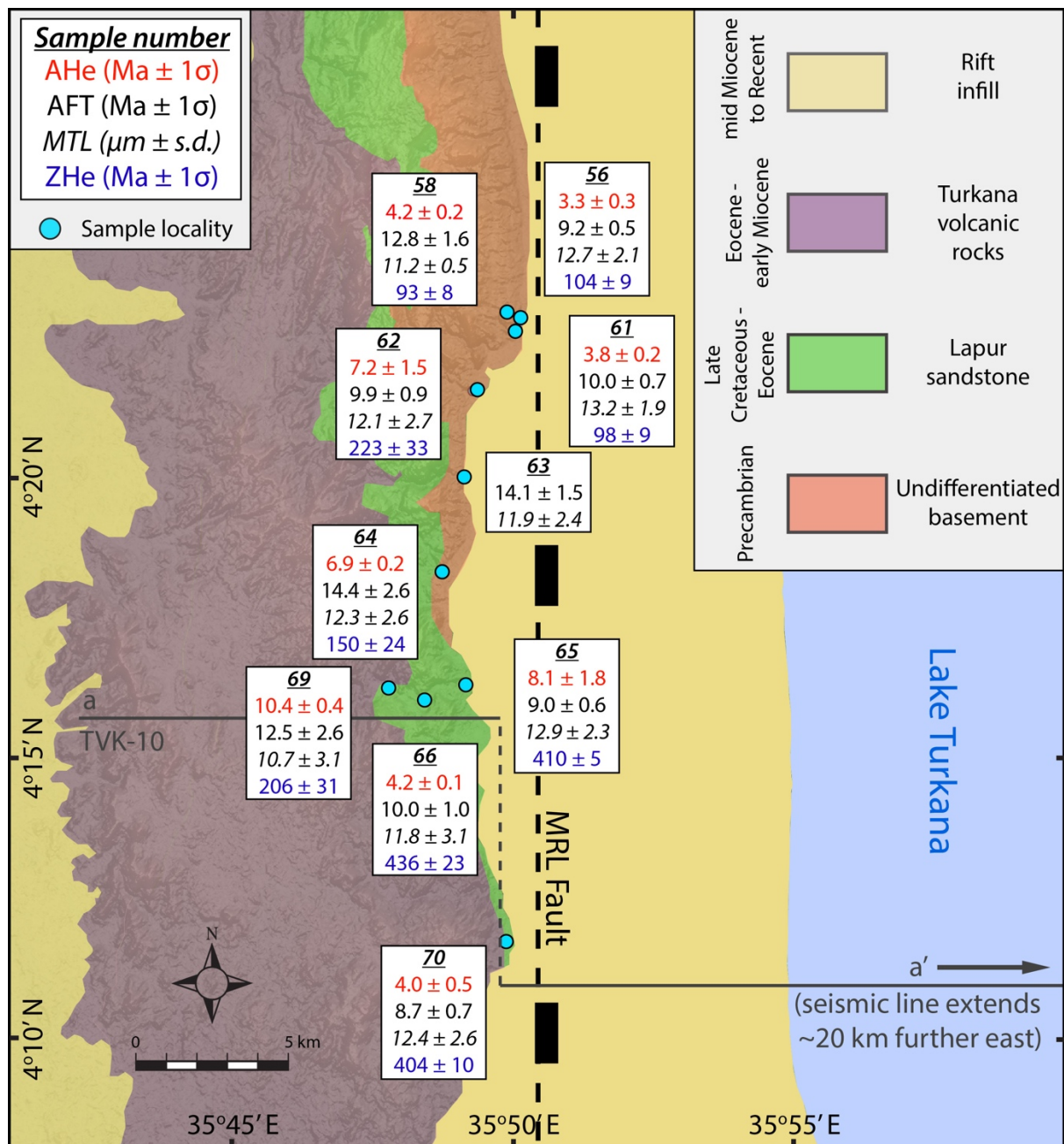


Figure 5.3. Simplified geological map of the Lapur Range with sample localities (blue dots) and low-temperature thermochronology results of this study. Refer to Figure 5.2 for geographic location. ZHe (blue), AFT and mean track length (MTL; see Section 5.2) (black), and AHe data (red) for each sample are presented in white boxes. The approximate location of the MRL normal fault is illustrated with a black hashed line. The location of seismic line TVK-10 (Figure 5.4a) is illustrated in black (Wescott et al., 1999). Adapted from Walsh and Dodson (1969); Abdelfettah et al. (2016).

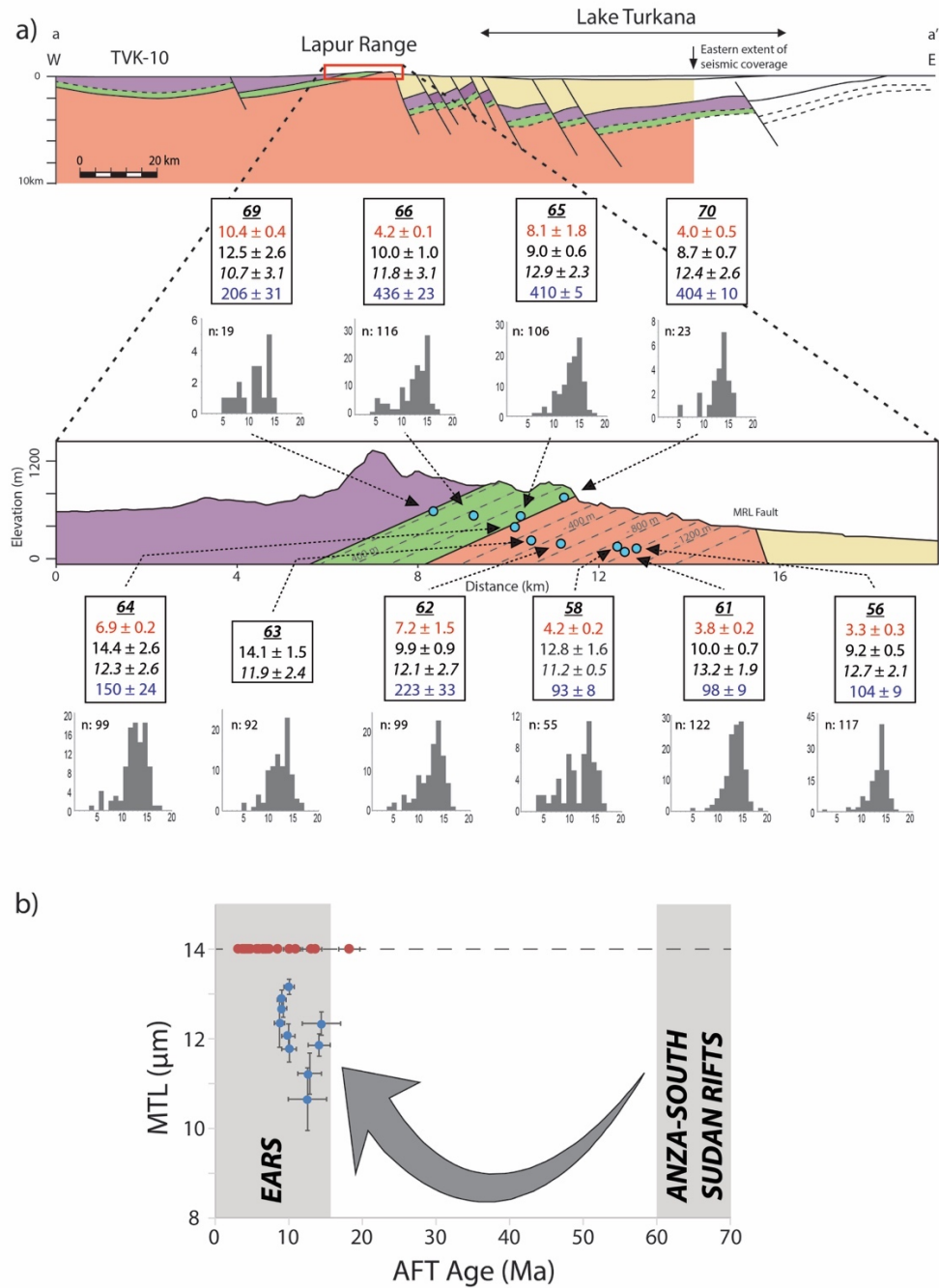


Figure 5.4. a) Sample locations (blue dots) projected on a geological cross-section of the Lapur Range and northern Turkana Basin adapted from Wescott et al. (1999). ZHe, AFT, AHe and confined fission track length distributions are shown for each sample. n = number of confined track length measurements included in the histogram and MTL result for that sample. Dashed lines in upper cross-section represent inferred unit boundaries from seismic interpretation of Wescott et al. (1999). Grey hashed lines on lower cross-section represent the relative stratigraphic depth of samples in relation to the basement-Lapur Sandstone nonconformity, prior to Miocene faulting and block rotation. Note that the dips of the MRL Fault and unit boundaries appear steeper due to the slight vertical exaggeration applied (true dips are $\sim 65^\circ$ E for the MRL Fault and $\sim 15^\circ$ W for the unit boundaries). Geological units and sample data follow the key in Figure 3. **b)** Lapur AFT data (blue circles) plotted as AFT age versus MTL, forming the ascending portion a boomerang plot (see Section 3.4). Red circles represent AHe ages.

5.1.1.4 Lapur Range

The Lapur Range, attaining elevations of up to 1560 m a. s. l., is located along the northwest shores of Lake Turkana at the intersection of the failed Anza-South Sudan Rifts (Figure 5.3). The Lapur block sits in the footwall of the MRL Fault, a N-S trending, east-dipping normal fault that bounds the western side of the northern Turkana Basin (Figure 5.4a) (Abdelfettah et al., 2016; Morley et al., 1992; Tiercelin et al., 2012). The northern Turkana Basin is a generally N-S trending, westward-dipping half graben that hosts present-day Lake Turkana. Based on seismic interpretations, it is thought to have begun to form in the middle Miocene (15-10 Ma) (Dunkelman, Rosendahl, & Karson, 1989; Morley et al., 1992; Morley, Stone, et al., 1999).

The basement of the Lapur Range is composed of Precambrian Mozambique Belt gneisses, amphibolites, granulites and other metamorphic assemblages (Begg et al., 2009). The basement is overlain by the Upper Cretaceous-lower Paleogene Lapur Sandstone (Arambourg & Wolff, 1969), which is up to 500 m thick and thinning to the north and south. The unconformity and bedding in the Lapur Sandstone dip $\sim 15^\circ$ to the WSW to SW, as dip direction varies slightly throughout the range (Morley et al., 1992; Tiercelin et al., 2012). Rocks exposed in the NE Lapur block therefore represent structurally deeper crustal levels than those in the south and west of the range. The Lapur Sandstone is interpreted as a fluvial deposit that was axially fed into the Anza Rift (Morley et al., 1992), with the lowest strata estimated as Turonian-early Campanian based on the presence of dinosaur and reptilian fossils (O'Connor, Sertich, & Manthi, 2011). The uppermost beds are locally interbedded with basal flows of the overlying Turkana Volcanics (Tiercelin et al., 2012).

The Turkana Volcanics were emplaced between ~ 37 -17 Ma and consist of upper Eocene-Oligocene basalts and interbedded tuffs, overlain by lower-middle Miocene rhyolites (Morley et al., 1992; Walsh & Dodson, 1969; Zanettin, Justin Visentin, Bellieni, Piccirillo, & Francesca, 1983). The total thickness of volcanics varies between basins to the west and east of the Lapur Range, ranging from ~ 3650 m in the Gatome Basin to ~ 1250 m under northern Lake Turkana, respectively (Morley et al., 1992; Wescott et al., 1999).

The subsurface stratigraphy in the hanging wall of the MRL Fault is primarily known from an $\sim$ E-W trending onshore seismic line (TVK-10; Figures 5.3 & 5.4a). Wescott et al. (1999) interpreted the seismic data as showing the Middle Miocene Turkana Volcanics conformably

overlain by a thick pile (up to 2 s TWTT) of late Miocene-Pliocene sedimentary rocks. Due to the highly degraded quality of seismic data below the Turkana Volcanics, the deeper basin structure is almost entirely unknown (Wescott et al., 1999). However, a recent reinterpretation of the same seismic line identified a ca. 300 m thick sedimentary package in the MRL hanging wall underlying the Turkana Volcanics, which was interpreted as a continuation of the Lapur Sandstone (Tiercelin et al., 2012). Only the uppermost portion of the basin fill crops out and comprises the Plio-Pleistocene Nachukui Formation, fluvio-lacustrine sediments dated between 4.2 and 0.7 Ma (Feibel, Brown, & McDougall, 1989). Offshore seismic data from Lake Turkana indicate a 4-km thick sedimentary package overlying the Turkana Volcanics (Dunkelman et al., 1989).

5.1.1.5 Previous low-temperature thermochronology studies

Previous thermochronological studies in East Africa are primarily based on AFT and show that the regional basement underwent early Mesozoic monotonic cooling related to craton exhumation. The oldest AFT ages (>250 Ma) are reported from the Tanzanian Craton in SW Kenya (Charles H. Kasanzu, 2016; Charles Happe Kasanzu et al., 2016; M Wagner et al., 1992). In some places however, this pattern has been disrupted by Cretaceous and Paleogene episodes of rapid cooling associated with rift-related denudation. Cretaceous to Paleogene cooling events are recorded in several locations throughout Kenya, including the Karisia Hills (>120 and 70-50 Ma; Foster and Gleadow, 1992), the Cherangani Hills (~70-50 Ma; Foster and Gleadow 1993, 1996) as well as the Samburu Hills and Elgeyo Escarpment (~50-35 Ma; Torres Acosta et al., 2015) (Figure 5.2). Similar ages are found in neighbouring Tanzania (W P Noble et al., 1997) and Ethiopia (Balestrieri, Bonini, Corti, Sani, & Philippon, 2016; Philippon et al., 2014). This pervasive Cretaceous-Paleogene phase of rapid cooling has been interpreted as resulting from denudation of the Anza rift shoulders during the graben's formation (Foster & Gleadow, 1992).

To date, Neogene cooling in Kenya related to EARS faulting and footwall erosion has only been inferred from thermal history modelling of low-temperature thermochronology data from the base of the Kenyan Rift flanks (Spiegel et al., 2007; Torres Acosta et al., 2015). This suggests that the magnitude of exhumation there was not sufficient to expose rocks which have cooled through the AFT and AHe temperature sensitivity ranges (~60-120 °C and ~30-80 °C, respectively). By contrast, Miocene ages from southern Ethiopia clearly record the onset of denudational cooling beginning at ~20-18 Ma in the Chew Bahir Basin (Pik et

al., 2008), followed by more widespread rapid cooling beginning at ~14-8 Ma in the Beto Basin and the Amaro Horst (Balestrieri et al., 2016; Philippon et al., 2014). Rapid cooling in central Ethiopia appears to have initiated slightly later at ~7 Ma (Abebe, Balestrieri, & Bigazzi, 2010).

5.2 Samples and Methods

Ten samples were collected to constrain the low-temperature thermal history of the Lapur Range in northern Turkana (Figures 5.3 and 5.4a). Six are from Neoproterozoic-early Paleozoic gneisses and quartzites of the Mozambique Belt, collected at the base of the exposed footwall, ca. ~25-1000m below the Lapur Sandstone nonconformity. An additional four samples were collected from the Upper Cretaceous-Paleogene conglomerates and arkosic sandstones of the Lapur Sandstone; three of these are from an ~E-W striking dry river canyon, the Lokitaung Gorge, forming a footwall transect near-perpendicular to the fault strike to sample progressively shallower-crustal rocks. Sample details, including rock types, coordinates, and elevations are listed in Table 5.2.

AFT thermochronology is based on the formation of radiation damage trails in apatite, known as fission tracks, that form as a result of spontaneous fission of ^{238}U (e.g. Fleischer et al., 1975; Wagner and Van den Haute, 1992). Fission tracks are sensitive to temperature, remaining relatively stable in apatite below ~60 °C but annealing nearly instantaneously above ~110-120 °C (A. J. W. Gleadow & Duddy, 1981). Between these temperatures, in what is called the partial annealing zone (PAZ), apatite fission tracks anneal (shorten) progressively as a function of time and temperature (Laslett, Green, Duddy, & Gleadow, 1987). The annealing behaviour of fission tracks in apatite is compositionally controlled, with chlorine-rich apatites being more resistant to annealing than their fluorine-rich counterparts (Barbarand, Carter, Wood, & Hurford, 2003; Donelick, O'Sullivan, & Ketcham, 2005). By combining age and fission track length information, the thermal history of a sample within the PAZ can be reconstructed (Kerry Gallagher, 1995; Ketcham, 2005).

AHe and ZHe thermochronology is based on the production of ^4He during the radioactive decay of ^{238}U , ^{235}U and ^{232}Th . ^4He is also produced from the radioactive decay of ^{147}Sm , however this generally constitutes <1% of the total ^4He , compared to typical He age

precisions of a few percent (Kenneth A Farley & Stockli, 2002). When analysing U and Th-rich minerals, such as zircon, the ^4He contribution from ^{147}Sm is considered statistically insignificant. However, in some U and Th poor phosphate minerals (e.g. apatite), the consideration of the ^{147}Sm α -source is required. ^4He diffusivity is a function of time, temperature, crystal size and accumulated radiation damage (K A Farley, 2000; Reiners & Farley, 2001; Shuster, Flowers, & Farley, 2006). In apatite, ^4He diffusion accelerates significantly above $\sim 40^\circ\text{C}$ and nearly instantaneously above $\sim 80^\circ\text{C}$, between which ^4He is partially retained (Wolf, Farley, & Kass, 1998). This temperature range is referred to as the apatite partial retention zone (AHe PRZ). The ZHe system is sensitive to higher temperatures, with the zircon partial retention zone (ZHe PRZ) ranging between $\sim 130\text{--}200^\circ\text{C}$ (Reiners et al., 2002; Stockli 2005; Biswas et al. 2007, Wolfe & Stockli 2010).

Multiple studies have shown that He diffusion in both apatite and zircon is dependent on a range of crystallographic and compositional characteristics, sometimes causing He age dispersion. These include U-Th rich inclusions (Danišik et al., 2017; Lippolt, Leitz, Wernicke, & Hagedorn, 1994), which are excluded during the crystal selection procedure, He implantation from external sources for very low U-Th (< 3 ppm) grains (Spiegel, Kohn, Belton, Berner, & Gleadow, 2009), U and Th zonation (Ault & Flowers, 2012; Danišik et al., 2017; Orme, Reiners, Hourigan, & Carrapa, 2015; Tagami, Farley, & Stockli, 2003; Tripathy-Lang, Fox, & Shuster, 2015), and Cl content in apatite (Gautheron, Tassan-Got, Barbarand, & Pagel, 2009). The destructive nature of the AHe and ZHe analytical procedures employed in this study prevent the quantification of the effects of the latter two variables. Three additional variables that may produce intra sample He age variation, i.e. grain size (K A Farley, 2000), accumulated radiation damage (Danišik et al., 2017; Flowers, Ketcham, Shuster, & Farley, 2009; Gautheron et al., 2009; Guenther, Reiners, Ketcham, Nasdala, & Giester, 2013; Guenther, Reiners, & Tian, 2014; Nasdala et al., 2004; Shuster & Farley, 2009; Shuster et al., 2006), and crystal breakage (Brown et al., 2013), have been more comprehensively investigated, each of which can lead to over 50% intra-sample age dispersion. Here, we consider grain size, morphology and radiation damage in our thermal history modelling and data interpretation. We use the effective uranium concentration ($e\text{U}$; $\text{U} + 0.235 \times \text{Th}$), which considers the decay rate and abundance of radiogenic parent elements, as a proxy for α -radiation damage.

A comprehensive review of the AFT, ZHe and AHe methodologies and technical procedures used in this study is given by Gleadow et al. (2015). Apatites and zircons were separated using the standard heavy mineral separation techniques. A synopsis of the AFT, ZHe and AHe analytical protocols are given in the footnotes of Tables 1, 2, and 3, respectively

5.3 Data

ZHe, AFT and AHe analytical data are listed in Tables 5.1-3, respectively. A simplified geological map and a cross-section of the Lapur Range, showing sample locations, ages and mean track lengths (MTL) data are presented in Figures 5.3 and 5.4a, respectively. Pooled and central AFT ages are indistinguishable within analytical uncertainties and all age values displayed in figures and discussed below are central ages. MTL measurements discussed in the text are reported as un-projected mean 3D lengths, although lengths were *c*-axis projected using the approach of Ketcham et al. (2007b) for thermal history modelling (Section 5.3.5). All samples yielded sufficient grains for experimentation except for sample 63, where no appropriate apatite or zircon geometries were found for He analyses. None of the AHe or ZHe ages exhibit correlations with either grain size or morphology. All ages are reported to one decimal place in the tables but rounded to whole numbers in the text when 1σ errors are > 5 Ma.

Table 5.1. Lapur Range Zircon (U-Th)/He Data Summary

Sample No.	Lab #	⁴ He [ncc]	Mass [mg]	U [ppm]	Th [ppm]	Th/U ratio	[eU] (ppm) ^a	F _r ^b	Corrected Age [Ma ± 1σ]	Grain length [μm]	Grain half-width [μm]	^c Grain morphology
56	9623	34.194	0.0136	202.5	84.5	0.42	222.4	0.81	93.1 ± 7.8	352.1	51.3	2T
56	9624	22.078	0.0107	135.6	42.7	0.31	145.6	0.80	115.6 ± 9.7	315.8	48.6	2T
56	9625	28.071	0.0131	193.2	71.3	0.37	209.9	0.82	84.1 ± 7.0	321.1	53.8	2T
56	9626	46.853	0.0130	224.0	86.4	0.39	244.3	0.80	121.0 ± 10.1	371.6	48.1	2T
58	9628	21.117	0.0100	221.3	36.8	0.17	230.0	0.82	75.5 ± 6.4	238.1	59.6	2T
58	9630	28.179	0.0182	145.7	31.2	0.21	153.0	0.85	83.2 ± 7.1	300.4	70.3	2T
58	9629	17.338	0.0078	164.1	81.1	0.49	183.2	0.79	99.8 ± 8.2	245.6	49.0	2T
58	9627	18.203	0.0145	83.8	29.1	0.35	90.6	0.82	113.4 ± 9.5	323.2	114.0	2T
61	9631	35.061	0.0154	169.9	77.8	0.46	188.2	0.83	99.3 ± 8.2	316.6	60.1	2T
61	9632	52.839	0.0158	200.6	88.9	0.44	221.5	0.82	123.4 ± 10.2	358.4	55.3	2T
61	9633	15.234	0.0081	174.6	71.7	0.41	191.4	0.79	81.1 ± 6.8	253.0	48.8	2T
61	9634	23.948	0.0117	172.6	98.4	0.57	195.8	0.81	86.2 ± 7.1	296.6	53.6	2T
62	9636	51.148	0.0052	305.1	122.0	0.40	333.7	0.75	240.8 ± 19.8	245.6	38.2	2T
62	9637	136.517	0.0131	259.1	115.1	0.44	286.2	0.82	294.8 ± 24.0	289.6	58.6	2T
62	9638	101.394	0.0105	304.4	107.7	0.35	329.7	0.82	237.2 ± 19.6	244.3	60.0	2T
62	9635	102.899	0.0104	261.1	102.6	0.39	285.2	0.79	279.7 ± 22.9	331.1	92.1	2T
62	11993	211.990	0.0195	299.5	115.0	0.38	326.5	0.84	269.1 ± 22.1	330.6	67.1	2T
62	11994	142.896	0.0097	417.7	166.5	0.40	456.8	0.80	259.6 ± 21.3	295.4	48.2	2T
62	11995	171.091	0.0114	553.0	169.8	0.31	592.9	0.81	205.5 ± 17.1	302.3	52.0	2T
62	11996	111.797	0.0097	339.3	137.2	0.40	371.6	0.80	251.6 ± 20.7	282.7	49.8	2T
62	11997	631.565	0.0548	1607.5	664.3	0.41	1763.7	0.89	53.8 ± 4.5	426.2	103.4	2T
62	11998	480.950	0.0317	554.2	164.1	0.30	592.8	0.87	208.1 ± 17.3	400.1	76.9	2T
64	9639	34.369	0.0042	680.2	247.2	0.36	738.3	0.73	91.7 ± 7.7	236.7	69.4	2T
64	9640	124.164	0.0073	1006.0	268.7	0.27	1069.2	0.80	129.8 ± 10.9	191.8	124.9	2T
64	9641	130.430	0.0106	585.9	152.7	0.26	621.8	0.81	161.4 ± 13.5	280.5	105.9	2T
64	9642	140.765	0.0099	519.1	133.8	0.26	550.5	0.80	210.6 ± 17.6	305.0	94.8	2T
65	9660	29.686	0.0092	55.1	33.4	0.61	63.0	0.80	407.4 ± 32.5	252.3	106.7	2T
65	9661	37.576	0.0118	54.1	30.3	0.56	61.3	0.81	413.6 ± 33.1	296.6	108.1	2T
65	9662	65.229	0.0086	133.0	76.8	0.58	151.1	0.81	401.2 ± 32.1	212.3	122.6	2T
65	9663	44.482	0.0078	102.5	28.1	0.27	109.1	0.79	418.9 ± 34.3	255.7	94.5	2T
66	9664	34.752	0.0079	62.8	31.3	0.50	70.1	0.78	497.3 ± 39.6	270.9	91.1	2T
66	9665	22.412	0.0069	52.6	31.4	0.60	59.9	0.78	433.5 ± 34.5	227.8	97.0	2T
66	9666	30.381	0.0067	69.4	65.6	0.95	84.8	0.78	425.7 ± 33.2	212.2	102.5	2T
66	9667	110.589	0.0121	174.8	77.7	0.44	193.1	0.83	378.1 ± 30.6	245.6	132.1	2T
69	9677	78.620	0.0088	397.1	213.1	0.54	447.2	0.79	163.6 ± 13.4	273.8	96.2	2T
69	9678	33.264	0.0061	151.3	47.8	0.32	162.5	0.78	270.8 ± 22.4	217.3	93.9	2T
69	9679	56.151	0.0068	320.2	273.4	0.85	384.5	0.78	174.0 ± 13.9	239.9	92.3	2T
70	9680	29.724	0.0083	60.5	39.9	0.66	69.8	0.79	411.3 ± 32.7	252.8	99.2	2T
70	9682	161.661	0.0179	165.4	37.8	0.23	174.2	0.85	413.2 ± 34.0	296.1	141.1	2T
70	9683	94.691	0.0089	186.0	83.6	0.45	205.7	0.80	412.1 ± 33.2	252.2	104.3	2T
70	9684	37.823	0.0082	86.5	48.8	0.56	98.0	0.80	376.0 ± 30.2	216.8	115.4	2T
Fish Canyon	9745	11.142	0.0076	350.2	174.8	0.50	391.3	0.79	30.7 ± 2.6	242.0	98.2	-
Fish Canyon	9746	8.824	0.0086	284.1	156.2	0.55	320.8	0.80	26.4 ± 2.2	240.3	106.6	-
Fish Canyon	9747	6.546	0.0059	304.2	100.0	0.33	327.7	0.78	28.0 ± 2.4	208.1	95.7	-

Notes: Clear, euhedral, non-fractured zircons with two terminations were separated manually by hand-picking under an Olympus SZX12 binocular microscope. Zircon grain geometries were imaged and measured microscopically and the loaded into acid-treated platinum capsules. Zircons were then outgassed under vacuum at ~1300°C for 20 minutes, using a semiconductor Coherent Quattro FAP 820 ηm diode laser with a fibre-optic coupling to the sample chamber. He volume was determined by isotope dilution using a pure ³He spike, calibrated against an independent ⁴He standard and measured using a Balzers quadrupole (QMS 200 – Prisma) mass spectrometer. A hot blank was run after each gas extraction to verify complete outgassing of the apatite grains. Outgassed zircon grains were transferred from their Pt capsules into Parr bombs, where they were spiked with ²³³U and ²²⁹Th and digested at 240°C for 40 hours in HF. Fish Canyon Tuff zircon (Gleadow et al., 2015) was also run as an ‘unknown’ with each batch zircon samples and served as a check on sample accuracy. Standards and unspiked reagent blanks, were treated identically as unknowns. Samples underwent a second bombing in HCl for 24 hours at 200°C, ensuring dissolution of fluoride salts. After drying, samples were dissolved in HNO₃ solutions and diluted in H₂O to 5%. For single crystals digested in small volumes (0.3–0.5 ml), U and Th isotope ratios were measured to a precision of up to 2%, but typically better than 1% on an Agilent 7700x ICP-MS. Overall precision for the ZHe method is ~6% or less. α-emission corrected ZHe ages were calculated following the approach of Hourigan et al. (2005).

^aeU = U ppm + Th ppm * 0.235

^bF_r is the α-ejection correction after Farley et al., (1996).

^cGrain morphology - 0T = no crystal terminations, 1T = one crystal termination and 2T = 2 crystal terminations.

Table 5.2. Lapur Range Apatite Fission Track Data Summary

Sample No.	Rock Type	Longitude/Latitude	Elevation [m]	No. of grains	N _s	ρ _s [10 ⁵ cm ⁻²]	²³⁸ U [ppm ± 1σ]	Cl [wt%]	Dispersion [%]	Pooled age [Ma ± 1σ]	Central age [Ma ± 1σ]	N _{length}	Mean track length [μm ± se]	c-axis corrected mean track length [μm ± se]	St.Dev. [μm]
56	Mozambique Belt, bt paragneiss w/ amphibolite	35.8354/4.3815	462	48	954	2.3254	59 ± 25	0.17	19	8.8 ± 0.7	9.2 ± 0.5	117	12.7 ± 0.2	13.9 ± 1.4	2.13
58	Mozambique Belt, bt gneiss w/ amphibole lenses	35.8326/4.3823	506	46	108	0.4476	10 ± 16	0.09	43	9.7 ± 1.8	12.8 ± 1.6	55	11.2 ± 0.5	12.9 ± 2.1	3.38
61	Mozambique Belt, brecciated gneiss	35.8354/4.3783	453	58	606	1.1838	29 ± 27	0.20	40	9.1 ± 1.1	10.0 ± 0.7	122	13.2 ± 0.2	14.2 ± 1.4	1.93
62	Mozambique Belt, bt-ms mylonitic gneiss	35.8218/4.3593	480	59	286	0.7585	19 ± 15	0.03	41	8.4 ± 1.2	9.9 ± 0.9	99	12.1 ± 0.3	13.5 ± 1.6	2.69
63	Mozambique Belt, ms mylonitic gneiss	35.8195/4.3338	474	52	141	0.7060	15 ± 14	0.01	36	10.6 ± 1.7	14.1 ± 1.5	92	11.9 ± 0.3	13.2 ± 1.5	2.37
64	Mozambique Belt, quartzite w chl ms bt schist	35.8129/4.3053	505	34	80	0.4065	10 ± 18	0.01	61	9.3 ± 4.5	14.4 ± 2.6	99	12.3 ± 0.3	13.7 ± 1.6	2.57
65	Lapur Sandstone, granules sst	35.8200/4.2718	485	30	515	1.9730	48 ± 53	0.14	23	9.0 ± 0.8	9.0 ± 0.6	106	12.9 ± 0.2	14.0 ± 1.5	2.28
66	Lapur Sandstone, sst	35.8078/4.2674	530	31	161	0.7165	19 ± 26	0.11	0	8.5 ± 1.3	10.0 ± 1.0	116	11.8 ± 0.3	13.3 ± 1.8	3.09
69	Lapur Sandstone, sst	35.7970/4.2708	556	19	52	1.1294	29 ± 37	0.11	47	10.1 ± 3.4	12.5 ± 2.6	19	10.7 ± 0.7	12.5 ± 2.0	3.06
70	Lapur Sandstone, arkosic sst	35.8326/4.1957	507	23	157	1.4880	38 ± 26	0.11	0	8.0 ± 0.7	8.7 ± 0.7	23	12.4 ± 0.5	13.7 ± 1.6	2.59

Notes: Apatite grains were mounted and polished, and fission tracks analysed using the LA-ICP-MS technique following procedures described in Gleadow et al. (2015). Analyses were performed by Samuel C Boone on image sets captured by *TrackWorks* using a 3.2MP AVT Oscar F-320C camera mounted on a Zeiss AxioImager microscope with a 1000x total magnification and a 100x dry objective (calibration = 0.07 μm by 0.07 μm per pixel). Spontaneous track densities were measured on prismatic internal apatite surfaces after etching with 5M HNO₃ for 20sec at 20°C.

Track counts were obtained by automated counting in *FastTracks* using the 'coincidence mapping' technique of Gleadow et al. (2009) followed by manual inspection. Uranium concentrations of each grain were determined by LA-ICP-MS single spot analysis using a New Wave Nd:YAG Laser (λ=213nm with 5Hz @ 45% power, spot size=30μm) connected to an Agilent 7700 mass spectrometer. NIST612 was used as the primary reference material and a sintered Mud Tank Carbonatite apatite as an in-house secondary reference material during LA-ICP-MS analyses, following the protocol of Boone et al. (2016). Data reduction of LA-ICP-MS results were performed using the TraceElementst_IS data reduction scheme in software package *Iolite* (Paton et al., 2011). Single grain and pooled ages were calculated according to Hasebe et al. (2004). Central ages were estimated from single grain ages and errors according to the formulas given by Galbraith (2005, p.100) using the Newton-Raphson method. All ages are "model" ages obtained using a range factor (Rs) of 7.17μm (average mean track length of Durango and Fish Canyon Tuff standards) and are directly comparable to conventional External Detector Method ages (Seiler et al., 2014). Confined track lengths (TINTs) were measured as true 3D lengths using *FastTracks* after irradiation by ²⁵²Cf and are corrected for a refractive index of 1.634 for apatite. c-axis projected confined track lengths (Ketcham et al., 2007b) were used for thermal history modelling (Section 3.5). Chlorine concentrations of individual apatite grains were measured using a Cameca SX50 electron microprobe.

N_s = number of spontaneous tracks counted; ρ_s = spontaneous track density; N_{length} = number of lengths measured; se = standard error, bt = biotite, ms = muscovite, chl = chlorite, sst = sandstone, w/ = with

Table 5.3. Lapur Range Apatite (U-Th-Sm)/He Data Summary

Sample No.	Lab #	⁴ He [ncc]	Mass [mg]	U [ppm]	Th [ppm]	Sm [ppm]	Th/U ratio	[eU] (ppm) ^a	F _T ^b	Corrected Age [Ma ± 1σ]	Grain length [μm]	Grain half-width [μm]	^c Grain morphology
56	9709	0.059	0.0065	23.7	1.1	21.3	0.05	23.96	0.78	4.0 ± 0.4	235.4	52.3	OT
56	9710	0.078	0.0121	19.3	3.2	33.1	0.16	20.05	0.81	3.3 ± 0.3	306.6	62.6	OT
56	9711	0.024	0.0071	11.2	3.8	83.8	0.34	12.09	0.77	3.0 ± 0.3	286.5	49.6	OT
56	9712	0.024	0.0033	29.0	3.2	47.3	0.11	29.75	0.71	2.8 ± 0.3	216.7	39.0	OT
58	9749	0.064	0.0239	6.1	0.9	96.0	0.14	6.31	0.83	4.2 ± 0.4	293.2	68.5	1T
58	9751	0.081	0.0338	5.1	0.8	90.9	0.16	5.29	0.87	4.2 ± 0.4	399.8	91.7	OT
61	9717	0.077	0.0150	15.7	3.7	41.7	0.23	16.57	0.83	3.1 ± 0.3	298.6	70.7	OT
61	9719	0.075	0.0121	14.3	3.8	55.3	0.27	15.19	0.80	4.2 ± 0.4	359.7	58.0	OT
61	9720	0.071	0.0091	17.4	4.6	88.0	0.26	18.48	0.80	4.4 ± 0.4	252.5	59.9	OT
62	9721	0.252	0.0084	22.8	25.9	79.1	1.14	28.89	0.80	10.7 ± 0.9	250.5	70.4	1T
62	9722	0.074	0.0173	5.5	4.9	26.6	0.88	6.65	0.84	6.3 ± 0.5	241.9	84.3	OT
62	9723	0.079	0.0088	9.5	11.1	37.6	1.17	12.11	0.79	7.7 ± 0.6	233.8	61.4	OT
62	9724	0.025	0.0085	8.1	3.1	89.4	0.38	8.83	0.79	3.4 ± 0.3	239.5	59.6	OT
64	9725	0.035	0.0206	2.9	2.2	46.4	0.75	3.42	0.85	4.8 ± 0.4	218.4	96.9	OT
64	9727	0.077	0.0134	4.5	3.9	68.5	0.85	5.42	0.82	10.5 ± 0.9	253.0	72.7	OT
64	9728	0.092	0.0073	5.5	7.3	107.0	1.32	7.22	0.78	18.2 ± 1.4	143.8	71.0	OT
65	9730	0.116	0.0135	7.0	23.7	266.9	3.37	12.57	0.83	6.7 ± 0.5	281.3	85.1	1T
65	9731	0.042	0.0074	2.2	25.5	157.3	11.54	8.19	0.77	7.3 ± 0.6	264.7	63.1	1T
65	9732	0.089	0.0076	17.9	3.8	36.4	0.21	18.79	0.81	6.3 ± 0.5	218.2	73.7	1T
65	10370	0.024	0.0078	3.5	1.9	26.4	0.54	3.95	0.80	8.1 ± 0.7	155.2	70.6	OT
66	9733	0.03	0.0079	5.6	16.3	217.3	2.92	9.43	0.80	4.0 ± 0.3	217.5	75.5	1T
66	9734	0.139	0.0355	8.0	7.8	138.2	0.97	9.83	0.88	3.7 ± 0.3	422.3	110.7	1T
66	9735	0.076	0.0125	18.4	6.5	77.7	0.36	19.93	0.83	3.1 ± 0.3	203.1	78.1	OT
69	9737	0.063	0.0087	5.2	1.8	99.1	0.34	5.62	0.80	12.9 ± 1.1	163.5	72.9	OT
69	9738	0.235	0.0043	76.6	8.3	7.0	0.11	78.55	0.76	7.5 ± 0.7	154.3	52.9	OT
69	9739	0.029	0.0058	2.6	5.5	118.2	2.08	3.89	0.76	13.4 ± 1.0	169.0	58.6	OT
69	9740	0.16	0.0075	19.8	9.6	157.6	0.49	22.06	0.79	10.1 ± 0.8	157.2	68.8	OT
70	9741	0.277	0.0179	38.8	1.9	35.0	0.05	39.25	0.84	3.9 ± 0.3	334.0	73.1	OT
70	9743	0.104	0.0095	20.3	12.2	92.5	0.60	23.17	0.82	4.7 ± 0.4	232.6	80.1	1T
70	9744	0.056	0.0044	40.5	12.2	35.6	0.30	43.37	0.76	3.2 ± 0.3	214.8	54.4	1T
Durango	9748	10.264	-	-	-	-	19.30	-	1.00	32.9 ± 2.0	-	-	-
Durango	9753	6.605	-	-	-	-	17.42	-	1.00	32.8 ± 2.0	-	-	-

Notes: Clear, euhedral, non-fractured apatites were separated manually by hand-picking under an Olympus SZX12 binocular microscope. Apatites, in acid-treated platinum capsules, were outgassed under vacuum at ~900°C for 5 minutes, using a semiconductor Coherent Quattro FAP 820 nm diode laser with fibre-optic coupling to the sample chamber. He volume was determined by isotope dilution using a pure ³He spike, calibrated against an independent ⁴He standard and measured using a Balzers quadrupole (QMS 200 – Prisma) mass spectrometer. A hot blank was run after each gas extraction to verify complete outgassing of the apatite grains. The second re-extract contributed less than 0.5% of the total measured ⁴He for all samples. ²³⁸U, ²³⁵U, ²³²Th and ¹⁴⁷Sm content was obtained by total dissolution of outgassed apatites, still in platinum capsules, in HNO₃ and analysed using an Agilent 7700x ICP-MS. Durango apatite (McDowell et al., 2005) and Fish Canyon Tuff zircon (Gleadow et al., 2015) were also run as ‘unknowns’ with each batch of apatite or zircons samples and served as a check on sample accuracy. (U-Th-Sm)/He ages were calculated and corrected for α-emission following the approach of Farley et al. (1996). Analytical uncertainties, including the α-ejection correction, an estimated 5 μm uncertainty in grain dimension, gas analysis (<1%) and ICP-MS analytical uncertainties, are conservatively assessed at ~6.2%. Accuracy and precision of U, Th and Sm content range up to 2%, but are typically <1%.

^aeU (ppm) = U ppm + 0.235*Th ppm

^bF_T is the α-ejection correction after Farley et al., (1996).

^cGrain morphology - OT = no crystal terminations, 1T = one crystal termination and 2T = 2 crystal terminations.

5.3.1 Zircon (U-Th)/He (ZHe) results

Samples 56, 58 and 61, sampled from within ca. 1000m below the basement nonconformity, yield reproducible ZHe single grain ages within a range of 76 ± 6 Ma to 123 ± 10 Ma. The corresponding weighted mean ZHe ages (4 aliquots each sample) fall within 2σ error, ranging from 93 ± 8 Ma to 104 ± 9 Ma. eU contents of single grain aliquots from these samples are relatively low (90.6-244.3 ppm, Figure 5.5). Samples 62 and 64, collected ~500 and 25 m below the nonconformity and about ~1.5 km and ~8 km south along-strike of the fault, yield older ZHe weighted mean ages of 223 ± 33 Ma (10 aliquots) and 150 ± 24 Ma (4 aliquots), respectively. Sample 62 yielded nine reproducible single grain ZHe ages (206 ± 17 to 295 ± 24 Ma) with moderate eU concentrations (285.2-592.9 ppm). A tenth single grain analysis from sample 62 yields a much younger single grain age (54 ± 5 Ma) with very high eU (1763.7 ppm; Figure 5.5), suggesting that the high accumulated radiation damage of this grain has led to decreased He retentivity and a younger ZHe age. By contrast, sample 64 exhibits much greater age dispersion in single grain ZHe ages (92 ± 8 to 211 ± 18 Ma), with younger ages corresponding to higher eU concentrations (ranging between 1069.2 and 550.5 ppm; Figure 5.5).

Zircon He analyses of Lapur Sandstone samples 65, 66 and 70 yielded concordant weighted mean ages between 404 ± 10 Ma to 436 ± 23 Ma (4 aliquots each; Table 5.1). Sample 69, collected ~600 m stratigraphically above the basement nonconformity (Figure 5.4a), yielded a younger Triassic ZHe weighted mean age of 206 ± 31 Ma (3 aliquots). Single grain aliquots of this sample show a marked variation in ZHe ages (164 ± 13 to 271 ± 22 Ma) that are also inversely proportional to eU concentrations (Figure 5.5; 447.2 to 162.5 ppm).

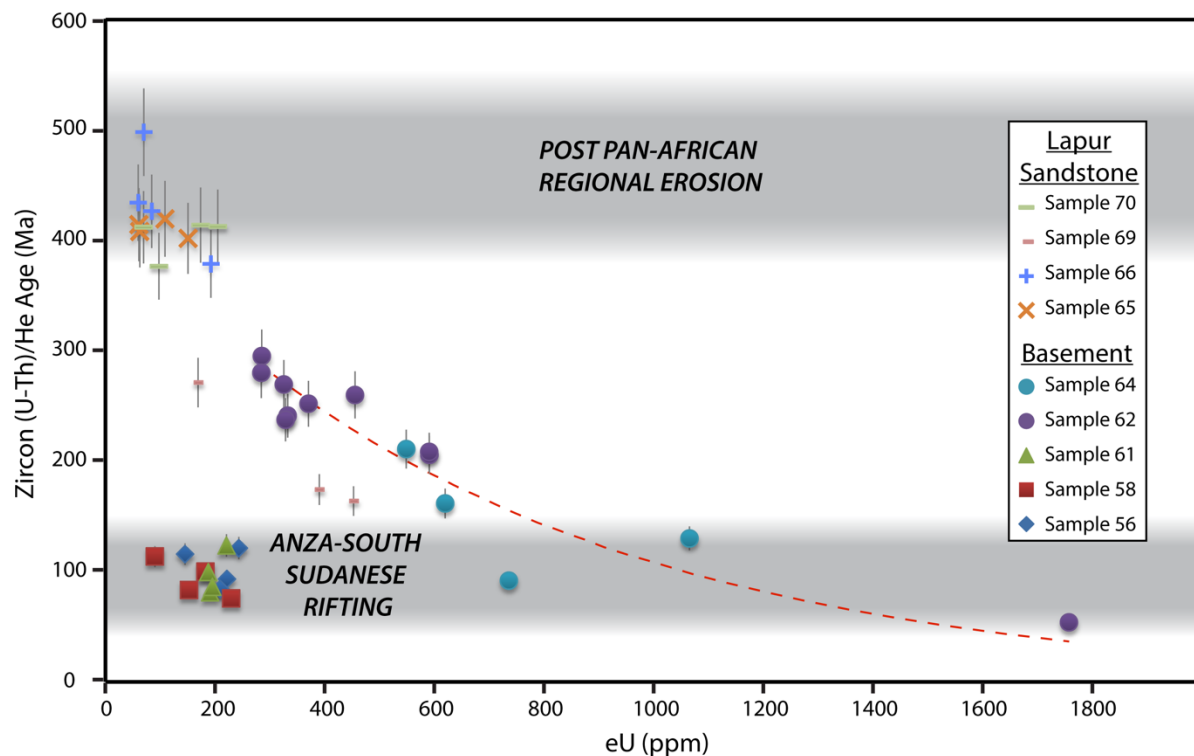


Figure 5.5. Plot of single grain ZHe ages versus effective uranium (eU) content of Lapur Range samples, with the approximate duration of relevant major regional tectonothermal events recorded in East Africa shown in grey. Basement samples 56, 58 and 61, which, prior to the mid-Miocene extension and block tilting were at structurally deepest in the basement levels, record Cretaceous single grain ZHe ages that are likely related to coeval Cretaceous-Paleogene Anza-South Sudanese rifting. The remaining samples display a range of Cambrian to Paleogene single grain ZHe ages. Some samples show a negative correlation between single grain ages and the eU content (red line), which suggests that the intermediate Permian-Triassic ZHe ages of samples 62, 64 and 69 are products of increased He diffusivity due to radiation damage effects and, thus, do not record distinct tectonothermal event(s) at those times.

5.3.2 Apatite fission track (AFT) results

AFT analyses from Lapur Range basement samples record middle-late Miocene ages (Table 5.2), showing a trend of apparently older ages closer to the basement nonconformity (14.4 ± 2.6 Ma) and younger ages (9.2 ± 0.5 Ma) further north, at greater crustal depths. MTL are moderate to relatively long and range from 11.9 to 13.2 μm , with moderate standard deviations (Table 5.2 & Figure 5.4; 1.93 to 2.69 μm). An exception to this range is sample 63, which yielded a shorter MTL (11.2 μm) and higher standard deviation (3.4 μm). However, this sample yielded about half the number of confined track measurements ($n=55$) obtained for other samples (> 100), making its MTL statistically less robust. Chlorine concentrations are consistently low, ranging from 0.01 to 0.20 wt%.

Similar to samples from the underlying basement, AFT ages from the Lapur Sandstone range from 9.0 ± 0.6 to 12.5 ± 2.6 Ma, with MTL of 10.7-12.9 μm and standard deviations of 2.59-3.09 μm (Table 5.2). Lapur sandstone apatites have low average chlorine compositions, ranging from 0.11 to 0.14 wt%.

5.3.3 Apatite (U-Th-Sm)/He (ZHe) results

Single grain AHe analyses from Lapur basement samples (56, 58, 61 and 62) yield ages of 3.4 ± 0.3 to 18.2 ± 1.4 Ma, with samples from shallower crustal levels yielding older AHe ages than those from structurally deeper levels (Figures 5.4a and 5.6). eU concentrations are relatively low (3.4-28.9 ppm; Table 5.3) and show no correlation to AHe age.

Lapur sandstone samples yielded similar AHe ages ranging from 3.1 ± 0.3 Ma to 13.4 ± 1.0 Ma, which exhibit a slight negative correlation with eU concentrations (3.9-78.6 ppm).

5.3.4 Summary of Results

New ZHe, AFT and AHe data from the Lapur Range in the northern Turkana Depression reveal multiple periods of cooling. ZHe results yield a wide range of dispersed ages that, upon further review, exhibit lithological controls, depth dependency and a correlation to accumulated radiation damage. Lapur basement samples collected from the deepest level below the top basement nonconformity (56, 58 and 61) record Cretaceous cooling (~ 125 -75 Ma) through the ZHe PRZ. In contrast, shallower basement samples (62 and 64) yield a wide range of dispersed ages that negatively correlate to eU content (Figure 5.5). These consist of 4 younger early Late Jurassic-early Paleogene aliquots (54-161 Ma, 1763.7-621.8 ppm) that are in general agreement with ZHe results from deeper basement samples and 10 older Triassic ages (206-295 Ma, 592.9-285.2 ppm). Many authors have noted negative age-eU correlations in previous ZHe studies (e.g. Nasdala et al., 2004; Reiners, 2005; Ketcham et al., 2013), however the exact nature of this relationship remains poorly understood. Guenthner et al. (2013) suggested that the correlation is a product of intra-grain differences in accumulated radiation damage in zircons. In certain cases, such variations result in varying He diffusion rates (or He retentivity) between aliquots, effectively lowering the temperature range of the PRZ for grains with significant radiation damage (e.g. Johnson et al., 2017; Mackintosh et al., 2017). The age-eU relationship observed in shallower Lapur basement samples is interpreted as indicating age dispersion as the product of variations in He diffusivity related to different amounts of α -radiation damage accumulation. In this case, younger aliquots record fully

rejuvenated ages despite cooling from the same temperatures as their intra-sample counterparts, as their ZHe PRZ has been effectively lowered by the effects of radiation damage. By contrast, older aliquots record partial overprinting during cooling from temperatures within their ZHe PRZ during a later cooling phase, such as Cretaceous cooling recorded by deeper basement samples that postdate their Triassic ages.

Lapur Sandstone ZHe ages fall into two populations: twelve zircon ages from samples 65, 66 and 70 yielding late Cambrian to Devonian ages (eU: 61.3-205.7 ppm) and three younger aliquots from sample 69 (eU: 162.5-447.2 ppm), yielding two Jurassic ages and a Permian age. All ZHe ages predate Late Cretaceous-Eocene Lapur Sandstone sedimentation and, thus, record thermal histories inherited from their respective source provenance(s). The low eU concentrations of aliquots from the former group suggest their late Cambrian to Devonian ages may record cooling of the source rocks through the ZHe PRZ during this time. The younger aliquots from sample 69 may record separate periods of cooling through the ZHe PRZ at times corresponding to their respective ZHe ages. Alternatively, their younger ages may be a result radiation damage-related He diffusion that postdates late Cambrian to Devonian cooling recorded by other Lapur Sandstone zircons.

AFT results are summarized in a boomerang plot (Figure 5.4b), which are often used to illustrate the relationship between fission track age and track length distribution for a set of samples that have experienced a multi-phase thermal history (e.g. Gallagher and Brown, 1997). Samples that record the earlier rapid cooling event yield older AFT ages and long MTL. Whereas, samples that record the subsequent rapid cooling event yield younger AFT ages and long MTL. By contrast, samples which record both thermal events yield intermediate, mixed AFT ages and shortened MTL resulting from partial annealing during the later rapid cooling phase, thus forming a boomerang shape. AFT analyses from this study are interpreted to represent the ascending left-hand portion of a boomerang plot, with basement and Lapur Sandstone samples recording moderate to relatively long MTL (10.7-13.2 μm) and relatively young middle Miocene (9-14 Ma) ages. This suggests that the earlier thermal history of the Lapur Range recorded by ZHe analyses was substantially overprinted by a later cooling phase that, at least partially, postdates the central AFT ages (i.e. < 9 Ma). This is in agreement with their middle Miocene-Pliocene AHe ages (Table 5.3). Thus, the low-temperature thermochronology data from the Lapur Range indicate that it underwent two

episodes of cooling: Cretaceous cooling prior to Lapur Sandstone deposition and a middle Miocene-Pliocene phase.

5.3.5 Thermal History Reconstructions

In order to quantify time-temperature histories for Lapur Range samples, ZHe, AFT and AHe data were inversely modelled (Figure 5.6) using *QTQt* (Kerry Gallagher, 2012), with the fission track annealing model of Ketcham et al. (2007a) on the basis of *c*-axis projected track lengths (Ketcham et al., 2007b). Zero termination AHe grains (see Table 5.3) were omitted during modelling, as the current version of *QTQt* does not accept this morphology. Radiation damage models were employed for modelling of AHe data (Flowers et al., 2009) and ZHe data (Guenther et al., 2013) to account for radiation-enhanced annealing as recorded by age-eU relationships. Two time-temperature constraints were applied to the models. The first constraint spans 100-35 Ma, which represents deposition of the Late Cretaceous to middle Paleogene Lapur Sandstone, at which time samples are expected to have resided at near-surface temperatures (10-30 °C). The second requires the samples to be at surface temperatures (20 ± 10 °C) in the present day.

Samples were grouped by lithology (Lapur basement and Lapur Sandstone) during thermal history modelling. Stratigraphic contacts and bedding relationships (Figure 5.4a) suggest that the current tilted fault block geometry of the Lapur Range and Turkana Basin is the result of Neogene extension. Therefore, Lapur basement samples were modelled together as a pseudo-vertical profile based on the approximate pre-EARS (i.e. tilt-corrected) position of samples relative to the Cretaceous erosional nonconformity (Figure 5.4a). Lapur Sandstone samples were also modelled collectively in a similar fashion as basement samples were, whilst allowing for independent pre-depositional t-T paths. This assumes that for any individual sandstone sample, all its respective detrital aliquots experienced a similar pre-depositional thermal history. In the case of samples 65, 66 and 70, the very low inter-aliquot ZHe age dispersion suggests this assumption is valid. This assumption, however, is perhaps invalid for sample 69, whose zircons may record multiple, distinct pre-depositional thermal evolutions (Section 5.3.4).

Preliminary thermal history modelling of Lapur Sandstone samples suggested a post-depositional thermal history consisting of Late Cretaceous-Paleogene reheating followed by Neogene-Recent cooling to surface temperatures (see below; Section 5.3.5.1.3 and 5.3.5.1.4).

While the preliminary results were consistent with those of independently modelled basement samples in both character and timing, the maximum paleo-temperatures predicted for the former exceeded those of the basement, which is inconsistent with stratigraphic relationships. Hence, Lapur Sandstone samples were then remodelled with an additional t-T constraint (30-10 Ma, 90-120 °C).

5.3.5.1 Time-temperature histories

In general, time-temperature reconstructions match the observed data very well. The exception being ZHe data from basement samples 62, 64 and 69 (Figure 5.6). Sample 64 exhibits a very large spread in aliquot ages that are likely related to radiation damage as discussed above. While the zircon radiation damage accumulation annealing model (ZRDAAM) model of Guenthner et al. (2013) was employed to account for radiation damage during modelling, the model predicts only a moderate amount of inter-grain age variation (~50 Ma), much smaller than observed variations (~120 Ma). Despite these differences, the 1-1 correlation in the predicted vs observed age plot does pass through the spread of aliquots. The ZRDAAM model did a similarly poor job of predicting the oldest ZHe age from sample 69. The largest incongruence however was found between predicted and observed single grain ages for aliquots of sample 62, suggesting that perhaps the ZRDAAM model is unable to fully account for wide ranges of accumulated radiation damage when predicting ZHe ages. Alternatively, the observed dispersion may be the result of intragrain zonation in U and Th. Similar inconsistencies have been reported by Guenthner et al. (2014), where the ZRDAAM model was unable to account completely for intra-sample ZHe age variation.

5.3.5.1.1 Pre-Cretaceous Thermal History

Thermal history modelling results suggest that the Lapur basement may have undergone a period of gradual heating beginning in the Precambrian, with all but the shallowest sample (64) reaching temperatures > 200 °C by the Early Cretaceous. However, given the total thermal resetting of samples 56, 58 and 61 and almost complete thermal overprinting of samples 62 and 64, the pre-Cretaceous thermal history is not constrained by the data. Eight of the predicted single grain ZHe values for sample 62 are significantly lower than the observed ages (Figure 5.6), indicating that sample 62 also resided at temperatures within the ZHe PRZ prior to the middle Cretaceous, as opposed to residing at temperatures above the PRZ, as suggested by the thermal history model.

Thermal history reconstructions for the Lapur Sandstone samples consist of an array of best fit t-T paths that seem to suggest monotonic gradual cooling from paleotemperatures ranging between ~20-170 °C in the early Paleozoic and reaching near-surface temperatures by the Upper Cretaceous when detrital zircon grains were redeposited as the Lapur Sandstone (Figure 5.6).

5.3.5.1.2 Early-Late Cretaceous Cooling

Between ~120 and 110 Ma, Lapur basement thermal history models record a period of relatively rapid monotonic cooling (~15 °C/Myr), represented by t-T paths for deeper basement samples (56, 58 and 61) which pass through the ZHe PRZ at this time. This thermal history predicts totally reset middle Cretaceous ZHe ages, in agreement with observed data. During this time, shallower samples (62 and 64) cooled from temperatures near or within the upper end of the ZHe PRZ (between ~180-200 °C) to near-surface temperatures. Best-fit t-T paths then transition to a period of very gradual cooling (~0.25 °C/Myr) or thermal stability into the early Late Cretaceous (~90-70 Ma). By the Turonian-early Campanian (~93-75 Ma), the shallowest basement samples were near surface, coinciding with deposition of the overlying Lapur Sandstone.

5.3.5.1.3 Late Cretaceous-earliest Neogene Reheating

Between ~80 and 60 Ma, all Lapur Range basement and sandstone samples transitioned to a period of moderately rapid reheating (~1.7 °C/km). Thermal history models predict that by the early Miocene (between ~20-14 Ma), samples reached temperatures between 100 and 170°C depending on their relative burial depth. Although time-temperature models appear to suggest that this episode of Paleogene reheating was monotonic and linear, this portion of the Lapur thermal history is not well constrained due to substantial rejuvenation of AFT data prior to Miocene times. Thus, a more complex Paleogene thermal history is possible.

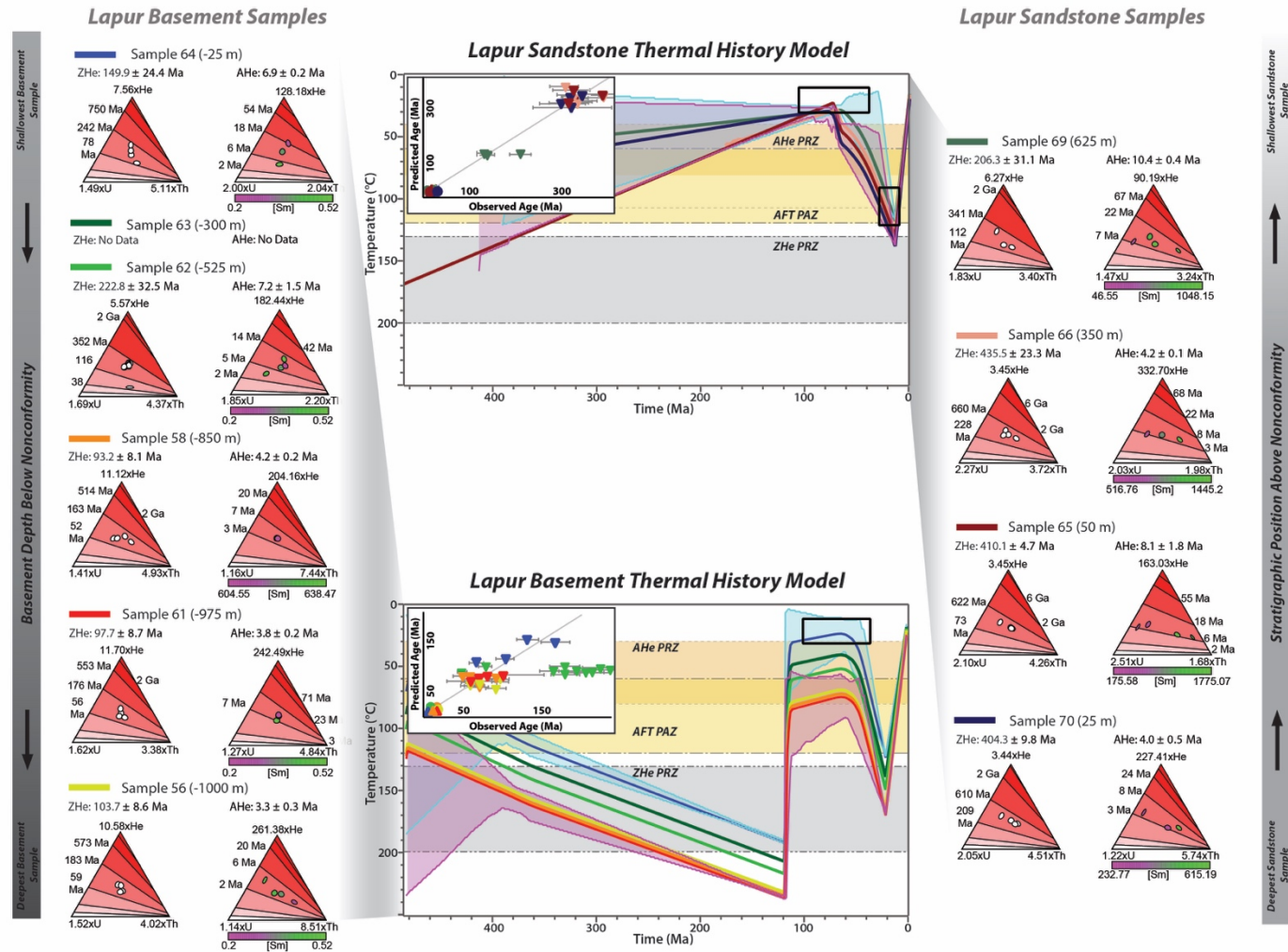


Figure 5.6. ZHe and AHe data and inverse thermal history models of Lapur basement samples (left side and bottom model) and Lapur Sandstone samples (right side and top model). Single sample ZHe and AHe data are arranged by relative stratigraphic depth relative to the top basement nonconformity prior to mid-Miocene tilting of fault blocks, with deepest samples at the bottom. Ternary plots for ZHe (left column) and AHe (right column) data, generated with HelioPlot (Vermeesch 2010), display the relationship between ages and U, Th and He concentrations for individual aliquots. The colour of single grain AHe ages shifts towards green with increasing Sm content. Corresponding central He ages ($\pm 1\sigma$) are listed above the ternary diagrams. The approximate temperature sensitivity range for each thermochronometer is highlighted in the thermal models. Applied time-temperature constraints are marked by black boxes (see Section 5.3.5). The best-fit t-T path for each individual sample is shown in a colour corresponding to that of the line displayed above its He data. In each vertical profile the envelope of acceptable t-T paths for the uppermost (blue path) and lowermost (red path) sample shown by light blue and purple envelopes, respectively. A plot of values predicted by thermal models versus observed values are shown for each model. ZHe data = inverted triangles, AHe data = upright triangles and AFT data = circles.

5.3.5.1.4 Middle Miocene-Pliocene Cooling

Between ~20 to 14 Ma, thermal history models indicate a transition to rapid cooling (6-10 °C/Myr) that is well constrained by the data. Basement samples had passed through the AFT PAZ and AHe PRZ by the end of the Pliocene (between 5-2 Ma) and attained near-surface temperatures (~20°C) by the present-day. Close inspection reveals that Lapur basement thermal history models appear to suggest that the onset of rapid cooling may have occurred as early as ~20 Ma, although this portion of the thermal history model is below the upper-temperature portion of the PAZ (~100-120°C) and thus not constrained by the basement data. By comparison, thermal history models from the Lapur Sandstone suggest that Neogene cooling likely commenced between ~15 and 14 Ma. This is consistent with observed MTL for these samples, which are of intermediate length (11.8-13.2 μm for samples with a statistically significant number of measurements; Table 5.2), suggesting that the Lapur Sandstone has not been fully reset and thus records the entire Neogene cooling history from maximum paleotemperatures (< ~120 °C).

5.3.5.2 Paleogeothermal gradients in Turkana

When interpreting low-temperature (up to ~200°C) thermal history reconstructions to constrain geological processes in the upper crust, temperature is generally used as a proxy for depth (Kohn et al., 2005). Therefore, when making geological inferences from t-T models, accurate assumptions must be made regarding paleo-heat flow to robustly distinguish between tectonic processes, such as long-term denudation and burial rates, and changes in the upper crustal thermal regime (K. Gallagher & Brown, 1999; A. J. W. Gleadow & Brown, 2000).

Heat flow measurements in the Loperot-1 well, ~110 km south of the Lapur Range, indicate that the present day geothermal gradient in Turkana is very high (42 ± 5 °C /km) (Morley, Stone, et al., 1999). This is attributed to a combination of factors including local volcanic activity, very thin crust (~20 km) and, as a consequence, the shallowness of the brittle-ductile transition (Morley, Stone, et al., 1999). The presence of the African Superplume, an elongate mantle thermal anomaly beneath East Africa, may also be a contributing factor to the observed high geothermal gradient (e.g. Hansen et al., 2012). Although it is uncertain when this elevated thermal regime was first initiated, some first-order temporal constraints can nevertheless be made from indirect evidence. Hendrie et al. (1994) showed that the ~35-40 km of extension that occurred since the late Eocene-Oligocene can account for the degree of

crustal thinning observed in Turkana (Hendrie et al., 1994). The presence of latest Eocene-Oligocene volcanism in Turkana (Hendrie et al., 1994; Morley et al., 1992; Wescott et al., 1999) and a proposed Oligocene-early Miocene timing for development of the Ethiopian and East African Domes in response to thermal uplift (Pik et al., 2008; Smith & Mosley, 1993; Wichura, Bousquet, Oberhänsli, Strecker, & Trauth, 2010) suggest a late Paleogene arrival of the African Superplume beneath Turkana.

Thus, when interpreting thermal history models in this study and estimating amounts of denudation, a Mesozoic-early Paleogene geothermal gradient of 25-35 °C/km is used, consistent with the average geothermal gradients for Archean-Proterozoic shields and mobile belts of East Africa (Nyblade, 1990). For late Paleogene-Recent times, a broader geotherm range of 25-45 °C/km is used to allow for the development of the elevated present-day heat flow and spatial and temporal variation within Turkana.

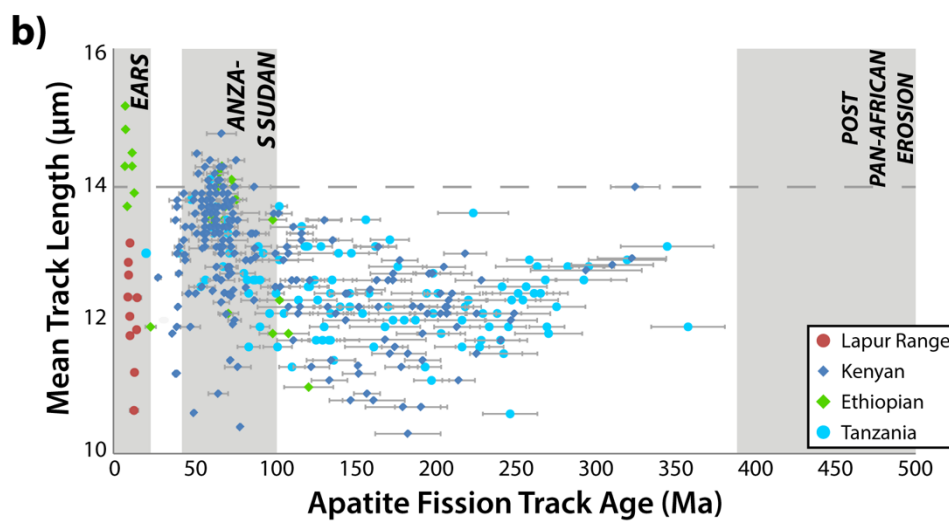
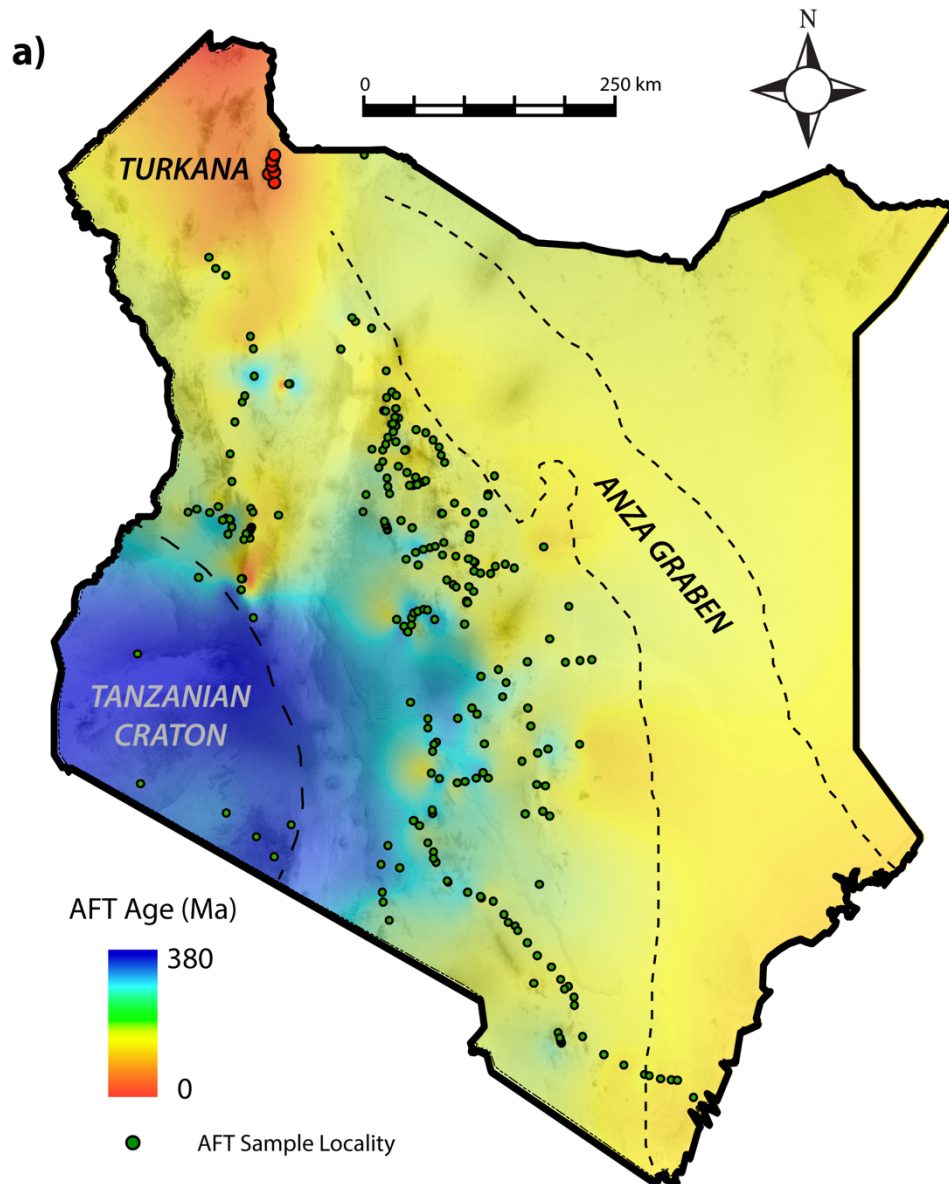


Figure 5.7 (Previous page). Image showing Lapur Range apatite fission track (AFT) ages within a regional context. **a)** Interpolation of Kenyan AFT data from previous studies (Gleadow, 1980; Foster and Gleadow, 1992, 1996; Wagner et al., 1992b; Spiegel et al., 2007; Majer-Kielbaska, 2015; Torres Acosta et al., 2015) and data presented in this paper superimposed on a DEM image. Samples from the Tanzanian Craton exhibit AFT ages >250 Ma, while most of Kenya records Late Cretaceous-early Paleogene AFT ages related to uplift and denudation of the Cretaceous-Paleogene Anza Rift margin. All samples from the Lapur Range of the Turkana Depression (this study; red circles) exhibit young Miocene AFT ages. The small orange area in the Kenyan Rift, just north of the Tanzanian Craton, corresponds to samples that yield Eocene-Oligocene AFT ages, interpreted to represent cooling from the lower PAZ during Neogene EARS-related extension (Torres Acosta et al., 2015). Sample localities are represented by green dots. **b)** Boomerang plot of AFT ages versus MTL from all Kenyan (see references for Figure 7a), Ethiopian (Abebe et al., 2010; Balestrieri et al., 2016; Philippon et al., 2014), and Tanzania (Noble, 1997; Noble et al., 1997; van der Beek et al., 1998). AFT studies. See Section 5.3.4 for a description of a Boomerang Plot.

5.4 Discussion

The Lapur Range is one of the best-exposed basement blocks in the Turkana Depression, making it an important area to address the question of how this long-lived portion of the EARS has been affected by its superposition with the late Mesozoic-early Cenozoic Anza-South Sudan Rift Systems. Thermochronological data and time-temperature history modelling reveal a complex cooling history spanning Early Cretaceous to Recent time. We suggest that these phases of cooling and reheating are associated with specific periods of regional and local tectonic activity.

5.4.1 Rapid Early Cretaceous cooling of the Lapur basement

ZHe ages (samples 56, 58 and 61 and high-eU aliquots from samples 62 and 64) and thermal history modelling suggest that the basement rocks of the Lapur Range experienced pronounced and rapid cooling during the Early Cretaceous, beginning sometime between ~120 and 100 Ma. The presence of a top basement erosional unconformity at the contact with the Lapur Sandstone and minor degree of volcanism associated with Anza Rift extension (Morley et al., 1999a) suggest that this cooling was largely a result of denudation.

Wells in the Chalbi Desert (Figure 5.2) encountered substantial thicknesses of Neocomian-Barremian (up to 695 m) and Aptian (up to 1100 m) sediment (Morley, Bosworth, et al., 1999). The Aptian-Albian was also an important period of source rock deposition in the Muglad Basin of South Sudan (up to 3,353 m thick; Figure 1) (Schull, 1988). Thus, while the

Lapur area, according to low-temperature thermochronology data presented here, was experiencing Early Cretaceous exhumation, the Anza and South Sudan Rifts were subsiding. This suggests that the northern Turkana area may have acted as a basement high between the Anza and South Sudan Rifts, as well as an axial sediment source for those grabens. However, the scale of exhumation is substantial. Assuming that erosion was the main cooling mechanism at play, then the ~170-180 °C of Early Cretaceous cooling recorded by thermal history models would suggest between ~5 and 7 km of denudation (based on a reasonable geothermal gradient of 25-35 °C/km as discussed earlier). This Early Cretaceous denudation estimate would be significantly reduced if the paleogeothermal gradient in Turkana was higher due to factors such as thermal blanketing effects since removed (e.g. Łuszczak et al., 2017), low thermal conductivity overburden (e.g. Braun et al., 2016) or elevated crustal heat flow that are not taken into consideration.

5.4.2 Late Cretaceous-earliest Neogene subsidence: Deposition of the Lapur Sandstone and emplacement of the Turkana Volcanics

Lapur basement subsidence began in the early Campanian (ca. 80 Ma), as recorded by thermal history models, with deposition of the Lapur Sandstone (Tiercelin et al., 2012). Continued burial beneath latest Eocene-Oligocene basalt flows and early Miocene rhyolites of the Turkana Volcanics reheated the Lapur basement rocks and overlying Lapur Sandstone, substantially rejuvenating their AFT ages and totally resetting their AHe ages. Thermal history reconstructions indicate that Late Cretaceous-Paleogene reheating was substantial, equivalent to burial beneath ~2-4 km of overburden assuming a geothermal gradient of 25-45 °C /km. However, only ~500 m of this can be attributed to the Late Cretaceous-early Paleogene Lapur Sandstone in the Lokitaung Gorge area and as little as 100-120 m in the northern Lapur Range (Tiercelin et al., 2012). The remaining overburden likely consisted of latest Eocene-early Miocene Turkana Volcanics, whose lateral equivalents attains thicknesses of up to 3,500 m in the neighbouring Lotikipi and Gatome Basins to the west (Wescott et al., 1999).

As the Lapur Range underwent reheating associated with subsidence and burial during the Late Cretaceous to early Paleogene, basement rocks from other parts of Kenya, southern Ethiopia and northern Tanzania appear to have experienced further cooling related to exhumation of the Anza Rift shoulders (Foster and Gleadow, 1992, 1993, 1996; Noble et al., 1997; Spiegel et al., 2007; Philippon et al., 2014; Torres Acosta et al., 2015; Balestrieri et al.,

2016). It is still unclear as to how the accommodation space was created in the Lapur area during the Late Cretaceous-early Paleogene. The Lapur Sandstone shows no evidence for rift-related faulting in outcrop, instead having been initially interpreted to represent fluvial deposits that fed laterally into the Anza Rift to the east (Morley et al., 1992). However, a more recent sedimentologic study (Tiercelin et al., 2012) showed the Lapur Sandstone paleodrainage to be far more complex, beginning in a south-southwest direction and later rotating to west-northwest to northeast for much of its history, before returning to south-southwest in the uppermost part of the type-section. Alternating paleoflow directions suggest a poorly understood paleotopographic environment in flux throughout Late Cretaceous-early Paleogene times.

Considering that the South Sudanese and Anza Rifts are thought to potentially link in the Turkana region (e.g. Fairhead, 1988; Bosworth, 1992; Hendrie et al., 1994), our data indicates a surprisingly small degree of Late Cretaceous-early Paleogene subsidence in the Lapur area. This is particularly apparent when considering that total crustal extension across the Anza Rift is estimated to be 60-65 km, equating to a Beta factor (β) of ~ 2 (Reeves et al., 1987) compared to the relatively minor degree across the Kenyan Rift ($\beta = \sim 1.15$) (Latin, Norry, & Tarzey, 1993). An erosional unconformity in the Lapur Sandstone type-section ~ 150 m above the basement nonconformity, however, may represent a period of significant erosion and removal of an unknown thickness of sediment (Abdelfettah et al., 2016), perhaps allowing for greater Upper Cretaceous sedimentation undetected by low-temperature thermochronology due to later rejuvenation of the system. The identification of thick piles of sediment in the subsurface of the northern Turkana, Gatome, and Lotikipi Basins (up to 300, 800, and 3,000 m, respectively) thought to possibly correspond to the Lapur Sandstone (Tiercelin et al., 2012; Wescott et al., 1999) suggests that Late Cretaceous-early Paleogene sedimentation may have been far more laterally extensive (~ 100 km E-W) than outcrop observations would suggest. Tiercelin et al. (2012) proposed a series of en echelon north-dipping normal faults in the Gatome Basin subsurface of suggested mid-Cretaceous age from a reinterpreted seismic line, possibly indicating that active rifting continued elsewhere in northern Turkana during the Cretaceous.

Despite significant effort, the inferred Turkana connection between the Anza and South Sudan Rifts remains speculative and enigmatic. Even if some degree of Cretaceous extension

occurred in Turkana, the amount of subsidence indicated by the low-temperature thermal evolution of the Lapur Range in this work is not consistent with a through-going NW-SE Cretaceous-early Paleogene rift of constant width and geometry. Such an initial setup was employed in the numerical and analog modelling study of Brune et al. (2017), designed to test the hypothesis that the anomalous tectonic and geomorphological evolution of the Turkana Depression resulted from the superposition of Anza-South Sudan and EARS rifting. In that study, Neogene-Recent EARS ~E-W extension was superimposed onto a geometrically homogenous NW-SE trending domain of thinned crust (30 km) extending from the Anza Graben to South Sudan. While their models reproduced some present-day morphological characteristics, such as the development of large escarpments on the NE and SW Turkana Depression margins and depicted an eastward propagation of rifting in southern Ethiopia consistent with current thinking (e.g. Ebinger et al., 2000; Balestrieri et al. 2016). However, their model also predicted an east-to-west propagation of extension in Turkana that is opposite to that observed (e.g. Morley et al., 1992; Morley, Stone et al., 1999; Wescott et al., 1999).

Instead, our findings suggest that later Paleogene tectonics may have played a more pivotal role in modifying the Turkana crustal architecture than perhaps thought. As pointed out by Tiercelin et al. (2012), the abrupt termination of Lapur Sandstone sedimentation coincided with the first eruptions between 45-35 Ma of plume-generated volcanics in southern Ethiopia (Ebinger & Sleep, 1998; Ebinger et al., 1993; R. M. George & Rogers, 2002), reaching northern Turkana by 37-35 Ma (Ebinger et al., 1993; R. George, Rogers, & Kelley, 1998; McDougall & Brown, 2009; Pik et al., 2008). The simultaneous arrival of a shallow plume beneath the Turkana Depression and an end to Lapur Sandstone deposition perhaps suggests that dynamic topography was responsible for ending early Paleogene subsidence in the Lapur Range. Thermal uplift has also been suggested to have caused inversion in the Muruanachok Basin (Morley, Stone, et al., 1999), whose eroded basin infill remnants consist of Cretaceous?-Paleogene sandstone (Figure 5.2) (Morley et al., 1992; Tiercelin et al., 2012). During the late Paleogene, widespread extension is thought to have continued in the Turkana Depression resulting in a series of basins running from the Sudanese-Ethiopian border, through Turkana and reactivated portions of the eastern Anza Rift (Figure 5.1) (Bosworth, 1992; Hendrie et al., 1994). Locally, late Paleogene extension was expressed by the development of the Lotikipi, Gatome, North Lokichar and Lokichar Basins (Morley, Stone, et

al., 1999; Wescott et al., 1999), marking a period of significant crustal thinning in Turkana (Hendrie et al., 1994).

5.4.3 Middle Miocene initiation of extension in the northern Turkana Basin

Beginning in the middle Miocene (~14 Ma), the Lapur basement and overlying Lapur Sandstone underwent a period of rapid cooling. The observed paleo-depth - AFT data relationship, whereby rocks at greater stratigraphic or structural depths yield younger AFT ages, longer MTL and lower standard deviations (Figure 5.4a), is indicative of denudation-related cooling. Thus, we interpret the observed mid-Miocene cooling of the Lapur Range to reflect flexural isostatic uplift of the MRL footwall in response to normal displacement of the hanging wall rocks. Since the MRL master fault accommodated the vast majority of Neogene extension across the basin (Abdelfettah et al., 2016), the onset of MRL isostatic footwall uplift at ~14 Ma likely indicates a middle Miocene initiation of extension across the entire northern Turkana Basin. This is supported by the presence of a pile of westward thickening sedimentary rocks overlying the Turkana Volcanics within the northern Turkana Basin hanging wall, thought to be of upper Miocene-Pliocene age (Morley, Stone, et al., 1999; Wescott et al., 1999). Our interpretation is therefore in strong agreement with the rift propagation model of Morley et al. (1999b), whereby the mid-Miocene, extension had migrated from western Turkana to the basins underlying Lake Turkana. A middle Miocene phase of extension has also been suggested for numerous other EARS basins elsewhere in the eastern Turkana Depression and the flanks of the Kenyan and Ethiopian Domes (Figure 5.2). This suggests a simultaneous migration of EARS rifting both north- and southward into the thicker crustal domains of southern Ethiopia and central Kenya at this time (Bonini et al., 2005).

Low-temperature thermochronology data indicate that basement samples from the deepest pre-EARS crustal levels (samples 56, 58 and 61) cooled from ~110-120 °C to surface temperatures by the Pleistocene-Holocene. Depending on spatial and temporal variations in Turkana geothermal gradients during the last ~14 Ma, this degree of cooling is consistent with ~2-4 km of footwall uplift in the Lapur Range. By comparison, previous AFT studies in the Kenya Rift have reported maximum middle Miocene-Recent denudational cooling associated with EARS-related extension of between 40-60 °C (Foster & Gleadow, 1996; Torres Acosta et al., 2015).

5.4.4 Lapur AFT data in a regional context: How has rift superposition affected the Turkana lithosphere's response to EARS extension?

AFT analyses from the Lapur Range yield the first Neogene apparent ages that record cooling directly related to the morphotectonic evolution of the EARS in Kenya (Figure 5.7). This finding is surprising considering the northern Turkana Basin exhibits a similar duration of EARS extension (mid Miocene-Recent), amount of extension (≤ 10 km) and similar style of faulting to many other basins in the Kenyan Rift (Ebinger et al., 2000; Morley, Ngenoh, et al., 1999; Morley, Stone, et al., 1999). The previous lack of Neogene AFT ages suggests that exhumation rates in the Kenyan Rift, south of Turkana, have been insufficient to exhume deep-seated basement rocks that record EARS-related cooling through the PAZ. The greater degree of Neogene cooling recorded in the Lapur Range is therefore attributed to either increased footwall uplift due to a difference in lithospheric response during rifting, or a relatively higher geothermal gradient in Turkana, resulting in greater cooling for an equivalent amount of denudation, or a combination of the two. A third possibility, whereby crustal heat-flow in the Turkana region decreased since the mid-Miocene could also result in a greater-degree of observed cooling. However, continued volcanic activity into late Pleistocene time (Morley, Stone, et al., 1999) and further Neogene-Pliocene extension throughout much of Turkana, resulting in further crustal thinning, makes this explanation highly unlikely.

Egan (1992) proposed that the amplitude of isostatic footwall uplift is dependent on the flexural rigidity of the lithosphere undergoing extension (Figure 5.8). The inherited thin crust and resulting high heat flow of the Turkana Depression, both in part products of its earlier Cretaceous-Paleogene rifting history, would have amplified its lithospheric response to EARS-related crustal extension, as predicted by thin plate flexure theory (Walcott, 1970). Hendrie et al. (1994), using the flexural cantilever model of continental extension (e.g. Kusznir and Ziegler, 1992), calculated an effective elastic thickness (T_e) estimate of just 3.5 km for the Turkana region, compared to ~25 km estimated for the central Kenyan Rift (Bechtel, Forsyth, & Swain, 1987; Ebinger, Deino, Drake, & Tesha, 1989). Such a large difference in T_e would allow the Turkana lithosphere to undergo approximately twice the amount of footwall uplift than the more rigid central Kenyan lithosphere, assuming a similar amount of basin extension (~10 km; Figure 5.8).

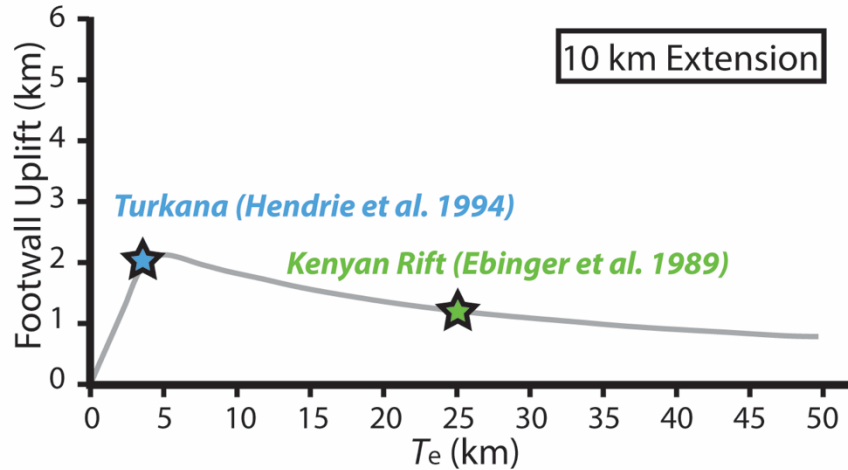


Figure 5.8. Plot of isostatic footwall uplift (km) versus effective elastic thickness (T_e) for 10 km of extension, illustrating how the flexural isostatic response to extension depends on lithospheric rigidity (adapted from Egan, 1992). Effective elastic thickness estimates for Turkana (Hendrie et al., 1994) and the Kenyan Rift (Ebinger et al., 1989) are plotted, illustrating that after 10 km of extension, the relatively low T_e of the Turkana lithosphere allows for twice the amount of isostatic footwall uplift than the thicker, cooler central Kenyan lithosphere.

While elevated crustal heat flow in Turkana is evidenced by a geothermal gradient of 4.2 °C/100 m from the Loperot-1 well (Morley, Stone, et al., 1999), similarly high geothermal gradients have been observed in multiple drill holes and wells from the central Kenya Rift (~3-6 °C/100 m) (Wheildon, Morgan, Williamson, Evans, & Swanberg, 1994), suggesting that a difference in heat flow alone cannot account for the greater degree of cooling observed in the Lapur Range. Thus, we believe that the larger amount of Miocene cooling in the Lapur Range is largely the result of increased isostatic uplift in the MRL footwall due to the low effective elastic thickness in the Turkana Depression.

5.5 Conclusions

ZHe, AFT and AHe analyses have been applied to 10 samples from the Lapur Range, an uplifted basement block in the footwall of the basin-bounding MRL Fault of the northern Turkana Basin. These data combined with inverse thermal history modelling reveal the following:

1. Basement rocks of the Lapur Range show a pronounced phase of Early Cretaceous (between 100-120 Ma) cooling consistent with ~5-7 km of denudation. Coeval subsidence in the adjacent Anza and South Sudan Rifts suggests that the northern

Turkana region may have acted as a basement high separating these rift systems and provided a local source for axially fed sediment.

2. The Lapur region transitioned to a period of reheating in the Late Cretaceous, between ca. 80 Ma and 20 Ma, coinciding with the deposition of ~500m of Upper Cretaceous-Eocene Lapur Sandstone and emplacement of up to 3.5 km of late Eocene-early Miocene Turkana Volcanics.
3. EARS-related faulting commenced in the Lapur area in the mid-Miocene at ~ 14 Ma, with the formation of the northern Turkana Basin bounding MRL Fault, coeval with rift initiation in other parts of eastern Turkana and southern Ethiopia.
4. Cooling rates associated with EARS-related flexural isostatic footwall uplift are significantly higher in the Turkana region compared to the Kenyan Rift, further south. This is best explained by the comparatively thin crust and high heat flow of the Turkana Depression, lowering the effective elastic thickness of its lithosphere. These crustal features are inherited from the Cretaceous-Paleogene rifting history of this region and amplified isostatic footwall uplift in response to extension.

References

- Abdelfettah, Y., Tiercelin, J., Tarits, P., Hautot, S., Maia, M., & Thuo, P. (2016). Subsurface structure and stratigraphy of the northwest end of the Turkana Basin, Northern Kenya Rift, as revealed by magnetotellurics and gravity joint inversion. *Journal of African Earth Sciences*, 119, 120–138. <https://doi.org/10.1016/j.jafrearsci.2016.03.008>
- Abebe, T., Balestrieri, M. L., & Bigazzi, G. (2010). The Central Main Ethiopian Rift is younger than 8 Ma: confirmation through apatite fission-track thermochronology. *Terra Nova*, 22(6), 470–476. <https://doi.org/10.1111/j.1365-3121.2010.00968.x>
- Arambourg, C., & Wolff, R. G. (1969). Nouvelles données paléontologique sur l'âge des "Gres du Lubur" (Turkana Grits) al'ouest du lac Rodolphe. *Comptes Rendus Société Géologique de France*, 9, 190–202.
- Ault, A. K., & Flowers, R. M. (2012). Is apatite U-Th zonation information necessary for accurate interpretation of apatite (U-Th)/He thermochronometry data? *Geochimica et Cosmochimica Acta*, 79, 60–78. <https://doi.org/10.1016/j.gca.2011.11.037>
- Baker, B. H., & Wohlenberg, J. (1971). Structure and Evolution of the Kenya Rift Valley. *Nature*, 229, 538–542.
- Balestrieri, M. L., Bonini, M., Corti, G., Sani, F., & Philippon, M. (2016). A refinement of the chronology of rift-related faulting in the Broadly Rifted Zone, southern Ethiopia, through apatite fission-track analysis. *Tectonophysics*, 671, 42–55. <https://doi.org/10.1016/j.tecto.2016.01.012>
- Barbarand, J., Carter, A., Wood, I., & Hurford, T. (2003). Compositional and structural control of fission-track annealing in apatite. *Chemical Geology*, 198(1–2), 107–137. [https://doi.org/10.1016/S0009-2541\(02\)00424-2](https://doi.org/10.1016/S0009-2541(02)00424-2)
- Bauer, F. U., Glasmacher, U. A., Ring, U., Karl, M., Schumann, A., & Nagudi, B. (2013). Tracing the exhumation history of the Rwenzori Mountains, Albertine Rift, Uganda, using low-temperature thermochronology. *Tectonophysics*, 599, 8–28. <https://doi.org/10.1016/j.tecto.2013.03.032>
- Bauer, F. U., Karl, M., Glasmacher, U. A., Nagudi, B., Schumann, A., & Mroszewski, L. (2012). Journal of African Earth Sciences The Rwenzori Mountains of western Uganda – Aspects on the evolution of their remarkable morphology within the Albertine Rift. *Journal of African Earth Sciences*, 73–74, 44–56. <https://doi.org/10.1016/j.jafrearsci.2012.07.001>
- Bechtel, T. D., Forsyth, D. W., & Swain, C. J. (1987). Mechanisms of isostatic compensation in the vicinity of the East African Rift, Kenya. *Geophysical Journal of the Royal Astronomical Society*, 90, 445–465. <https://doi.org/10.1111/j.1365-246X.1987.tb00734.x>
- Begg, G. C., Griffin, W. L., Natapov, L. M., O'Reilly, S. Y., Grand, S. P., O'Neill, C. J., ... Bowden, P. (2009). The lithospheric architecture of Africa: Seismic tomography, mantle petrology, and tectonic evolution. *Geosphere*, 5(1), 23–50. <https://doi.org/10.1130/GES00179.1>
- BEICIP. (1987). Geological Map of Kenya. Nairobi: Ministry of Energy and Regional Development.
- Bellieni, G., Justin Visentin, E., Zanettin, B., Piccirillo, E. M., Radicati di Brozolo, F., & Rita, F. (1981). Oligocene transitional tholeiitic magmatism in Northern turkana (Kenya): Comparison with the Coeval Ethiopian volcanism. *Bulletin Volcanologique*, 44(3), 411–427. <https://doi.org/10.1007/BF02600573>
- Benoit, M. H., Nyblade, A. A., & Pasyanos, M. E. (2006). Crustal thinning between the Ethiopian and East African plateaus from modeling Rayleigh wave dispersion. *Geophysical Research Letters*, 33(13). <https://doi.org/10.1029/2006GL025687>

- Biswas, S., Coutand, I., Grujic, D., Hager, C., Stockli, D., & Grasemann, B. (2007). Exhumation and uplift of the Shillong plateau and its influence on the eastern Himalayas: New constraints from apatite and zircon (U-Th-[Sm])/He and apatite fission track analyses. *Tectonics*, 26. <https://doi.org/10.1029/2007TC002125>
- Bonini, M., Corti, G., Innocenti, F., Manetti, P., Mazzarini, F., Abebe, T., & Pecskey, Z. (2005). Evolution of the Main Ethiopian Rift in the frame of Afar and Kenya rifts propagation. *Tectonics*, 24(1), n/a-n/a. <https://doi.org/10.1029/2004TC001680>
- Boone, S. C., Seiler, C., Reid, A. J., Kohn, B., & Gleadow, A. (2016). An Upper Cretaceous paleo-aquifer system in the Eromanga Basin of the central Gawler Craton, South Australia: evidence from apatite fission track thermochronology. *Australian Journal of Earth Sciences*, 63(3), 315–331. <https://doi.org/10.1080/08120099.2016.1199050>
- Bosworth, W. (1992). Mesozoic and early Tertiary rift tectonics in East Africa. *Tectonophysics*, 209(1–4), 115–137. [https://doi.org/10.1016/0040-1951\(92\)90014-W](https://doi.org/10.1016/0040-1951(92)90014-W)
- Bosworth, W., & Morley, C. K. (1994). Structural and stratigraphic evolution of the Anza rift, Kenya. *Tectonophysics*, 236(1–4), 93–115. [https://doi.org/10.1016/0040-1951\(94\)90171-6](https://doi.org/10.1016/0040-1951(94)90171-6)
- Braun, J., Stippich, C., & Glasmacher, U. A. (2016). The effect of variability in rock thermal conductivity on exhumation rate estimates from thermochronological data. *Tectonophysics*, 690, 288–297. <https://doi.org/10.1016/j.tecto.2016.09.027>
- Brown, R. W., Beucher, R., Roper, S., Persano, C., Stuart, F., & Fitzgerald, P. (2013). Natural age dispersion arising from the analysis of broken crystals. Part I: Theoretical basis and implications for the apatite (U-Th)/He thermochronometer. *Geochimica et Cosmochimica Acta*, 122(120), 478–497. <https://doi.org/10.1016/j.gca.2013.05.041>
- Brown, R. W., Rust, D. J., Summerfield, M. A., Gleadow, A. J. W., & De Wit, M. C. J. (1990). An Early Cretaceous Phase of Accelerated Erosion on the South-Western Margin of Africa: Evidence from Apatite Fission Track Analysis and the Offshore Sedimentary Record. *International Journal of Radiation Applications and Instrumentation, Part D*, 17(3), 339–350.
- Browne, S. E., & Fairhead, J. D. (1983). Gravity study of the Central African Rift system: A model of continental disruption 1. The Ngaoundere and Abu Gabra Rifts. *Tectonophysics*, 94(1–4), 187–203. [https://doi.org/10.1016/0040-1951\(83\)90016-1](https://doi.org/10.1016/0040-1951(83)90016-1)
- Brune, S., Corti, G., & Ranalli, G. (2017). Controls of inherited lithospheric heterogeneity on rift linkage: Numerical and analog models of interaction between the Kenyan and Ethiopian rifts across the Turkana depression. *Tectonics*, 1–30. <https://doi.org/10.1002/2017tc004739>
- Chorowicz, J. (2005). The East African rift system. *Journal of African Earth Sciences*, 43, 379–410. <https://doi.org/10.1038/215578a0>
- Danišik, M., McInnes, B. I. A., Kirkland, C. L., McDonald, B. J., Evans, N. J., & Becker, T. (2017). Seeing is believing: Visualization of He distribution in zircon and implications for thermal history reconstruction on single crystals. *Science Advances*, 3, 1–9.
- Donelick, R. A., O'Sullivan, P. B., & Ketcham, R. A. (2005). Apatite Fission-Track Analysis. *Reviews in Mineralogy and Geochemistry*, 58(1), 49–94. <https://doi.org/10.2138/rmg.2005.58.3>
- Dunkelman, T. J., Rosendahl, B. R., & Karson, J. A. (1989). Structure and stratigraphy of the Turkana rift from seismic reflection data. *Journal of African Earth Sciences*, 8(2–4), 489–510. [https://doi.org/10.1016/S0899-5362\(89\)80041-7](https://doi.org/10.1016/S0899-5362(89)80041-7)
- Ebinger, C. J., Deino, A. L., Drake, R. E., & Tesha, A. L. (1989). Chronology of Volcanism and Rift Basin Propagation: Rungwe Volcanic Province, East Africa. *Journal of Geophysical Research*, 94(B11), 15,785–15,803.
- Ebinger, C. J., & Sleep, N. H. (1998). Cenozoic magnetism throughout east Africa resulting

- from impact of a single plume. *Nature*, 395, 788–791.
- Ebinger, C. J., Yemane, T., Harding, D., Tesfaye, S., Kelley, S., & Rex, D. C. (2000). Rift deflection, migration, and propagation: Linkage of the Ethiopian and Eastern rifts, Africa. *GSA Bulletin*, 112(2), 163–176.
- Ebinger, C. J., Yemane, T., Woldegabriel, G., Aronson, J. L., & Walter, R. C. (1993). Late Eocene-Recent volcanism and faulting in the southern main Ethiopian rift. *Journal of the Geological Society*, 150(1), 99–108. <https://doi.org/10.1144/gsjgs.150.1.0099>
- Egan, S. S. (1992). The flexural isostatic response of the lithosphere to extensional tectonics. *Tectonophysics*, 202(2–4), 291–308. [https://doi.org/10.1016/0040-1951\(92\)90115-M](https://doi.org/10.1016/0040-1951(92)90115-M)
- Fairhead, J. D. (1988). Mesozoic plate tectonic reconstructions of the central South Atlantic Ocean: The role of the West and Central African rift system. *Elsevier Science Publishers B.V., Amsterdam*, 155, 181–191. [https://doi.org/10.1016/0040-1951\(88\)90265-X](https://doi.org/10.1016/0040-1951(88)90265-X)
- Farley, K. A. (2000). Helium diffusion from apatite: General behavior as illustrated by Durango fluorapatite. *Journal of Geophysical Research*, 105(B2), 2903–2914.
- Farley, K. A., & Stockli, D. F. (2002). (U-Th)/He Dating of Phosphates: Apatite, Monazite, and Xenotime. *Reviews in Mineralogy and Geochemistry*, 48(1), 559–577. <https://doi.org/10.2138/rmg.2002.48.15>
- Farley, K. A., Wolf, R. A., & Silver, L. T. (1996). The effects of long alpha-stopping distances on (U-Th)/He ages. *Geochimica et Cosmochimica Acta*, 60(21), 4223–4229. [https://doi.org/10.1016/S0016-7037\(96\)00193-7](https://doi.org/10.1016/S0016-7037(96)00193-7)
- Feibel, C. S., Brown, F. H., & McDougall, I. (1989). Stratigraphic context of fossil hominids from the Omo group deposits: Northern Turkana Basin, Kenya and Ethiopia. *American Journal of Physical Anthropology*, 78(4), 595–622. <https://doi.org/10.1002/ajpa.1330780412>
- Feinstein, S., Eyal, M., Kohn, B. P., Steckler, M. S., Ibrahim, K. M., Moh'd, B. K., & Tian, Y. (2013). Uplift and denudation history of the eastern Dead Sea rift flank, SW Jordan: Evidence from apatite fission track thermochronometry. *Tectonics*, 32, 1513–1528. <https://doi.org/10.1002/tect.20082>
- Fleischer, R., Price, P., & Walker, R. (1975). *Nuclear Tracks in Solids: principles and applications*. Berkeley: University of California Press.
- Flowers, R. M., Ketcham, R. A., Shuster, D. L., & Farley, K. A. (2009). Apatite (U-Th)/He thermochronometry using a radiation damage accumulation and annealing model. *Geochimica et Cosmochimica Acta*, 73(8), 2347–2365. <https://doi.org/10.1016/j.gca.2009.01.015>
- Foster, D. A., & Gleadow, A. J. W. (1992). The morphotectonic evolution of rift-margin mountains in central Kenya: Constraints from apatite fission-track thermochronology. *Earth and Planetary Science Letters*, 113(1–2), 157–171. [https://doi.org/10.1016/0012-821X\(92\)90217-J](https://doi.org/10.1016/0012-821X(92)90217-J)
- Foster, D. A., & Gleadow, A. J. W. (1993). Episodic Denudation in East Africa: A Legacy of Intracontinental Tectonism. *Geophysical Research Letters*, 20(21), 2395–2398.
- Foster, D. A., & Gleadow, J. W. (1996). Structural framework and denudation history of the flanks of the Kenya and Anza Rifts, East Africa. *Tectonics*, 15(2), 258–271. <https://doi.org/10.1029/95TC02744>
- Galbraith, R. F. (2005). *Statistics for fission track analysis*. CRC Press.
- Gallagher, K. (1995). Evolving temperature histories from apatite fission-track data. *Earth and Planetary Science Letters*, 136(3–4), 421–435.
- Gallagher, K. (2012). Transdimensional inverse thermal history modeling for quantitative thermochronology. *Journal of Geophysical Research: Solid Earth*, 117. <https://doi.org/10.1029/2011JB008825>

- Gallagher, K., & Brown, R. (1997). The onshore record of passive margin evolution. *Journal of the Geological Society, London*, 154, 451–457.
<https://doi.org/10.1144/gsjgs.154.3.0451>
- Gallagher, K., & Brown, R. (1999). Denudation and uplift at passive margins: the record on the Atlantic Margin of southern Africa. *Philosophical Transactions of the Royal Society London A: Mathematical, Physical and Engineering Sciences*, 357, 835–859.
<https://doi.org/10.1098/rsta.1999.0354>
- Gautheron, C., Tassan-Got, L., Barbarand, J., & Pagel, M. (2009). Effect of alpha-damage annealing on apatite (U-Th)/He thermochronology. *Chemical Geology*, 266, 157–170.
<https://doi.org/10.1016/j.chemgeo.2009.06.001>
- George, R. M., & Rogers, N. W. (2002). Plume dynamics beneath the African plate inferred from the geochemistry of the Tertiary basalts of southern Ethiopia. *Contribution to Mineralogy and Petrology*, 144, 286–304. <https://doi.org/10.1007/s00410-002-0396-z>
- George, R., Rogers, N., & Kelley, S. (1998). Earliest magnetism in Ethiopia: Evidence for two mantle plumes in one flood basalt province. *Geology*, 26(10), 923–926.
- Gleadow, A. J. W., Gleadow, S. J., Belton, D. X., Kohn, B. P., Krochmal, M. S., & Brown, R. W. (2009). Coincidence mapping – a key strategy for the automatic counting of fission tracks in natural minerals. *Geological Society of London Special Publications, Thermochronological Methods: From Palaeotemperature Constraints to Landscape Evolution Models.*, 324, 25–36. <https://doi.org/10.1144/SP324.2>
- Gleadow, A., Harrison, M., Kohn, B., Lugo-zazueta, R., & Phillips, D. (2015). The Fish Canyon Tuff: A new look at an old low-temperature thermochronology standard. *Earth and Planetary Science Letters*, 424, 95–108. <https://doi.org/10.1016/j.epsl.2015.05.003>
- Gleadow, A. J. W. (1980). Fission track age of the KBS Tuff and associated hominid remains in northern Kenya. *Nature*, 284, 225–230. <https://doi.org/10.1038/284225a0>
- Gleadow, A. J. W., & Brown, R. W. (2000). Fission track thermochronology and the long-term denudational response to tectonics. *Geomorphology and Global Tectonics*, (May), 57–75. Retrieved from
y:%5CARCHIV_GEOWATT%5CPaper%5CPaper_archiviert%5CGleadow_1999_Wiley_FissionTrackAnalysis.pdf
- Gleadow, A. J. W., & Duddy, I. R. (1981). A Natural Long-Term Track Annealing Experiment for Apatite. *Nuclear Tracks*, 5(1–2), 169–174.
- GMRD-Sudan. (1981). Geological Map of the Sudan. Khartoum: Geological & Mineral Resources Department (G. M. R. D.) Sudan.
- Greene, L. C., Richards, D. R., & Johnson, R. A. (1991). Crustal structure and tectonic evolution of the anza rift, northern Kenya. *Tectonophysics*, 197(2–4), 203–211.
[https://doi.org/10.1016/0040-1951\(91\)90041-P](https://doi.org/10.1016/0040-1951(91)90041-P)
- Guenther, W. R., Reiners, P. W., Ketcham, R. A., Nasdala, L., & Giester, G. (2013). Helium diffusion in natural zircon: radiation damage, anisotropy, and the interpretation of zircon (U-Th)/He thermochronology. *American Journal of Science*, 313(3), 145–198.
<https://doi.org/10.2475/03.2013.01>
- Guenther, W. R., Reiners, P. W., & Tian, Y. (2014). Interpreting date-eU correlations in zircon (U-Th)/He datasets: A case study from the Longmen Shan, China. *Earth and Planetary Science Letters*, 403(OCTOBER), 328–339.
<https://doi.org/10.1016/j.epsl.2014.06.050>
- Guiraud, R., & Bosworth, W. (1997). Senonian basin inversion and rejuvenation of rifting in Africa and Arabia: synthesis and implications to plate-scale tectonics. *Tectonophysics*, 282, 39–82.
- Guiraud, R., & Maurin, J.-C. (1992). Early Cretaceous rifts of Western and Central Africa: an overview. *Tectonophysics*, 213, 153–168.

- Hansen, S. E., Nyblade, A. A., & Benoit, M. H. (2012). Mantle structure beneath Africa and Arabia from adaptively parameterized P-wave tomography: Implications for the origin of Cenozoic Afro-Arabian tectonism. *Earth and Planetary Science Letters*, 319–320, 23–34. <https://doi.org/10.1016/j.epsl.2011.12.023>
- Hasebe, N., Barbarand, J., Jarvis, K., Carter, A., & Hurford, A. J. (2004). Apatite fission-track chronometry using laser ablation ICP-MS. *Chemical Geology*, 207(3), 134–145.
- Hautot, S., Tarits, P., Whaler, K., Le Gall, B., Tiercelin, J.-J., & Le Turdu, C. (2000). Deep structure of the Baringo Rift Basin (central Kenya) from three-dimensional magnetotelluric imaging: Implications for rift evolution. *Journal of Geophysical Research*, 105(B10), 23,493–23,518. <https://doi.org/10.1029/2000JB900213>
- Hendrie, D. B., Kusznir, N. J., Morley, C. K., & Ebinger, C. J. (1994). Cenozoic extension in northern Kenya: a quantitative model of rift basin development in the Turkana region. *Tectonophysics*, 236(1–4), 409–438. [https://doi.org/10.1016/0040-1951\(94\)90187-2](https://doi.org/10.1016/0040-1951(94)90187-2)
- Henry, W. J., Mechie, J., Maguire, P. K. H., Khan, M. A., Prodehl, C., Keller, G. R., & Patel, J. (1990). A Seismic Investigation of the Kenya Rift Valley. *Geophysical Journal International*, 100(1), 107–130. <https://doi.org/10.1111/j.1365-246X.1990.tb04572.x>
- Hourigan, J. K., Reiners, P. W., & Brandon, M. T. (2005). U-Th zonation-dependent alpha-ejection in (U-Th)/He chronometry. *Geochimica et Cosmochimica Acta*, 69(13), 3349–3365. <https://doi.org/10.1016/j.gca.2005.01.024>
- Ibrahim, A. E., Ebinger, C. J., & Fairhead, J. D. (1991). Interpretation of the Central African Rift system in Sudan based on new gravity and aeromagnetic data. *EOS*, 462.
- Johnson, J. E., Flowers, R. M., Baird, G. B., & Mahan, K. H. (2017). “Inverted” zircon and apatite (U–Th)/He dates from the Front Range, Colorado: High-damage zircon as a low-temperature (<50 °C) thermochronometer. *Earth and Planetary Science Letters*, 466, 80–90. <https://doi.org/10.1016/j.epsl.2017.03.002>
- Kasanzu, C. H. (2016). Apatite fission track and (U-Th)/He thermochronology from the Archean Tanzania craton: Contributions to cooling histories of Tanzanian basement rocks. *Geoscience Frontiers*, 1–9. <https://doi.org/10.1016/j.gsf.2016.09.007>
- Kasanzu, C. H., Linol, B., de Wit, M. J., Brown, R., Persano, C., & Stuart, F. M. (2016). From source to sink in central Gondwana: Exhumation of the Precambrian basement rocks of Tanzania and sediment accumulation in the adjacent Congo basin. *Tectonics*, 35(9), 2034–2051. <https://doi.org/10.1002/2016TC004147>
- Kazmin, V. (1972). Geological Map of Ethiopia. Addis Ababa: Geological Survey of Ethiopia, Ministry of Mines.
- Ketcham, R. A. (2005). Forward and Inverse Modeling of Low- Temperature Thermochronometry Data. *Reviews in Mineralogy and Geochemistry*, 58, 275–314. <https://doi.org/10.2138/rmg.2005.58.11>
- Ketcham, R. A., Carter, A., Donelick, R. A., Barbarand, J., & Hurford, A. J. (2007a). Improved measurement of fission-track annealing in apatite using c-axis projection. *American Mineralogist*, 92(5–6), 789–798. <https://doi.org/10.2138/am.2007.2280>
- Ketcham, R. A., Carter, A., Donelick, R. A., Barbarand, J., & Hurford, A. J. (2007b). Improved modeling of fission-track annealing in apatite. *American Mineralogist*, 92, 799–810. <https://doi.org/10.2138/am.2007.2281>
- Ketcham, R. A., Guenther, W. R., & Reiners, P. W. (2013). Geometric analysis of radiation damage connectivity in zircon, and its implications for helium diffusion. *American Mineralogist*, 98(2–3), 350–360. <https://doi.org/10.2138/am.2013.4249>
- Kohn, B. P., Gleadow, A. J. W., Brown, R. W., Gallagher, K., Lorencak, M., & Noble, W. P. (2005). Visualizing Thermotectonic and Denudational Histories Using Apatite Fission Track Thermochronology. *Reviews in Mineralogy and Geochemistry*, 58, 527–565. <https://doi.org/10.2138/rmg.2005.58.20>

- Kreuser, T. (1995). Rift to drift evolution in Permian-Jurassic basins of East Africa. *Geological Society, London, Special Publication*, 80(1), 297–315.
- Kröner, a, & Stern, R. J. (2005). Pan-African Orogeny. *Encyclopedia Of Geology*, 1(2004), 1–12. [https://doi.org/10.1016/S1342-937X\(05\)70162-3](https://doi.org/10.1016/S1342-937X(05)70162-3)
- Kusznir, N. J., & Ziegler, P. A. (1992). The mechanics of continental extension and sedimentary basin formation : A simple-shear / pure-shear flexural cantilever model, 215, 117–131.
- Laslett, G. M., Green, P. F., Duddy, I. R., & Gleadow, A. J. W. (1987). Thermal annealing of fission tracks in apatite 2 . A quantitative analysis. *Chemical Geology (Isotope Geoscience Section)*, 65, 1–13. [https://doi.org/10.1016/0168-9622\(87\)90057-1](https://doi.org/10.1016/0168-9622(87)90057-1)
- Latin, D., Norry, M. J., & Tarzey, R. J. E. (1993). Magmatism in the gregory rift, East Africa: Evidence for melt generation by a plume. *Journal of Petrology*, 34(5), 1007–1027. <https://doi.org/10.1093/petrology/34.5.1007>
- Lehto, T., Westerhof, A. B., Lehtonen, M. I., Manninen, T., Mäkitie, H., Virransalo, P., ... Elepu, D. (2014). Geological Map of Uganda, Scale 1:1 000 000. Espoo, Finland: Geological Survey of Finland.
- Lippard, S. J. (1972). *Stratigraphy and structure of the Elgeyo escarpment southern Kamasia Hills and adjoining regions, Rift Valley Province, Kenya*. Royal Holloway, University of London.
- Lippolt, H. J., Leitz, M., Wernicke, R. S., & Hagedorn, B. (1994). (Uranium + thorium)/helium dating of apatite: experience with samples from different geochemical environments. *Chemical Geology*, 112(1–2), 179–191. [https://doi.org/10.1016/0009-2541\(94\)90113-9](https://doi.org/10.1016/0009-2541(94)90113-9)
- Lisker, F., Brown, R., & Fabel, D. (2003). Denudational and thermal history along a transect across the Lambert Graben , northern Prince Charles Mountains , Antarctica , derived from apatite fission track thermochronology, 22(5). <https://doi.org/10.1029/2002TC001477>
- Łuszczak, K., Persano, C., Braun, J., & Stuart, F. M. (2017). How local crustal thermal properties influence the amount of denudation derived from low-temperature thermochronometry. *Geology*, 45(9), 779–782. <https://doi.org/10.1130/G39036.1>
- Mackintosh, V., Kohn, B., Gleadow, A., & Tian, Y. (2017). Phanerozoic Morphotectonic Evolution of the Zimbabwe Craton: Unexpected Outcomes From a Multiple Low-Temperature Thermochronology Study. *Tectonics*, 36, 1–24. <https://doi.org/10.1002/2017TC004703>
- Maguire, P. K. H., Swain, C. J., Masotti, R., & Khan, M. A. (1994). A crustal and uppermost mantle cross-sectional model of the Kenya Rift derived from seismic and gravity data. *Tectonophysics*, 236(1–4), 217–249. [https://doi.org/10.1016/0040-1951\(94\)90178-3](https://doi.org/10.1016/0040-1951(94)90178-3)
- Majer-Kielbaska, D. A. (2015). *Low temperature thermochronology of the northernmost Mozambique Belt, Kenya*. University of Melbourne.
- Mbede, E., Andriessen, P., Delvaux, D., & van der Beek, P. (1998). Denudation history of the Malawi and Rukwa Rift flanks (East African Rift System) from apatite fission track thermochronology. *Journal of African Earth Sciences*, 26(3), 363–385.
- McDougall, I., & Brown, F. H. (2009). Timing of volcanism and evolution of the northern Kenya Rift. *Geological Magazine*, 146(1), 34–37. <https://doi.org/10.1017/S0016756808005347>
- McDowell, F. W., McIntosh, W. C., & Farley, K. A. (2005). A precise ^{40}Ar - ^{39}Ar reference age for the Durango apatite (U-Th)/He and fission track dating standard. *Chemical Geology*, 214(3), 249–263.
- Mechie, J., Keller, G. R., Prodehl, C., Gaciri, S., Braile, L. W., Mooney, W. D., ... Sandmeier, K.-J. (1994). Crustal structure beneath the Kenya Rift from axial profile

- data. *Tectonophysics*, 236, 179–200.
- Morley, C. K., Bosworth, W., Day, R. A., Lauck, R., Bosher, R., Stone, D. M., ... Bassett, N. (1999). Geology and Geophysics of the Anza Graben. In C. K. Morley (Ed.), *Geoscience of Rift Systems-Evolution of East Africa: AAPG Studies in Geology No. 44* (pp. 67–90).
- Morley, C. K., Ngenoh, D. K., & Ego, J. K. (1999). Introduction to the East African Rift System. In C. K. Morley (Ed.), *Geoscience of Rift Systems-Evolution of East Africa: AAPG Studies in Geology No. 44* (pp. 1–18).
- Morley, C. K., Stone, D. M., Harper, R. M., & Wigger, S. T. (1999). Geology and Geophysics of the Western Turkana Basins, Kenya. In C. K. Morley (Ed.), *Geoscience of Rift Systems—Evolution of East Africa: AAPG Studies in Geology No. 44* (pp. 19–54).
- Morley, C. K., Wescott, W. a., Stone, D. M., Harper, R. M., Wigger, S. T., & Karanja, F. M. (1992). Tectonic evolution of the northern Kenyan Rift. *Journal of the Geological Society*, 149(3), 333–348. <https://doi.org/10.1144/gsjgs.149.3.0333>
- Mugisha, F., Ebinger, C. J., Strecker, M., & Pope, D. (1997). Two-stage rifting in the Kenya rift: implications for half-graben models. *Tectonophysics*, 278, 63–81.
- Nasdala, L., Reiners, P. W., Garver, J. I., Kennedy, A. K., Stern, R. A., Balan, E., & Wirth, R. (2004). Incomplete retention of radiation damage in zircon from Sri Lanka. *American Mineralogist*, 89(1), 219–231. <https://doi.org/10.2138/am-2004-0126>
- Noble, W. P. (1997). *Post Pan African Tectonic Evolution of Eastern Africa: An Apatite Fission Track Study*. La Trobe University.
- Noble, W. P., Foster, D. A., & Gleadow, A. J. W. (1997). The post-Pan-African thermal and extensional history of crystalline basement rocks in eastern Tanzania. *Tectonophysics*, 275(4), 331–350. [https://doi.org/10.1016/S0040-1951\(97\)00026-7](https://doi.org/10.1016/S0040-1951(97)00026-7)
- Nyblade, A. A. (1990). Terrestrial heat flow in east and southern Africa. *Journal of Geophysical Research: Solid Earth*, 95(B11), 17371–17384.
- O'Connor, P. M., Sertich, J. J. W., & Manthi, F. K. (2011). A pterodactyloid pterosaur from the Upper Cretaceous Lapurr sandstone, West Turkana, Kenya. *Anais Da Academia Brasileira de Ciencias*, 83(1), 309–315. <https://doi.org/10.1590/S0001-37652011000100019>
- Orme, D. A., Reiners, P. W., Hourigan, J. K., & Carrapa, B. (2015). Effects of inherited cores and magmatic overgrowths on zircon (U-Th)/He ages and age-eU trends from Greater Himalayan sequence rocks, Mount Everest region, Tibet. *Geochemistry Geophysics Geosystems*, 16, 2499–2507. <https://doi.org/10.1002/2015GC005818>
- Paton, C., Hellstrom, J., Paul, B., Woodhead, J., & Herget, J. (2011). Iolite: Freeware for the visualisation and processing of mass spectrometric data. *Journal of Analytical Atomic Spectrometry*, 26, 2508–2518. <https://doi.org/10.1039/c1ja10172b>
- Philippon, M., Corti, G., Sani, F., Bonini, M., Balestrieri, M., Molin, P., ... Cloetingh, S. (2014). Evolution, distribution, and characteristics of rifting in southern Ethiopia. *Tectonics*, 33, 1–24. <https://doi.org/10.1002/2013TC003430>
- Pik, R., Marty, B., Carignan, J., Yirgu, G., & Ayalew, T. (2008). Timing of East African Rift development in southern Ethiopia : Implication for mantle plume activity and evolution of topography. *Geology*, 36(2), 167–170. <https://doi.org/10.1130/G24233A.1>
- Reeves, C. V., Karanja, F. M., & MacLeod, I. N. (1987). Geophysical evidence for a failed Jurassic rift and triple junction in Kenya. *Earth and Planetary Science Letters*, 81(2–3), 299–311. [https://doi.org/10.1016/0012-821X\(87\)90166-X](https://doi.org/10.1016/0012-821X(87)90166-X)
- Reiners, P. W. (2005). Zircon (U-Th)/He Thermochronometry. *Reviews in Mineralogy and Geochemistry*, 58, 151–179. <https://doi.org/10.2138/rmg.2005.58.6>
- Reiners, P. W., & Farley, K. A. (2001). Influence of crystal size on apatite (U-Th)/He thermochronology: an example from the Bighorn Mountains, Wyoming. *Earth and*

- Planetary Science Letters*, 188, 413–420.
- Schull, T. J. (1988). Rift Basins of Interior Sudan: Petroleum Exploration and Discovery, 72(10), 1128–1142.
- Seiler, C., Fletcher, J. M., Kohn, B. P., Gleadow, A. J. W., & Raza, A. (2011). Low - temperature thermochronology of northern Baja California , Mexico : Decoupled slip - exhumation gradients and delayed onset of oblique rifting across the Gulf of California. *Tectonics*, 30(1). <https://doi.org/10.1029/2009TC002649>
- Seiler, C., Kohn, B., & Gleadow, A. (2014). Apatite fission track analysis by LA-ICP-MS: An evaluation of the absolute dating approach. In *International Conference on Thermochronology*.
- Shuster, D. L., & Farley, K. A. (2009). The influence of artificial radiation damage and thermal annealing on helium diffusion kinetics in apatite. *Geochimica et Cosmochimica Acta*, 73(1), 183–196. <https://doi.org/10.1016/j.gca.2008.10.013>
- Shuster, D. L., Flowers, R. M., & Farley, K. A. (2006). The influence of natural radiation damage on helium diffusion kinetics in apatite. *Earth and Planetary Science Letters*, 249(3–4), 148–161. <https://doi.org/10.1016/j.epsl.2006.07.028>
- Sippel, J., Meeßen, C., Cacace, M., Mechie, J., Fishwick, S., Heine, C., Scheck-Wenderoth, M., & Strecker, M. R. (2017). The Kenya rift revisited: insights into lithospheric strength through data-driven 3-D gravity and thermal modelling. *Solid Earth*, 8, 45–81. <https://doi.org/10.5194/se-8-45-2017>
- Smith, M., & Mosley, P. (1993). Crustal Heterogeneity and Basement Influence on the Development of the Kenya Rift, East Africa. *Tectonics*, 12(2), 591–606.
- Spiegel, C., Kohn, B., Belton, D., Berner, Z., & Gleadow, A. (2009). Apatite (U-Th-Sm)/He thermochronology of rapidly cooled samples: The effect of He implantation. *Earth and Planetary Science Letters*, 285(1–2), 105–114. <https://doi.org/10.1016/j.epsl.2009.05.045>
- Spiegel, C., Kohn, B. P., Belton, D. X., & Gleadow, A. J. W. (2007). Morphotectonic evolution of the central Kenya rift flanks: Implications for late Cenozoic environmental change in East Africa. *Geology*, 35(5), 427–430. <https://doi.org/10.1130/G23108A.1>
- Stockli, D. F. (2005). Application of low-temperature thermochronology to extensional tectonic settings. *Reviews in Mineralogy and Geochemistry*, 58(1), 411–448.
- Sullivan, P. B. O. Y., Mitchell, M. M., Sullivan, A. J. O., Kohn, B. P., & Gleadow, A. J. W. (2000). Thermotectonic history of the Bassian Rise , Australia : implications for the breakup of eastern Gondwana along Australia ' s southeastern margins, 182, 31–47.
- Tagami, T., Farley, K. A., & Stockli, D. F. (2003). (U-Th)/He geochronology of single zircon grains of known Tertiary eruption age. *Earth and Planetary Science Letters*, 207(1–4), 57–67. [https://doi.org/10.1016/S0012-821X\(02\)01144-5](https://doi.org/10.1016/S0012-821X(02)01144-5)
- Tiercelin, J. J., Potdevin, J. L., Thuo, P. K., Abdelfettah, Y., Schuster, M., Bourquin, S., ... Ruffet, G. (2012). Stratigraphy, sedimentology and diagenetic evolution of the Lapur Sandstone in northern Kenya: Implications for oil exploration of the Meso-Cenozoic Turkana depression. *Journal of African Earth Sciences*, 71–72, 43–79. <https://doi.org/10.1016/j.jafrearsci.2012.06.007>
- Torres Acosta, V., Bande, A., Sobel, E. R., Parra, M., Schildgen, T. F., Stuart, F., & Strecker, M. R. (2015). Cenozoic extension in the Kenya Rift from low-temperature thermochronology: Links to diachronous spatiotemporal evolution of rifting in East Africa. *Tectonics*, 34(12), 2367–2386. <https://doi.org/10.1002/2015TC003949>
- Tripathy-Lang, A., Fox, M., & Shuster, D. L. (2015). Zircon 4He/3He thermochronometry. *Geochimica et Cosmochimica Acta*, 166, 1–14. <https://doi.org/10.1016/j.gca.2015.05.027>
- Unrug, R. (1997). Rodinia to Gondwana: The geodynamic map of Gondwana supercontinent

- assembly. *GSA Today*, 7(1). <https://doi.org/10.2460/javma.242.12.1666>
- van der Beek, P., Mbede, E., Andriessen, P., & Delvaux, D. (1998). Denudation history of the Malawi and Rukwa Rift flanks (East African Rift system) from apatite fission track thermochronology. *Journal of African Earth Sciences*, 26(3), 363–385. [https://doi.org/10.1016/S0899-5362\(98\)00021-9](https://doi.org/10.1016/S0899-5362(98)00021-9)
- Vermeesch, P. (2009). RadialPlotter: A Java application for fission track, luminescence and other radial plots. *Radiation Measurements*, 44(4), 409–410.
- Vermeesch, P. (2010). HelioPlot, and the treatment of overdispersed (U-Th-Sm)/He data. *Chemical Geology*, 271(3–4), 108–111. <https://doi.org/10.1016/j.chemgeo.2010.01.002>
- Wagner, G., & Van den Haute, P. (1992). *Fission-Track Dating*. Dordrecht: Kluwer Academic Publishers.
- Wagner, M., Altherr, R., & Van Den Haute, P. (1992). Apatite fission-track analysis of Kenyan basement rocks: constraints on the thermotectonic evolution of the Kenya dome. A reconnaissance study. *Tectonophysics*, 204(1–2), 93–110. [https://doi.org/10.1016/0040-1951\(92\)90272-8](https://doi.org/10.1016/0040-1951(92)90272-8)
- Walcott, R. I. (1970). Flexural Rigidity, Thickness, and Viscosity of the Lithosphere. *Journal of Geophysical Research*, 75(20), 3941–3954. <https://doi.org/10.1029/JB075i020p03941>
- Walsh, J., & Dodson, R. G. (1969). Geology of Northern Turkana, Degree Sheets 1, 2, 9, and 10. Geological Survey of Kenya.
- Wescott, W. A., Wigger, S. T., Stone, D. M., & Morley, C. K. (1999). Geology and Geophysics of the Lotikipi Plain. *Geoscience of Rift Systems-Evolution of East Africa*, (44), 55–66.
- Wheildon, J., Morgan, P., Williamson, K. H., Evans, T. R., & Swanberg, C. A. (1994). Heat flow in the Kenya rift zone. *Tectonophysics*, 236(1–4), 131–149. [https://doi.org/10.1016/0040-1951\(94\)90173-2](https://doi.org/10.1016/0040-1951(94)90173-2)
- Wichura, H., Bousquet, R., Oberhänsli, R., Strecker, M. R., & Trauth, M. H. (2010). Evidence for middle Miocene uplift of the East African Plateau. *Geology*, 38(6), 543–546. <https://doi.org/10.1130/G31022.1>
- Wolf, R. A., Farley, K. A., & Kass, D. M. (1998). Modeling of the temperature sensitivity of the apatite (U – Th)/He thermochronometer. *Chemical Geology*, 148, 105–114.
- WoldeGabriel, G., Aronson, J. L., & Walter, R. C. (1990). Geochronology and rift basin development in the central sector of the Main Ethiopian Rift. *Geological Society of America Bulletin*, 102, 439–485. [https://doi.org/10.1130/0016-7606\(1990\)102<0439](https://doi.org/10.1130/0016-7606(1990)102<0439)
- Wolfe, M. R., & Stockli, D. F. (2010). Zircon (U-Th)/He thermochronometry in the KTB drill hole, Germany, and its implications for bulk He diffusion kinetics in zircon. *Earth and Planetary Science Letters*, 295(1–2), 69–82. <https://doi.org/10.1016/j.epsl.2010.03.025>
- Wolfenden, E., Ebinger, C., Yirgu, G., Deino, A., & Ayalew, D. (2004). Evolution of the northern Main Ethiopian rift: birth of a triple junction. *Earth and Planetary Science Letters*, 224(1–2), 213–228. <https://doi.org/10.1016/j.epsl.2004.04.022>
- Zanettin, B., Justin Visentin, E., Bellieni, G., Piccirillo, E. M., & Francesca, R. (1983). The volcanism of the North Turkana Basin, Kenya: age, succession and structural evolution. *Bulletin Des Centres de Recherches Exploration-Production Elf-Aquitaine*, 7, 249–255.

Concluding Remarks

The determination of spatio-temporal chronologies of extensional deformation and landscape evolution are critical to enhancing our understanding of the development of the Turkana Depression in relation to multiple phases of superimposed rifting since the Cretaceous. The preceding four chapters have explored the tectono-thermal evolution of the northern Kenyan and southern Ethiopian upper crust through the application of multiple low-temperature thermochronometers (ZHe, AFT and AHe analyses) and thermal history modelling. Spatial patterns of Cretaceous-Recent upper crustal cooling of the Turkana Depression revealed by these data (Figure 6.1) provide important insights into extensional strain localisation and propagation, the development of topography, and paleostress field orientation in the region.

6.1 Did the Anza and South Sudan Rifts extend through Turkana?

Constraining the nature and extent of Cretaceous-early Paleogene upper crustal deformation in the Turkana Depression related to the development of the surrounding Anza and South Sudan Rift systems is crucial to understanding the lithospheric evolution of the region and its influence on subsequent EARS extension. Many authors, largely based on the compatible geometries, coeval evolution and similar gravity signals of these rift [*Schull, 1988; Ibrahim et al., 1991; Bosworth, 1992; Ebinger and Ibrahim, 1994*], suggest that the Anza and South Sudan rift systems once linked within the Turkana region [e.g. *Ebinger et al., 2000*]. This assumption is often invoked to explain the diffuse distribution of extensional deformation in the Turkana Depression [**Chapter 4**; *Brune et al., 2017*], as well as its anomalously thin crust [*Mariita and Keller, 2007*], low topography [e.g. *Benoit et al., 2006; Fishwick and Bastow, 2011*] and subdued rift morphology [e.g. *Morley et al., 1999a*]. However, the location and geometry of this proposed connection remain enigmatic due to a general data gap in eastern

South Sudan and the obscuring effect of subsequent Paleogene-Recent tectonism and volcanic activity in northern Kenya [*Guiraud and Bosworth, 1997*].

Low-temperature thermochronology from the Turkana Depression crystalline basement presented in this study provides previously unavailable insights into the tectono-thermal evolution of this region during the period of proposed Anza-South Sudan rift connection. Thermal history modelling of these data suggest that much of this region experienced pronounced, and seemingly uninterrupted cooling during the Cretaceous to middle Paleogene, incompatible with the regional subsidence that would be expected if the Anza and South Sudan rifts had established a hard linkage in the Turkana Depression. The proposed Anza-South Sudan rift linkage is further challenged by the scarcity of Cretaceous-early Paleogene strata in the region and high-gradient isobath pattern of pre-Cretaceous basement east of Lake Turkana [*Vétel and Le Gall, 2006*]. Thus, in contrast to the often-asserted assumption that the Turkana Depression was part of a through-going Anza-South Sudan rift system, this study suggests that the Turkana region may have acted as a regional high and possible axial sediment source for the adjacent grabens during Cretaceous-early Paleogene time. This may indicate that later Paleogene tectonism and/or plume activity played a more critical role in attenuating the Turkana crust than previously thought.

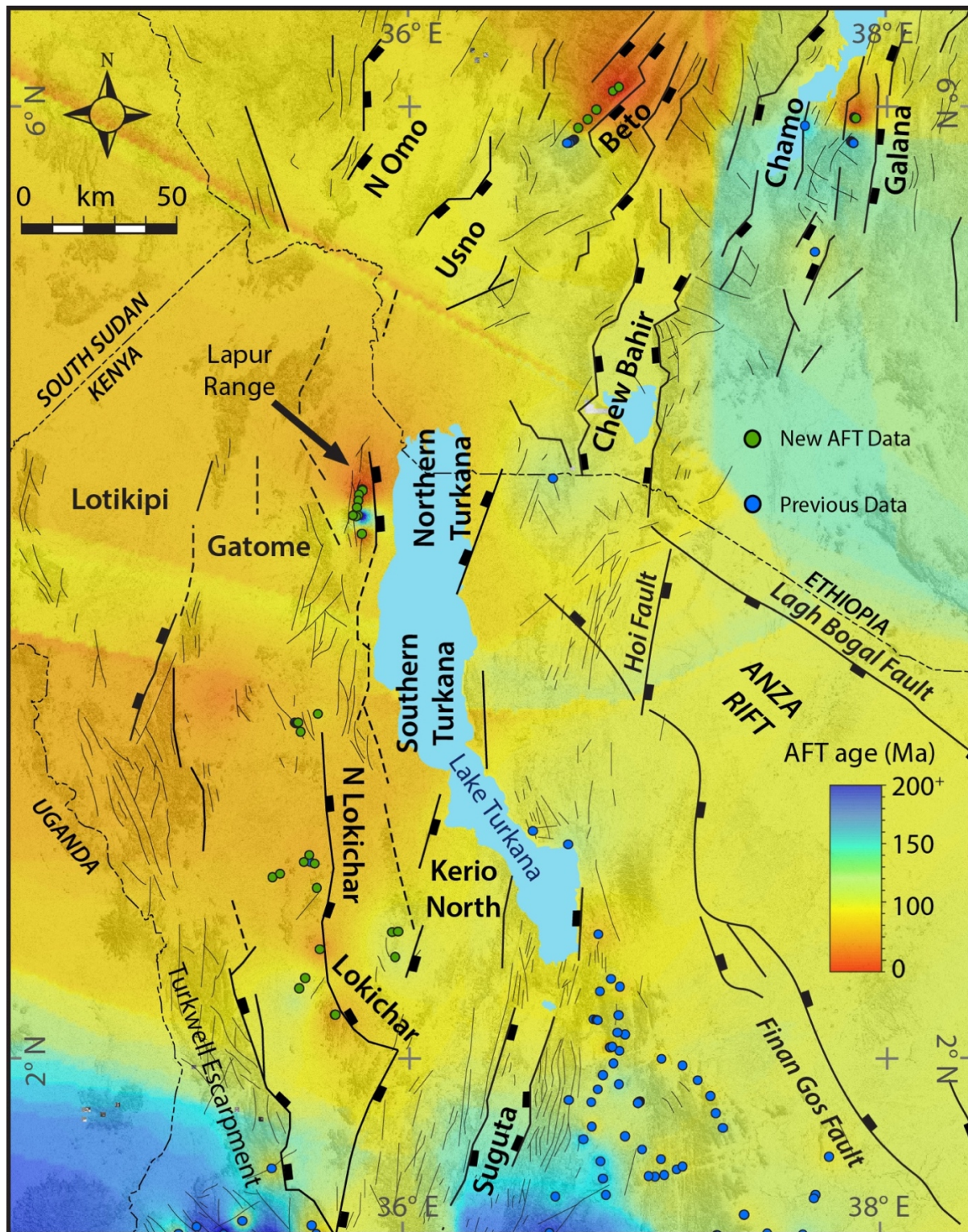


Figure 6.1. Data interpolation of Turkana Depression AFT ages (Ma), overlying a digital elevation model [NASA, 2000]. AFT ages are represented by colour, with colours becoming cooler with increasing ages. AFT data presented in this study are marked as green circles. Previous AFT ages (blue circles) were compiled from Gleadow [1980], Wagner et al. [1992], Foster & Gleadow [1992, 1993 and 1996], Philippon et al. [2014], Torres Acosta et al. [2015] and Balestrieri et al. [2016].

Evidence does exist, however, to suggest that isolated depocentres developed intermittently during the Cretaceous-early Paleogene throughout much of the greater Turkana Depression. The best example is the occurrence of up to 500 m of Upper Cretaceous-lower Paleogene fluvial sandstone in the Lapur Range [**Chapter 5**; *Wescott et al.*, 1999; *Tiercelin et al.*, 2012]. However, there is no evidence for rift-related faulting in the Lapur Sandstone, suggesting these strata may represent passive sedimentation during axial transport into the Anza graben [*Morley et al.*, 1992]. Other thin [generally < 100 m, *Wescott et al.*, 1993; *Tiercelin et al.*, 2012], discontinuous successions of immature, arkosic grits are found scattered throughout much of the Turkana Depression (Figure 1.2) [e.g. *Arambourg and Wolff*, 1969; *Walsh and Dodson*, 1969; *Savage and Williamson*, 1978; *Williamson and Savage*, 1986; *Wescott et al.*, 1993]. The age and nature of these so-called ‘Turkana Grits’ are poorly constrained due to their largely unfossiliferous and weathered state [*Wescott et al.*, 1993; *Tiercelin et al.*, 2004], with previous authors estimating that deposition occurred at any time between the Cretaceous and Miocene [*Savage and Williamson*, 1978; *Williamson and Savage*, 1986]. Where faulting has been observed in these depocentres, such as in the Muruanachok Hills [*Vétel and Le Gall*, 2006] and in the Mt Porr area [*Williamson and Savage*, 1986], fault plane analysis indicates a NE-SW extension direction, compatible with the Cretaceous paleostress field inferred to have controlled Anza rifting [*Morley et al.*, 1999b].

Sedimentation in these dispersed depocentres had ceased by the middle Eocene, in some cases abruptly (e.g. BRZ and Lapur Range), with the onset of extensive Eocene-Oligocene plume-related volcanism in the region [*Tiercelin et al.*, 2012]. If the apparent end of clastic sedimentation represents a cessation of subsidence in the Turkana Depression, then the contemporaneous Eocene arrival of a mantle plume in the region may suggest a causative relationship, whereby buoyancy forces associated with the mantle thermal anomaly destroyed any accommodation space (**Chapter 3**).

Despite Cretaceous?-early Paleogene sedimentation in the Turkana Depression appearing to have been widely distributed, the limited thickness of these scattered successions and concurrent regional cooling recorded by basement rocks (this study) suggest that subsidence in the region was limited in magnitude prior to the late Paleogene. Nonetheless, the causal mechanism for Cretaceous?-early Paleogene subsidence remains unknown, as the structural architecture and geometry of these basins are hidden in the subsurface or have largely been

eroded away. One possibility is that these small basins represent a sort of accommodation zone that developed late in the evolution of the surrounding Anza and South Sudan grabens.

Why the large and well developed Anza [> 600 km long; β of ~ 2 , *Reeves et al.*, 1987] and South Sudan rift systems [~ 800 km long, β of > 1.5 , *Mohamed*, 1996] were unable to fully link is unknown. One possible explanation could be that the lateral propagation of NW-SE striking Anza-South Sudan normal faults was restricted by the presence of pre-existing, transverse lithospheric-scale structures in the Turkana subsurface. Such a deep-seated $\sim$ N-S structure has been inferred to extend through much of Ethiopia, corresponding to a significant change in crustal thickness beneath the surrounding eastern (38-40 km) and western (34-36 km) rift shoulders [*Keranen and Klemperer*, 2008]. This structure has been interpreted as a Neoproterozoic N-NNE trending suture between East and West Gondwana, which approximately follows the trend of the Ethiopian Rift [*Berhe*, 1990; *Stern et al.*, 1990; *Keranen and Klemperer*, 2008]. Its southern extent however, is unknown. Alternatively, a hard linkage of the Anza and South Sudan rifts may have been interrupted by the Eocene onset of plume activity and/or a change in intracontinental stress field?

6.2 Eocene birth of the East African Rift System in the Turkana Depression

6.2.1 Arrival of the African Superplume

The emplacement of voluminous flood basalts in southern Ethiopia and northernmost Kenya ca. 45 Ma is widely regarded as marking the initiation of the EARS [e.g. *Ebinger et al.*, 2000; *Vétel and Le Gall*, 2006; *Purcell*, 2017; *Rooney*, 2017]. The onset of magmatism prior to extensional deformation in the Turkana Depression suggests that volcanism was generated by an asthenospheric thermal anomaly, not by decompression melting [e.g. *Morley*, 1994; *Ebinger et al.*, 2000; *Pik et al.*, 2008]. This interpretation is further supported by geochemical analysis of early volcanic flows [e.g. *Pik et al.*, 2006; *Furman et al.*, 2007; *Rooney et al.*, 2012, 2017]. While debate regarding the number and distribution of mantle plumes operating beneath East Africa is ongoing [*Ebinger and Sleep*, 1998; *Weeraratne et al.*, 2003; *Simmons et al.*, 2007; *Li et al.*, 2008; *Chang and Van der Lee*, 2011; *Fishwick and Bastow*, 2011], more recent studies seem to favour models which invoke the EARS having developed above one extensive ‘African Superplume’ [e.g. *Grand et al.*, 1997; *Ritsema et al.*, 1999; *Furman*, 2007; *Hansen et al.*, 2012].

The localisation of plume-related magmatism is thought to be controlled by pre-existing lithosphere-asthenosphere boundary topography [e.g. *Ebinger and Sleep, 1998; Sleep et al., 2002; Koptev et al., 2017*]. Thus, the nucleation of initial plume volcanism in the northern Turkana Depression has been explained by the channelling of upwelling magma into a ~NW-SE trending zone of pre-thinned lithosphere inherited from earlier Anza-South Sudan rifting [*Ebinger and Sleep, 1998*]. This theory requires that the Anza and South Sudan rifts once connected and extended through Turkana resulting in a significant degree of lithospheric attenuation. However, as discussed above, data presented here record regional Cretaceous-early Paleogene cooling west of Lake Turkana, suggesting that the region underwent little subsidence during that time (see **Chapter 3**). Thus, either the limited degree of Cretaceous-early Paleogene lithospheric thinning inferred to have affected Turkana was sufficient to promote ponding of plume material beneath the region [*Sleep et al., 2002*], or another mechanism for the early nucleation of plume-magmatism in the Turkana Depression is required. Alternatively, Cretaceous-early Paleogene lithospheric attenuation was more substantial further west beneath the poorly understood rifts along the South Sudanese-Ethiopian border [*Coffin and Rabinowitz, 1988; Ebinger et al., 2000*]. Such a scenario would explain the locus of earliest flood basalts being centred in southwestern Ethiopia.

6.2.2 Nucleation of strain circa 45-40 Ma

In comparison to the onset of rift-related volcanism, the exact timing and distribution of earliest ~E-W upper crustal extensional deformation in East Africa has been poorly constrained [e.g. *Macgregor, 2015; Purcell, 2017*]. This is primarily a result of poor surface exposure of earliest syn-rift strata, limited exploration well data and a lack of biostratigraphic and geochronological dating horizons. Nevertheless, the earliest manifestation of strain in the EARS is thought to have occurred in Turkana, with the development of the Paleogene-middle Miocene Lokichar Basin [e.g. *Morley et al., 1999a; Vétel and Le Gall, 2006; Purcell, 2017*]. The restriction of the deepest syn-rift infill to the basin subsurface [*Morley et al., 1999a*] has until now prevented a determination of the formational age of the Lokichar Basin. Low temperature thermochronology of the half-graben margins (**Chapter 2**) suggests that the Lokichar Basin began developing sometime between ~45-40 Ma, significantly predating faulting elsewhere in the EARS (Figure 6.2). Evidence for possible Eocene ~E-W extension has also been documented elsewhere in the Turkana Depression. Both the 1100 m expansion of Eocene-Oligocene basalts across a N-S fault east of Lake Turkana [*Vétel and Le Gall,*

2006] and the recognition of two ~N-S basins in the western Turkana subsurface [Wescott *et al.*, 1999] suggest that Eocene rifting may have been more widespread in the region.

The much earlier occurrence of E-W extension in the Turkana Depression compared to elsewhere in East Africa suggests that a unique set of circumstances facilitated strain nucleation in the region. While the nucleation of strain in other early segments of the EARS was assisted by thermal modification and weakening of the lithosphere by mantle plume activity [e.g. Afar, *Ebinger and Sleep*, 1998; *Bellahsen et al.*, 2003] or by pre-existing lithospheric weakness zones inherited from earlier tectonism [e.g. Rukwa Rift, *Roberts et al.*, 2010, 2012], faulting in the Turkana Depression may have been promoted by both mechanisms. The concentration of Eocene mantle plume activity in the region likely weakened the Turkana lithosphere [e.g. *Ebinger and Sleep*, 1998; *Ebinger et al.*, 2000], whose strength may have already been reduced by earlier Cretaceous-early Paleogene attenuation [*Brune et al.*, 2017], while pre-existing transverse brittle weakness zones may have served as focal points for strain accommodation [*Vétel and Le Gall*, 2006].

By constraining the spatio-temporal onset of ~E-W extension in East Africa to the Turkana Depression, questions regarding the forces driving early EARS development can now be sharpened. While modern intraplate deformation is thought to be generated by lateral gradients in gravitational potential energy arising from the high topography of East Africa [e.g. *Coblentz and Sandiford*, 1994; *Coblentz et al.*, 1994; *Stamps et al.*, 2010, 2014, 2015], it is uncertain if similar buoyancy forces existed during the Eocene. For instance, the high Ethiopian and East African plateaux are not thought to have developed until Oligocene time [e.g. *Pik et al.*, 2003, 2008; *Wichura et al.*, 2010; *Pik*, 2011; *Moucha and Forte*, 2011]. Marked Paleogene cooling recorded in the central Turkana Depression (**Chapter 3**) and western Anza graben [*Morley et al.*, 1999b], and a coeval termination of sedimentation throughout the region [*Ebinger et al.*, 1993; *Tiercelin et al.*, 2012] may indicate that a period of thermal uplift accompanied the Eocene arrival of the African Superplume beneath Turkana [*Ebinger and Sleep*, 1998]. However, strong supporting evidence for regional domal topography is still lacking [*Vétel and Le Gall*, 2006]. Nevertheless, even if Eocene tension was generated by buoyancy forces associated with a Paleogene ‘Turkana Dome’, the presence of a directional far-field stress field would have been required to develop non-axisymmetric, linear basins [*Burov and Gerya*, 2014], such as the N-S trending Lokichar Basin. Thus, regardless of the topographic state of the Turkana region at the onset of rifting,

some other mechanism was required to have initiated Eocene upper crustal dilation. The apparent synchronicity between the rotation of the East African paleostress regime from ~NE-SW to ~E-W and a major reorganization of plate motion in the Indian Ocean ca. ~45-42 Ma [Cande *et al.*, 2018] suggests that plate boundary forces may have played an important role during EARS inception.

6.3 Rift migration in the Turkana Depression

Thermochronology data presented in this study serve to better constrain the propagation of EARS-related strain within the Turkana Rift. A period of reheating between ~30-10 Ma (**Chapter 3**) is interpreted as representing burial during the Oligocene-middle Miocene formation of the Lothidok Basin and marking the northward migration of strain from the Lokichar region. Faulting had permeated into southern Ethiopia by the Miocene [**Chapter 4**; Ebinger *et al.*, 2000; Bonini *et al.*, 2005; Pik *et al.*, 2008]. The low lithospheric strength [Hendrie *et al.*, 1994; Pérez-Gussinyé *et al.*, 2009] and elevated heat flow of the northern Turkana Depression [Morley *et al.*, 1999a] resulted in amplified isostatic footwall uplift and increased denudational cooling of rift basin margins (**Chapter 5**), generating the only rift-related Neogene AFT ages in the region (Figure 6.3). Simultaneously, faulting propagated southwards into the central Kenyan Rift, with the South Kerio Rift developing sometime during the Oligocene-early Miocene [Morley *et al.*, 1992; Mugisha *et al.*, 1997; Hautot *et al.*, 2000; Purcell, 2017].

During Neogene-Recent time, extensional deformation also migrated eastwards within the Turkana Depression [Morley *et al.*, 1999a; Vétel and Le Gall, 2006]. The Turkana and Kerio North basins formed in the middle Miocene beneath what is now Lake Turkana (**Chapter 5**) and were later followed by the development of the Pliocene-Pleistocene Kino Sogo Fault Belt in eastern Turkana [Vétel *et al.*, 2005]. The eastward propagation of rifting may reflect strain hardening effects caused by a relatively low strain rate [Morley, 1994; Morley *et al.*, 1992, 1999a]. However, a similar eastward shift of Turkana surface volcanism [e.g. McDougall and Brown, 2009] may suggest that rift propagation reflected an eastward migration of the underlying mantle thermal anomaly over time [Morley, 1994; Ebinger *et al.*, 2000].

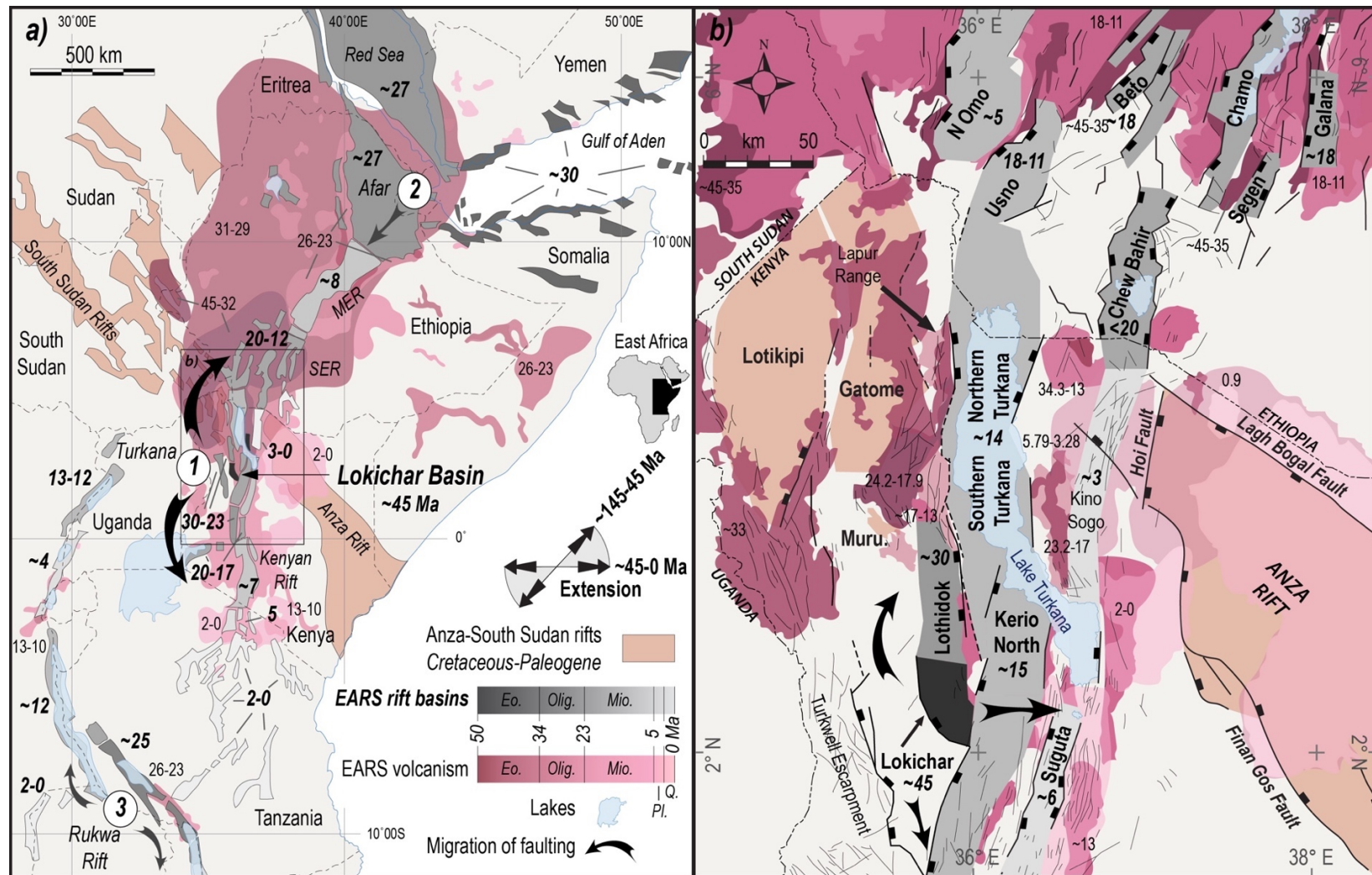


Figure 6.2. Chronology of Cretaceous-Recent rift basin formation and EARS-related volcanism in **a)** East Africa and **b)** the Turkana Depression. Colours of EARS basins (grey) and volcanics (red) darken with age. **a)** EARS volcanism and faulting is suggested to have nucleated within inherently weak lithospheric zones (1: Turkana Depression, 2: Afar, 3: Rukwa Rift) before propagating north and south. **b)** Volcanic activity within the Turkana Depression propagated both southwards and eastwards after initial eruption of voluminous flood basalts in the northern Turkana Depression ca. 45 Ma [Ebinger et al., 2000]. After nucleating in the Lokichar Basin area, strain within the Turkana Depression migrated towards the north, south and east with time. Basin names and formation ages (italics) in bold; volcanic ages in normal font style. Distribution of basins and surface volcanics, and corresponding ages from this study, Vetél and Le Gall [2006], Macgregor [2015], Purcell et al. [2017], Rooney et al. [2016] and references therein. MER: Main Ethiopian Rift; SER: Southern Ethiopian Rift; Muru: Muruanachok Hills.

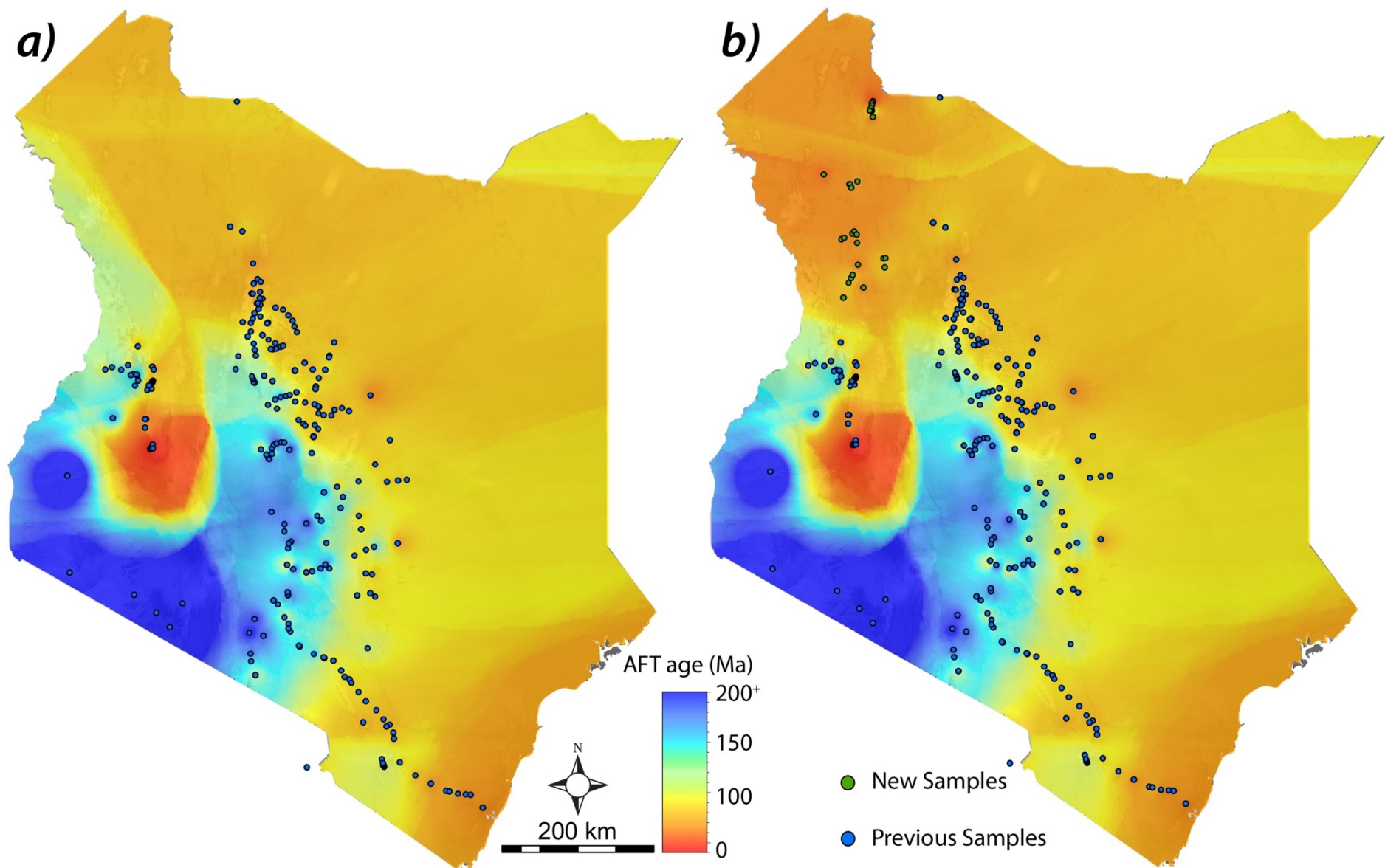


Figure 6.3. Data interpolations of **a)** previously published Kenyan AFT ages and **b)** including newly acquired results from this study. AFT ages are represented by colour, with colours becoming cooler with increasing ages. Interpolations overlay a digital elevation model (DEM) [NASA, 2000]. AFT age trends are discussed in the text. Previously published AFT ages (Ma) were compiled from Gleadow [1980], Wagner et al. [1992], Foster & Gleadow [1992, 1993 and 1996], Noble et al. [1997], Spiegel et al. [2007], Torres Acosta et al., [2015].

6.4 Remaining uncertainties and future work

This study provides a tectono-thermal framework to guide future investigations into Cretaceous-early Paleogene rifting in East Africa and the dynamic interplay of rift superposition, pre-existing rheological heterogeneities and mantle processes during inception of the EARS. Nevertheless, several key questions regarding the chronology of rifting and the evolution of topography in the region still need to be further addressed. These include:

- Why were the Anza and South Sudan rifts unable to develop a hard linkage within the Turkana Depression? Was the development of such a connection simply interrupted by the Eocene arrival of a mantle thermal anomaly beneath the region and/or a change in far-field stress field? Or did something about the inherent basement architecture of Turkana prevent the two systems from connecting, such as the presence of transverse lithospheric-scale fabrics?
- How extensive was early Paleogene sedimentation in the Turkana Depression? Was deposition associated with faulting? If not, what mechanism was responsible for the creation of accommodation space at that time?
- What is the chronology of crustal attenuation in the Turkana Depression? Was Cretaceous-early Paleogene crustal thinning sufficient by Eocene time to channel upwelling plume material, as suggested by *Ebinger and Sleep* [1998]? Or does the much earlier nucleation of EARS strain in the Turkana Depression account for the significant north-south discrepancy in East African crustal thickness? What does this tell us about the ability of the Turkana lithosphere to isostatically support large-scale topography in the Paleogene?
- What caused the inferred Eocene rotation of the paleostress field from ~NW-SE to ~W-E? Was this only a local phenomenon, or does the observed shift in paleostress orientation reflect a more regional change in plate motion?
- How extensive was Eocene ~E-W extensional deformation in the Turkana Depression?

Nevertheless, this study provides a tectono-thermal framework to guide future investigations into Cretaceous-early Paleogene rifting in East Africa and the dynamic interplay of rift superposition, pre-existing rheological heterogeneities and mantle processes during inception of the EARS.

References

- Arambourg, C., and R. G. Wolff (1969), Nouvelles données paléontologique sur l'âge des "Gres du Lubur" (Turkana Grits) al'ouest du lac Rodolphe, *Comptes Rendus Société géologique Fr.*, 9, 190–202.
- Balestrieri, M. L., M. Bonini, G. Corti, F. Sani, and M. Philippon (2016), A refinement of the chronology of rift-related faulting in the Broadly Rifted Zone, southern Ethiopia, through apatite fission-track analysis, *Tectonophysics*, 671, 42–55, doi:10.1016/j.tecto.2016.01.012.
- BEICIP (1987), Geological Map of Kenya,
- Bellahsen, N., C. Faccenna, F. Funiciello, J. M. Daniel, and L. Jolivet (2003), Why did Arabia separate from Africa? Insights from 3-D laboratory experiments, *Earth Planet. Sci. Lett.*, 216, 365–381, doi:10.1016/S0012-821X(03)00516-8.
- Benoit, M. H., A. A. Nyblade, and M. E. Pasyanos (2006), Crustal thinning between the Ethiopian and East African plateaus from modeling Rayleigh wave dispersion, *Geophys. Res. Lett.*, 33(13), doi:10.1029/2006GL025687.
- Berhe, S. M. (1990), Ophiolites in Northeast and East Africa: implications for Proterozoic crustal growth, *J. Geol. Soc. London*, 147, 41–57.
- Bonini, M., G. Corti, F. Innocenti, P. Manetti, F. Mazzarini, T. Abebe, and Z. Pecskey (2005), Evolution of the Main Ethiopian Rift in the frame of Afar and Kenya rifts propagation, *Tectonics*, 24(1), n/a-n/a, doi:10.1029/2004TC001680.
- Bosworth, W. (1992), Mesozoic and early Tertiary rift tectonics in East Africa, *Tectonophysics*, 209(1–4), 115–137, doi:10.1016/0040-1951(92)90014-W.
- Brune, S., G. Corti, and G. Ranalli (2017), Controls of inherited lithospheric heterogeneity on rift linkage: Numerical and analog models of interaction between the Kenyan and Ethiopian rifts across the Turkana depression, *Tectonics*, 1–30, doi:10.1002/2017tc004739.
- Burov, E., and T. Gerya (2014), Asymmetric three-dimensional topography over mantle plumes, *Nature*, 513(7516), 85–89, doi:10.1038/nature13703.
- Cande, S. C., P. Patriat, and J. Dymert (2018), Motion between the Indian, Antarctic and African plates in the early Cenozoic, *Geophys. J. Int.*, 183, 127–149, doi:10.1111/j.1365-246X.2010.04737.x.
- Chang, S.-J., and S. Van der Lee (2011), Mantle plumes and associated flow beneath Arabia and East Africa, *Earth Planet. Sci. Lett.*, 302, 448–454, doi:10.1016/j.epsl.2010.12.050.
- Coblentz, D., and M. Sandiford (1994), Tectonic stresses in the African plate: Constraints on the ambient lithospheric stress state, *Geology*, 22(September), 831–834, doi:10.1130/0091-7613(1994)022<0831.
- Coblentz, D. D., R. M. Richardson, and M. Sandiford (1994), On the gravitational potential of the Earth's lithosphere, *Tectonics*, 13(4), 929–945.
- Coffin, M. F., and P. D. Rabinowitz (1988), Evolution of the conjugate East African-Madagascar margins and the western Somali Basin, *Geol. Soc. Am. Spec. Pap.*, 226, 1–79.
- Davidson, A. (1983), *The Omo River Project: reconnaissance geology and geochemistry of parts of Ilubabor, Kefa, Gemu Gofa and Sidamo, Ethiopia.*
- Ebinger, C. J., and A. Ibrahim (1994), Multiple episodes of rifting in Central and East Africa: A re-evaluation of gravity data, *Geol. Rundschau*, 83(4), 689–702, doi:10.1007/BF00251068.
- Ebinger, C. J., and N. H. Sleep (1998), Cenozoic magnetism throughout east Africa resulting from impact of a single plume, *Nature*, 395, 788–791.

- Ebinger, C. J., T. Yemane, D. J. Harding, S. Tesfaye, S. Kelley, and D. C. Rex (2000), Rift deflection, migration, and propagation: Linkage of the Ethiopian and Eastern rifts, Africa, *GSA Bull.*, *112*(2), 163–176.
- Fishwick, S., and I. D. Bastow (2011), Towards a better understanding of African topography: A review of passive-source seismic studies of the African crust and upper mantle, *Geol. Soc. London, Spec. Publ.*, *357*, 343–371, doi:10.1144/SP357.19.
- Foster, D. A., and A. J. W. Gleadow (1992), The morphotectonic evolution of rift-margin mountains in central Kenya: Constraints from apatite fission-track thermochronology, *Earth Planet. Sci. Lett.*, *113*(1–2), 157–171, doi:10.1016/0012-821X(92)90217-J.
- Foster, D. A., and A. J. W. Gleadow (1993), Episodic Denudation in East Africa: A Legacy of Intracontinental Tectonism, *Geophys. Res. Lett.*, *20*(21), 2395–2398.
- Foster, D. A., and J. W. Gleadow (1996), Structural framework and denudation history of the flanks of the Kenya and Anza Rifts, East Africa, *Tectonics*, *15*(2), 258–271.
- Furman, T. (2007), Geochemistry of East African Rift basalts: An overview, *J. African Earth Sci.*, *48*, 147–160, doi:10.1016/j.jafrearsci.2006.06.009.
- Gleadow, A. J. W. (1980), Fission track age of the KBS Tuff and associated hominid remains in northern Kenya, *Nature*, *284*, 225–230, doi:10.1038/284225a0.
- GMRD-Sudan (1981), Geological Map of the Sudan,
- Grand, S. P., R. D. van der Hilst, and S. Widiyantoro (1997), Global Seismic Tomography: A Snapshot of Convection in the Earth, *GSA Today*, *7*(4), 1–7.
- Guiraud, R., and W. Bosworth (1997), Senonian basin inversion and rejuvenation of rifting in Africa and Arabia : synthesis and implications to plate-scale tectonics, *Tectonophysics*, *282*, 39–82.
- Hansen, S. E., A. A. Nyblade, and M. H. Benoit (2012), Mantle structure beneath Africa and Arabia from adaptively parameterized P-wave tomography: Implications for the origin of Cenozoic Afro-Arabian tectonism, *Earth Planet. Sci. Lett.*, *319–320*, 23–34, doi:10.1016/j.epsl.2011.12.023.
- Hautot, S., P. Tarits, K. Whaler, B. Le Gall, J.-J. Tiercelin, and C. Le Turdu (2000), Deep structure of the Baringo Rift Basin (central Kenya) from three-dimensional magnetotelluric imaging: Implications for rift evolution, *J. Geophys. Res.*, *105*(B10), 23,493–23,518, doi:10.1029/2000JB900213.
- Hendrie, D. B., N. J. Kusznir, C. K. Morley, and C. J. Ebinger (1994), Cenozoic extension in northern Kenya: a quantitative model of rift basin development in the Turkana region, *Tectonophysics*, *236*(1–4), 409–438, doi:10.1016/0040-1951(94)90187-2.
- Ibrahim, A. E., C. J. Ebinger, and J. D. Fairhead (1991), Interpretation of the Central African Rift system in Sudan based on new gravity and aeromagnetic data, *Eos (Washington. DC).*, *462*.
- Kazmin, V. (1972), Geological Map of Ethiopia,
- Keranen, K., and S. L. Klemperer (2008), Discontinuous and diachronous evolution of the Main Ethiopian Rift: Implications for development of continental rifts, *Earth Planet. Sci. Lett.*, *265*, 96–111, doi:10.1016/j.epsl.2007.09.038.
- Koptev, A., S. Cloetingh, T. Gerya, E. Calais, and S. Leroy (2017), Non-uniform splitting of a single mantle plume by double cratonic roots: Insight into the origin of the central and southern East African Rift System, *Terra Nov.*, *30*, 125–134, doi:10.1111/ter.12317.
- Lehto, T. et al. (2014), Geological Map of Uganda, Scale 1:1 000 000,
- Li, C., R. D. van der Hilst, E. R. Engdahl, and S. Burdick (2008), A new global model for P wave speed variations in Earth's mantle, *Geochemistry Geophys. Geosystems*, *9*(5), 1–21, doi:10.1029/2007GC001806.
- Macgregor, D. (2015), History of the development of the East African Rift System: A series of interpreted maps through time, *J. African Earth Sci.*, *101*, 232–252,

- doi:10.1016/j.jafrearsci.2014.09.016.
- Mariita, N. O., and G. R. Keller (2007), An integrated geophysical study of the northern Kenya rift, *J. African Earth Sci.*, 48(2–3), 80–94, doi:10.1016/j.jafrearsci.2006.05.008.
- McDougall, I., and F. H. Brown (2009), Timing of volcanism and evolution of the northern Kenya Rift, *Geol. Mag.*, 146(1), 34–37, doi:10.1017/S0016756808005347.
- Mohamed, A. Y. (1996), Basin analysis and hydrocarbon maturation, Unity and Kaikang area, Muglad Basin, Sudan, Aberdeen.
- Morley, C. K. (1994), Interaction of deep and shallow processes in the evolution of the Kenya rift, *Tectonophysics*, 236, 81–91.
- Morley, C. K., W. a. Wescott, D. M. Stone, R. M. Harper, S. T. Wigger, and F. M. Karanja (1992), Tectonic evolution of the northern Kenyan Rift, *J. Geol. Soc. London.*, 149(3), 333–348, doi:10.1144/gsjgs.149.3.0333.
- Morley, C. K., D. M. Stone, R. M. Harper, and S. T. Wigger (1999), Geology and Geophysics of the Western Turkana Basins, Kenya, in *Geoscience of Rift Systems—Evolution of East Africa: AAPG Studies in Geology No. 44*, edited by C. K. Morley, pp. 19–54.
- Moucha, R., and A. M. Forte (2011), Changes in African topography driven by mantle convection, *Nat. Geosci.*, 4(10), 707–712, doi:10.1038/ngeo1235.
- Mugisha, F., C. J. Ebinger, M. Strecker, and D. Pope (1997), Two-stage rifting in the Kenya rift: implications for half-graben models, *Tectonophysics*, 278, 63–81.
- NASA (2000), Shuttle Radar Topography Mission (SRTM) 1 Arc-Second Global,
- Noble, W. P., D. A. Foster, and A. J. W. Gleadow (1997), The post-Pan-African thermal and extensional history of crystalline basement rocks in eastern Tanzania, *Tectonophysics*, 275(4), 331–350, doi:10.1016/S0040-1951(97)00026-7.
- Pérez-Gussinyé, M., M. Metois, M. Fernández, J. Vergés, J. Fullea, and A. R. Lowry (2009), Effective elastic thickness of Africa and its relationship to other proxies for lithospheric structure and surface tectonics, *Earth Planet. Sci. Lett.*, 287, 152–167, doi:10.1016/j.epsl.2009.08.004.
- Philippon, M., G. Corti, F. Sani, M. Bonini, M. Balestrieri, P. Molin, E. Willingshofer, D. Sokoutis, and S. Cloetingh (2014), Evolution, distribution, and characteristics of rifting in southern Ethiopia, *Tectonics*, 33, 1–24, doi:10.1002/2013TC003430.
- Pik, R. (2011), Geodynamics: East Africa on the rise, *Nat. Geosci.*, 4(10), 660–661, doi:10.1038/ngeo1274.
- Pik, R., B. Marty, J. Carignan, and J. Lavé (2003), Stability of the Upper Nile drainage network (Ethiopia) deduced from (U-Th)/He thermochronometry: implications for uplift and erosion of the Afar plume dome, *Earth P.*, 215, 73–88, doi:10.1016/S0012-821X(03)00457-6.
- Pik, R., B. Marty, and D. R. Hilton (2006), How many mantle plumes in Africa? The geochemical point of view, *Chem. Geol.*, 226, 100–114, doi:10.1016/j.chemgeo.2005.09.016.
- Pik, R., B. Marty, J. Carignan, G. Yirgu, and T. Ayalew (2008), Timing of East African Rift development in southern Ethiopia : Implication for mantle plume activity and evolution of topography, *Geology*, 36(2), 167–170, doi:10.1130/G24233A.1.
- Purcell, P. G. (2017), Re-imagining and re-imaging the development of the East African Rift, *Pet. Geosci.*, 36, doi:https://doi.org/10.1144/petgeo2-17-036.
- Reeves, C. V., F. M. Karanja, and I. N. MacLeod (1987), Geophysical evidence for a failed Jurassic rift and triple junction in Kenya, *Earth Planet. Sci. Lett.*, 81(2–3), 299–311, doi:10.1016/0012-821X(87)90166-X.
- Ritsema, J., H.-J. van Heijst, and J. H. Wodhouse (1999), Complex Shear Wave Velocity Structure Imaged Beneath Africa and Iceland, *Science (80-.)*, 286(5446), 1925–1928.

- Roberts, E. M., P. M. O. Connor, N. J. Stevens, M. D. Gottfried, Z. A. Jinnah, S. Ngasala, A. M. Choh, and R. A. Armstrong (2010), Sedimentology and depositional environments of the Red Sandstone Group, Rukwa Rift Basin, southwestern Tanzania: New insight into Cretaceous and Paleogene terrestrial ecosystems and tectonics in sub-equatorial Africa, *J. African Earth Sci.*, *57*, 179–212, doi:10.1016/j.jafrearsci.2009.09.002.
- Roberts, E. M., N. J. Stevens, P. M. O'Connor, P. H. G. M. Dirks, M. D. Gottfried, W. C. Clyde, R. A. Armstrong, a. I. S. Kemp, and S. Hemming (2012), Initiation of the western branch of the East African Rift coeval with the eastern branch, *Nat. Geosci.*, *5*(4), 289–294, doi:10.1038/ngeo1432.
- Rooney, T. O. (2017), The Cenozoic magmatism of East-Africa: Part I — Flood basalts and pulsed magmatism, *Lithos*, *286–287*, 264–301, doi:10.1016/j.lithos.2017.05.014.
- Rooney, T. O., C. Herzberg, and I. D. Bastow (2012), Elevated mantle temperature beneath East Africa, *Geology*, *40*(1), 27–30, doi:10.1130/G32382.1.
- Savage, R. J. G., and P. G. Williamson (1978), The early history of the Turkana depression, *Geol. Soc. London Spec. Publ.*, *6*(1), 375–394.
- Schull, T. J. (1988), Rift Basins of Interior Sudan: Petroleum Exploration and Discovery, , *72*(10), 1128–1142.
- Simmons, N. A., A. M. Forte, and S. P. Grand (2007), Thermochemical structure and dynamics of the African superplume, *Geophys. Res. Lett.*, *34*, 1–5, doi:10.1029/2006GL028009.
- Sleep, N. H., C. J. Ebinger, and J.-M. Kendall (2002), Deflection of mantle plume material by cratonic keels, *Geol. Soc. London, Spec. Publ.*, *1999*, 135–150.
- Spiegel, C., B. P. Kohn, D. X. Belton, and A. J. W. Gleadow (2007), Morphotectonic evolution of the central Kenya rift flanks: Implications for late Cenozoic environmental change in East Africa, *Geology*, *35*(5), 427–430, doi:10.1130/G23108A.1.
- Stamps, D. S., L. M. Flesch, and E. Calais (2010), Lithospheric buoyancy forces in Africa from a thin sheet approach, *Int. J. Earth Sci.*, *99*, 1525–1533, doi:10.1007/s00531-010-0533-2.
- Stamps, D. S., L. M. Flesch, E. Calais, and A. Ghosh (2014), Current kinematics and dynamics of Africa and the East African Rift System, *J. Geophys. Res. Solid Earth*, *119*, 5161–5186, doi:10.1002/2013JB010717. Received.
- Stamps, D. S., G. Iaffaldano, and E. Calais (2015), Role of mantle flow in Nubia-Somalia plate divergence, *Geophys. Res. Lett.*, *42*, 290–296, doi:10.1002/2014GL062515. Received.
- Stern, R. J., K. C. Nielsen, E. Best, M. Sultan, R. E. Arvidson, and A. Kröner (1990), Orientation of late Precambrian sutures in the Arabian-Nubian Shield, *Geology*, *18*, 1103–1106, doi:10.1130/0091-7613(1990)018<1103.
- Tiercelin, J.-J., T. Nalpas, P. Thuvo, and J.-L. Potdevin (2012), Hydrocarbon Prospectivity in Mesozoic and Early–Middle Cenozoic Rift Basins of Central and Northern Kenya, Eastern Africa, *AAPG Mem.*, *100*, 179–208, doi:10.1306/13351553M1001742.
- Torres Acosta, V., A. Bande, E. R. Sobel, M. Parra, T. F. Schildgen, F. Stuart, and M. R. Strecker (2015), Cenozoic extension in the Kenya Rift from low-temperature thermochronology: Links to diachronous spatiotemporal evolution of rifting in East Africa, *Tectonics*, *34*(12), 2367–2386, doi:10.1002/2015TC003949.
- Vétel, W., and B. Le Gall (2006), Dynamics of prolonged continental extension in magmatic rifts: the Turkana Rift case study (North Kenya), *Geol. Soc. London Spec. Publ.*, *259*, 209–233.
- Vétel, W., B. Le Gall, and J. J. Walsh (2005), Geometry and growth of an inner rift fault pattern: the Kino Sogo Fault Belt, Turkana Rift (North Kenya), *J. Struct. Geol.*, *27*, 2204–2222, doi:10.1016/j.jsg.2005.07.003.

- Wagner, M., R. Altherr, and P. van den Haute (1992), Apatite fission-track analysis of Kenyan basement rocks: constraints on the thermotectonic evolution of the Kenya dome. A reconnaissance study, *Tectonophysics*, 204(1–2), 93–110.
- Walsh, J., and R. G. Dodson (1969), Geology of Northern Turkana, Degree Sheets 1, 2, 9, and 10,
- Weeraratne, D. S., D. W. Forsyth, K. M. Fischer, and A. A. Nyblade (2003), Evidence for an upper mantle plume beneath the Tanzanian craton from Rayleigh wave tomography, *J. Geophys. Res.*, 108(B9), 1–14, doi:10.1029/2002JB002273.
- Wescott, W. A., C. K. Morley, and F. M. Karanja (1993), Geology of the “Turkana Grits” in the Lariu range and Mt. Porr areas, southern Lake Turkana, Northwestern Kenya, *J. African Earth Sci.*, 16(4), 425–435, doi:10.1016/0899-5362(93)90101-U.
- Wescott, W. A., S. T. Wigger, D. M. Stone, and C. K. Morley (1999), Geology and Geophysics of the Lotikipi Plain, *Geosci. Rift Syst. East Africa*, (44), 55–66.
- Wichura, H., R. Bousquet, R. Oberhänsli, M. R. Strecker, and M. H. Trauth (2010), Evidence for middle Miocene uplift of the East African Plateau, *Geology*, 38(6), 543–546, doi:10.1130/G31022.1.
- Williamson, P. G., and R. J. G. Savage (1986), Early rift sedimentation in the Turkana basin, northern Kenya, *Geol. Soc. London, Spec. Publ.*, 25(1), 267–283, doi:10.1144/GSL.SP.1986.025.01.22.