Morley, C.K., D.K. Ngenoh, and J.K. Ego, 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, p. 1–18.

Chapter 1

Introduction to the East African Rift System

C.K. Morley (part 1)
Department of Petroleum Geoscience
University of Brunei Darussalam
Negara Brunei Darussalam

D.K. Ngenoh (part 2)
J.K. Ego (part 2)
National Oil Company of Kenya
Nairobi, Kenya

Abstract

The Tertiary age East African Rift System (EARS) is predominantly located in zones of Precambrian orogenic belts avoiding stable Archean cratonic areas. The most profound influence of preexisting fabrics is exerted by Precambrian ductile fabrics. However, later predominantly extensional events of Permo-Triassic (Karroo), Jurassic, Cretaceous, and Paleogene age have also variably influenced the location and orientation of the Tertiary rift systems.

The two branches of the East African Rift System have undergone different tectonic histories. In general, the western branch can be regarded as a good model of a young continental rift while the eastern branch is representative of a "failed" mature continental rift system. Both are characterized by large half graben systems filled by fluvio-deltaic and lacustrine sediments, and/or by volcanics and volcaniclastics. The basin fills can be up to 7–8 km thick. In comparison with the eastern branch, the western branch is younger (late Miocene-Recent), less volcanic-rich but more seismically active, with deeper earthquakes (down to about 30–40 km). Extension estimates for the eastern branch depend upon location, but range up to 40 km (Turkana area), maximum extension in the western branch is about 10–12 km.

The eastern branch was probably initiated in the Eocene. In the Kenya and Ethiopian Rifts, older half grabens were abandoned in the Pliocene in favor of a narrow rift zone characterized by minor fault swarms, dike intrusions, and volcanic centers. There is evidence for an important active rift component (very thin mantle lithosphere, large topographic domes, anomalously low velocities in the mantle) as well as passive rifting. Deep refraction profiles indicate extension ranges between 10–40 km in the Kenya Rift (increasing northwards). However, this may be an underestimate if the crust has been inflated by intrusion of magma during extension.

Part 1—Introduction to the East African Rift System

INTRODUCTION

This section aims to briefly review of some of the key geological features of the East African Rift System (EARS), as background information for the more detailed geology presented in following chapters.

The East African Rift System has been the classic area of continental rifting since the process was described in Kenya by Gregory (1896, 1921). Today, a number of different avenues of research are being undertaken in the rift system, the type of study and location very much dictated by the morphology of the region.

The East African Rift System forms a narrow (50–150 km wide), elongate system of normal faults that stretches some 3,500 km in a submeridian direction. It is connected to the world wide system of oceanic rifts via the Afar Triangle to the Gulf of Aden and the Red Sea (Figures 1 and 2). The EARS is composed of two rift trends called the eastern and western branches. The eastern branch, located north and east of Lake Victoria, is a volcanic-rich system that forms the Kenya and Ethiopian Rifts. In places the rift system was probably initiated in the early Miocene, but there is also evidence for earlier Paleogene rift activity in northern Kenya and Ethiopia (Hendrie et al. 1994, Ebinger et al. 1993, Chapters 2 and 3). The rift branch is well exposed at the surface and is dotted with relatively small lakes. Lake Turkana is the sole large lake in the eastern branch (Figure 3). The western branch appears to have been initiated later than the east-

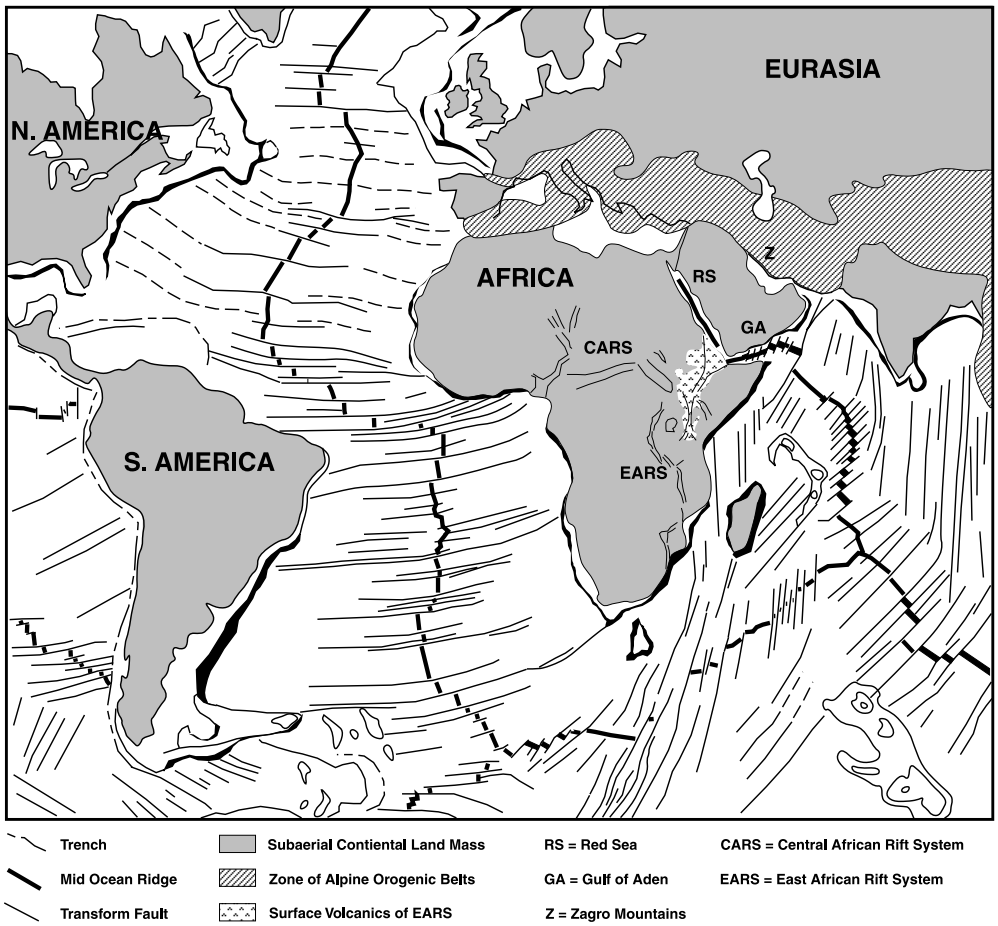


Figure 1. Simplified map of the rift systems surrounding Africa, (modified from Olivet, in Coussement 1995).

ern branch, during the late Miocene (Ebinger 1989). It is composed of a series of extensive deep and shallow lakes: Mobutu (or Albert), Edward (or Idi Amin), Kivu, Tanganyika, Rukwa, and Malawi (or Nyasa) (Figure 3). These lakes mask much of the rift structure. The western branch is associated with much less volcanism than is the eastern branch although both branches are seismically and volcanically active today (Figure 2).

At the largest scale, the topography of the EARS is characterized by two large lithospheric domes called the Afar and East African Domes (Figure 2). They are separated by the Turkana depression (600 m average elevation) of northern Kenya. The average elevations are 1,500 m for the Afar Dome and 1,200 m for the East African Dome. Away from these areas the topography ranges between 300–900 m (Ebinger et al. 1989). Both domes have diameters of about 1,000 km and are associated with large negative gravity anomalies (Figure 2). Within the East African Dome smaller domes (perhaps associated with magmatic underplating) with radii of 100–200 km are present, notably the Kivu and Kenya Domes.

The main rift basins of the western branch are largely covered by water. Hence outcrop studies are restricted to the rift margins. However, the lakes have two advantages—it is much easier to acquire seismic reflection data over water than land, and modern sedimentary processes in large lacustrine systems can be studied. These two advantages have provided the focus of much recent research into the western branch.

The eastern branch is dotted with small lakes, but outcrops within the rift system are numerous and extensive. Most of the exposed section is either volcanic flows, igneous intrusives, or pyroclastic sediments. Much of the outcrop work has centered around understanding the volcanic sequences stratigraphically and geochemically. It has taken a long time to build up a regional picture of the distribution and evolution of the igneous rocks because of the vast areas involved (e.g., Williams 1978, Macdonald et al. 1994). As a consequence, detailed geochemical studies of the igneous suites are relatively few and far between.

In places extensive deposits of young (Plio-Pleistocene), dominantly fluvio-lacustrine sediments are present. Some contain remarkable assemblages of fossils including vertebrate bones and a few have become well known sites for investigations into our hominid ancestors (e.g., Olduvai Gorge, Koobi Fora, Middle Awash Valley, and Hadar). However, exposures of older sedimentary sequences are comparatively rare. The volcanic rocks that cover much of the rift mask evidence of the older rift history. Nevertheless, in a few places within the rift, erosion through topographically high basement blocks provides a glimpse at the deeper rift section, which includes arkosic sandstones (e.g., Tugen

Hills, Kito Pass—Chapman et al. 1978, and the Turkana area—Williamson and Savage 1986, Wescott et al. 1993). Only in the Turkana area has the geology of these older sequences been studied in detail, and this was only possible due to hydrocarbon exploration. The field work, geophysical, and drilling results that shed much light on the Paleogene-Miocene history of the northern Kenya Rift are presented in Chapters 2 and 3.

The general structural pattern of the eastern branch has been ascertained from surface mapping and from satellite imagery (e.g., Baker et al. 1972, Bosworth and Maurin 1993, Grimaud et al. 1994). The relatively young deformation (late Miocene-Recent) is spectacularly exposed in places and displays the characteristics of normal fault systems very well (Crossley 1978, Griffiths 1980, Chapter 11). However, only a certain amount about the structure of the Kenya Rift can be learned from surface studies. One point that helps illustrate the problems of understanding the rift evolution from surface studies alone concerns the Lokichar Fault in the Turkana area (see Chapter 2). Despite the very high quality work conducted in the Kenya Rift during the 1970s and 1980s no published maps of the Kenya Rift system during that period showed the existence of the Lokichar Fault. Yet this fault is possibly the largest fault in the Kenya Rift and

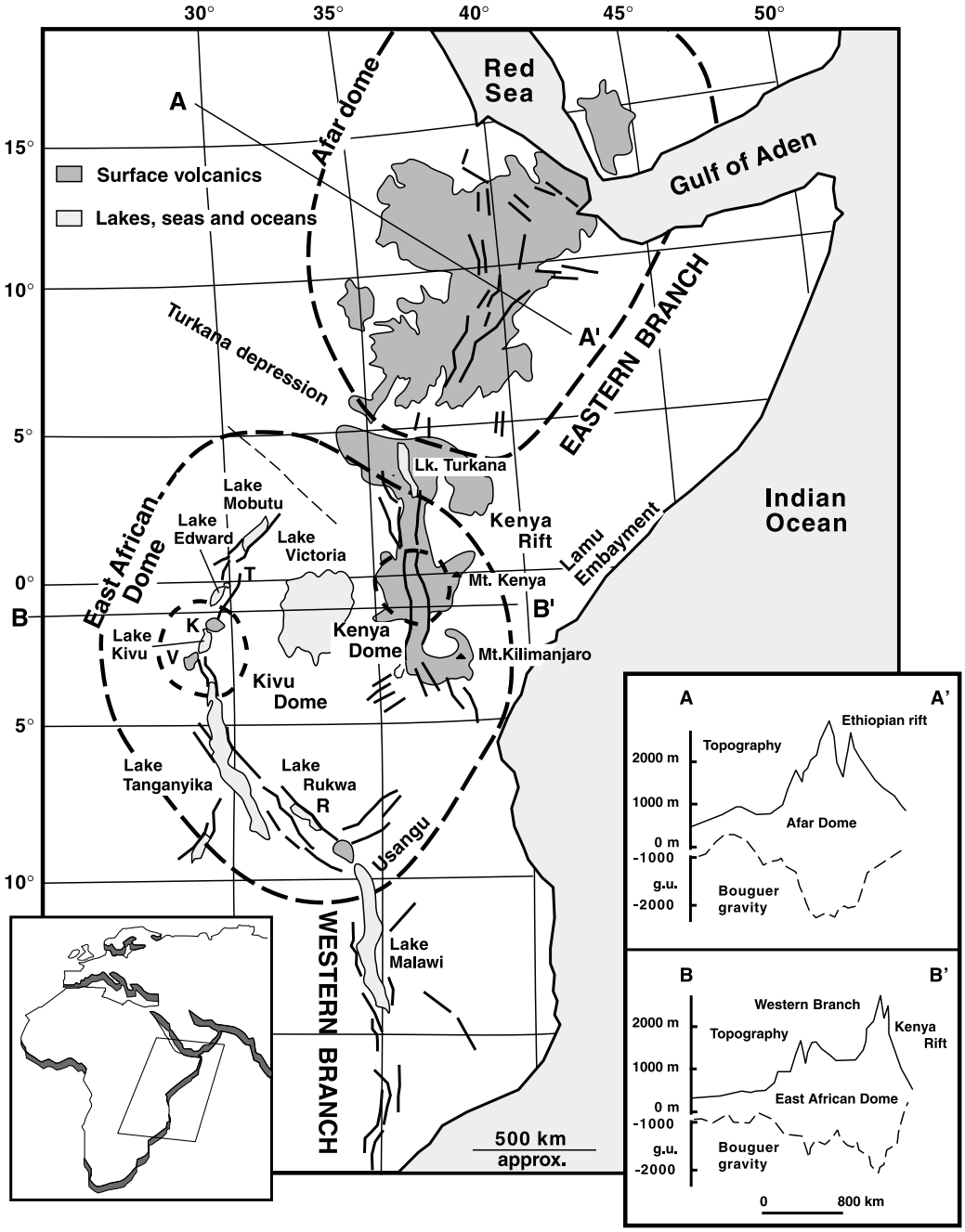


Figure 2. Distribution of topographic domes with relation to rift structure in East Africa (after Ebinger et al. 1989). Western branch volcanic centers: R = Rungwe, T= Toro-Ankole, V= Virunga, K = South Kivu.

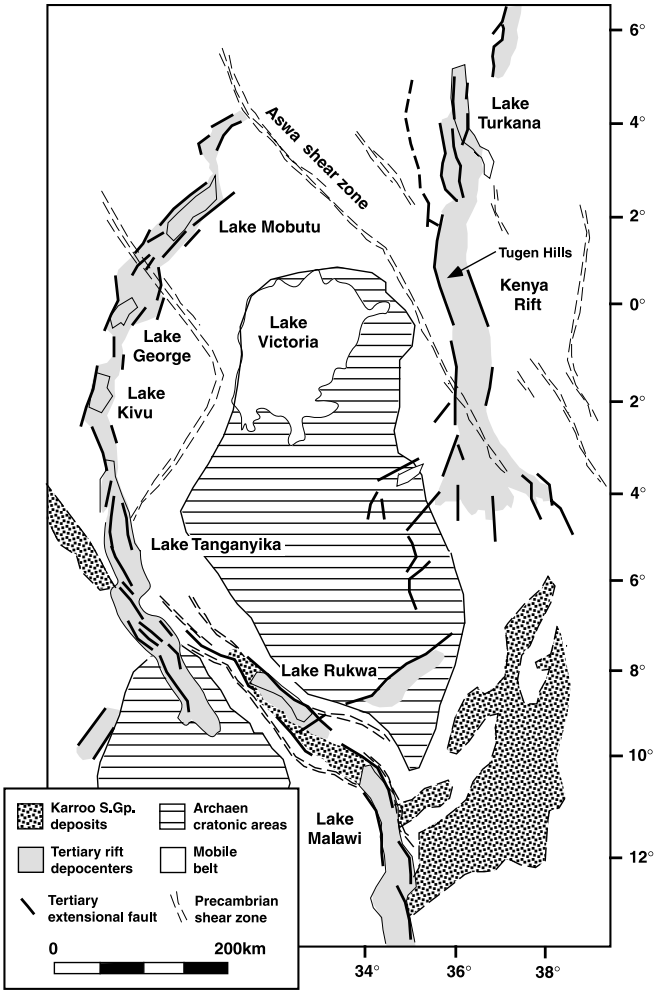


Figure 3. Location map for the East African Rift System (after Morley 1995).

may have accommodated over 10 km of extension (Hendrie et al. 1994). It took seismic reflection data to establish the presence of the fault (Morley et al. 1992).

Recent studies have helped unify some of the more fragmented aspects of previous research. In particular, the deep seismic refraction profiling program in the Kenya Rift (KRISP) has provided details about the structure of the lower crust and mantle that has important implications for the geodynamics, structure, and igneous evolution of the rift and has helped to pull different geoscience disciplines together (for example, the special volumes of *Tectonophysics* No. 236, 1994 and No. 278, 1997).

MAJOR ASPECTS OF THE EAST AFRICAN RIFT SYSTEM

Rift Basins of the East African Rift System

Seismic reflection surveys acquired during the 1980s revealed the existence of very deep, extensive rift basins in the EARS. The first data came from surveys by Project PROBE over lakes Tanganyika, Malawi, and Turkana (e.g., Rosendahl et al. 1988, Ebinger et al. 1987, Specht and Rosendahl 1989, Dunkleman et al. 1989). Oil exploration in the northern Kenya Rift, Lake Rukwa, Ruzizi Plain, and the Usangu Flats provided information on more basins of the East African Rift System.

The western branch of the rift contains a string of half graben basins that are mostly concealed beneath the major lakes. The basins tend to be characterized by the dominance of large boundary faults which produce asymmetrical half grabens. Long rift segments exhibit an alternation in the sense of "polarity" of the half grabens about every 60–100 km (e.g., Rosendahl et al. 1986, Ebinger et al. 1987). The basins are probably all relatively young (late Miocene-Recent), although the Rukwa Rift is the only one in the western branch whose age is well constrained (Wescott et al. 1991). These basins are filled predominantly by clastic sediments eroded from basement and sedimentary rock source areas, and deposited in fluvio-deltaic and lacustrine environments. Occasionally carbonates were deposited, particularly in shoreline and shallow lacustrine environments (e.g., Cohen and Thouin 1987). The basin fills reach maximum thicknesses of 6–7 km (Morley, 1989).

The rift basins of the eastern branch are more poorly known. Extensive Miocene-Recent lava flows have concealed much of the older rift history. However, in northern Kenya surface and subsurface geology show a history of rifting that extends not only throughout the Miocene, but into the Paleogene as well (Morley et al. 1992, Chapters 2 and 3). A brief review of these basins is provided in Part 2 of this chapter. While the western branch of the rift system provides good examples of relatively simple extensional structural geometries, the much longer-lived northern Kenya Rift shows how extensional systems evolve through time. A complex history of boundary fault propagation, activation, and deactivation occurs in northern Kenya, within an overall easterly migration of tectonic activity with time (Baker and Wohlenberg 1971, Morley et al. 1992). The basins are filled by basement-derived sedimentary rocks, volcaniclastics, and lava flows.

Extension across the western branch is estimated to be a maximum of about 13 km equating to a Beta (β) factor of 1.2–1.3 (Morley 1988, Ebinger 1989, Kusznir et al. 1995). Considerably more extension, up to about 40 km, affected the northern Kenya Rift (KRISP 1991, Morley et al. 1992, Hendrie et al. 1994).

Volcanism

Important volcanism has affected the East African Rift System since its formation. There is, however, a marked difference between the quantity and timing of volcanic emissions between the eastern and western branches. The western branch began volcanic activity in the late Miocene and activity has been restricted to four main areas: Rungwe; Virunga; South Kivu; and Toro-Ankole (see Ebinger 1989 for a review, Figure 2). Volcanic activity began in the eastern branch during the Oligocene and today outcrops of volcanic rocks are very widespread.

In the eastern branch, rifting was accompanied or preceded by voluminous eruptions of basalts, trachytes, and phonolites (Baker et al. 1972, Justin-Visentin et al. 1974) followed, in some places, by rhyolites. Trace element geochemistry suggests that most of the melts under the rift zone have been derived from the mantle, perhaps at depths between 40–80 km (transition from spinel to garnet peridotite, e.g., Latin et al. 1993, Macdonald 1994), while geophysical data indicate that zones of partial melts lie between 60–100 km (e.g., Banks and Beamish 1979, Achauer et al. 1994). The large volumes of basaltic magma (about 924,000 km^3 over the last 30 m.y.) are estimated to have required a source region larger than the volume of the lithosphere under the rift. This led Latin et al. (1993) to infer that the partial melt zone in the lithosphere was fed by extra material derived from a rising asthenospheric mantle plume. Some melting of the crust may also have occurred, but contributed a relatively small amount to the total volume of magma (see Macdonald et al. 1994 for a review). In the Kenya Rift some compositional changes in the extrusives can be attributed to variations in Precambrian basement composition (Smith and Mosley 1993).

From the Pliocene to the Present, magmatic and tectonic activity in the eastern branch has been concentrated within a narrow axial trough (e.g., Suguta Valley, Figure 4). This change in activity is marked by different petrological and geochemical characteristics of the extrusives, which are predominantly transitional series basalts (Barberi and Santacroce 1980). Despite the concentration of volcanic activity within the rift there is also marked Miocene-Quaternary "off-axis" volcanism on the rift flanks, such as Mount Elgon, Mount Kenya, and Mount Kilimanjaro.

It is commonly thought that the earliest volcanism in the Kenya Rift preceded the earliest known rifting (e.g., Baker and Wohlenberg 1971, Morley et al. 1992, Macdonald et al. 1994). Volcanic activity was initiated at younger times progressing southwards. Volcanism commenced around 33–36 Ma. in northern Kenya (Lotikipi area), and around 15 Ma. in the southern part of the rift. In the southern Kenya Rift crustal extension began around 10 Ma., while the splay faults further south were initiated around 5 Ma. The onset of extension in the Lotikipi and Turkana area is not so clear. Previous work has viewed the volcanics in the Lotikipi area as forming prior to the main rifting events (e.g., Baker and Wohlenberg 1971, Morley et al. 1992). However, in Chapter 3 it is argued that the Lotikipi volcanics may have occurred during and after a Paleogene rifting event. This extensional event may be related to the Central African Rift-Anza Graben system, not the East African Rift System—its relationship to the volcanism is unclear. In the central and northern Kenya Rift the onset of volcanism prior to rifting is also in doubt because feldspathic (basement derived) sandstones underlie middle Miocene volcanics and directly overlie Precambrian basement in the Lokichar Basin, the Kito Pass area, and the Tugen Hills (Chapman et al. 1978, Morley et al. 1992). These sandstones are of Paleogene-middle Miocene age in the Lokichar Basin and are clearly synrift deposits (Morley et al. 1992, Chapter 2). The data presently available suggests that faulting preceding volcanism in the northern and central Kenya Rift, south of the Tugen Hills area (Figure 3) the reverse seems to apply and volcanism probably preceded rifting.

Pre-Rift Structure

The East African Rift System is developed on crust that exhibits a relatively simple Phanerozoic history of successive rift and sag basins formed during the Paleozoic, Mesozoic, and early Tertiary. Most important in terms of extent and influence on the EARS geometry is the Precambrian crystalline basement. The basement is formed by a number of relatively stable Archean cratonic areas surrounded by orogenic sutures (orogenic or mobile belts). The good quality of exposures associated with the East African Rift, aided by the absence of a thermal sag basin, enables the interaction between rift structure and basement fabric to be investigated (see Chapters 9 and 11). The region serves as a good analog for the role of fabrics on rift structure. In the following section the history and nature of the largest scale fabric elements are considered.

Precambrian Basement

Of the five Archean cratons that make up southern Africa, two play an important part in the EARS, these are the Zambian and Tanzanian Cratons. Both are composed of para- and orthogneisses, with associated basic and ultrabasic rocks (Anhaeusser et al. 1969). Stabilization appears to have occurred between 3,600 and 3,000 Ma. for the Tanzanian Craton (Spooner et al. 1970, McConnell 1972, Kroner 1977), and at about 1,820 Ma. for the Zambian Craton (Brewer et al. 1979). Crustal thickness of the cold and rigid cratons may reach 50–60 km. The rift structure at the largest scale is controlled by the presence of these cratonic areas, both the western and eastern branches lie within the orogenic or mobile belts that surround the Tanzanian Craton (McConnell 1972, Figure 5).

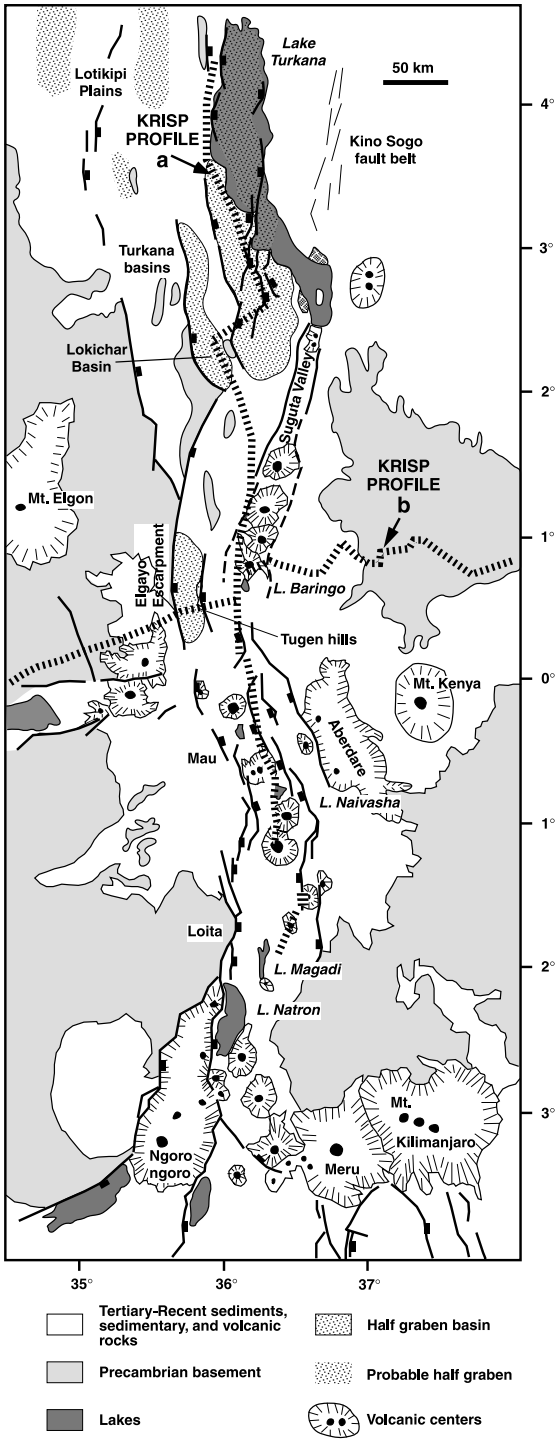
The orogenic belts display three main episodes of deformation. The oldest is the Eburnean (2,100–1,800 Ma.). It is important for creating the approximately 140° striking ductile metamorphic fabric that deflects the rifts from a north-south orientation to an northwest-southeast orientation in the western rift branch between Lakes Tanganyika, Rukwa, and Malawi.

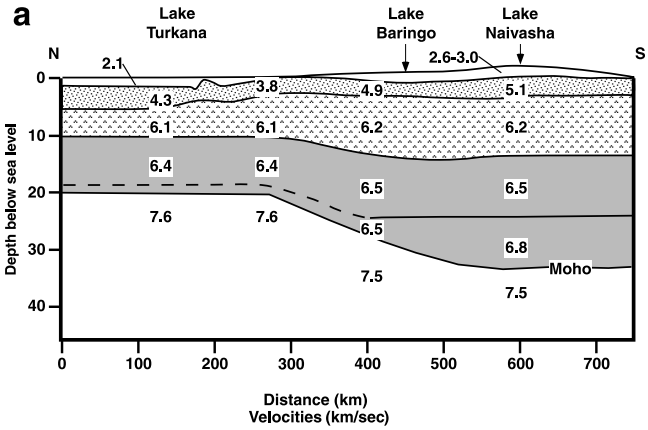
The Kibarian Orogeny (1,400–900 Ma.) forms north to north northeast trending zones of ductile deformation. In particular, one chain (Karagwe-Ankole) forms the basement to the northern Tanganyika-Lake Kivu area.

The Panafrican or Katangian Orogeny of about 600 Ma., created the most widespread orogenic belts including the crystalline basement which underlies the Kenya Rift (e.g., McConnell 1972, Smith and Mosley 1993). In the western branch the orogeny is represented by the Bukoban Belt, which is relatively weakly deformed and includes platform sedimentary rocks, that are exposed on the eastern side of Lake Tanganyika (Holmes 1952, Daly 1984).

Karroo Basins

The Karroo is a term used to identify dominantly continental deposits that occur in a number of separate basins





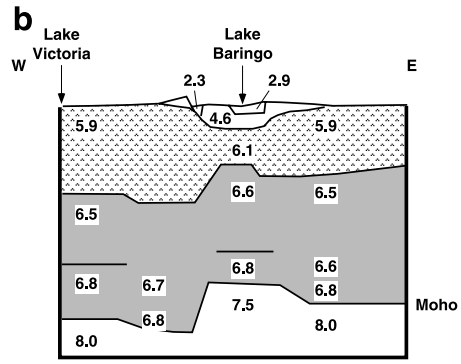


Figure 4. Location map for the Kenya Rift (combining data from Baker 1986, Morley et al. 1992, and Smith and Mosley 1993). Crustal cross sections are based on seismic refraction profiling (after KRISP 1991).

throughout southern Africa (Figure 6). They range in age from Late Paleozoic to the Early Jurassic. In South Africa the largest Karroo basin is a foreland basin associated with the Cape Fold and Thrust Belt. Many of the other more linear Karroo deposits have been interpreted as representing extensional or pull-apart basins (see Kreuser 1995 for a review). They most commonly trend northwest-southeast or northeast-southwest, and less commonly north-south and east-west (Daly et al. 1989). It has been proposed by Daly et al. (1989) that these basins developed in the foreland of the Cape Fold and Thrust Belt under conditions of approximately north-south compression, and east-west extension. The Karroo rifts followed pre-existing lines of weakness in the basement and opened under oblique shear. Several segments of the EARS follow the trends of Karroo basins. Most important for the data examined in this volume is the super-

imposition of the Karroo age and proto-Rukwa Rift and the late Tertiary-Recent age rifts in the vicinity of Lake Rukwa (Figure 3).

Although many of the Karroo outcrops lie in fault bounded basins it does not necessarily mean that all the Karroo sediments were deposited in rifts. Some units may well represent sag basin deposits that were subsequently preserved in grabens and eroded from the flanks. This problem has not been widely investigated. However, it seems clear that the Karroo deposits of northeastern Kenya are widespread and seismic data has revealed they change thickness only gradually and are little affected by faulting—these features point to a sag basin origin.

Jurassic Rifting

The middle Jurassic extensional phase is the most "successful" of the many rifting episodes that affected East Africa, but is very poorly represented onshore. Around Oxfordian time, Madagascar, India, and Antarctica separated from East Africa and oceanic crust was generated (Rabinowitz et al. 1983). The Lamu Embayment in Kenya (Figure 6) is the most likely site for the pre-rift location of Madagascar and coincides with the southeasterly termination of the Anza Graben (Reeves et al. 1987). Consequently, some control on the location and orientation of the Anza Graben might be due to the presence of an older Jurassic triple junction in the vicinity. There appears to be little influence by Jurassic rifting on the other rift basins examined in this volume. The Jurassic rift-related basins seem to be confined to the coastal regions of Kenya, Tanzania, and Mozambique.

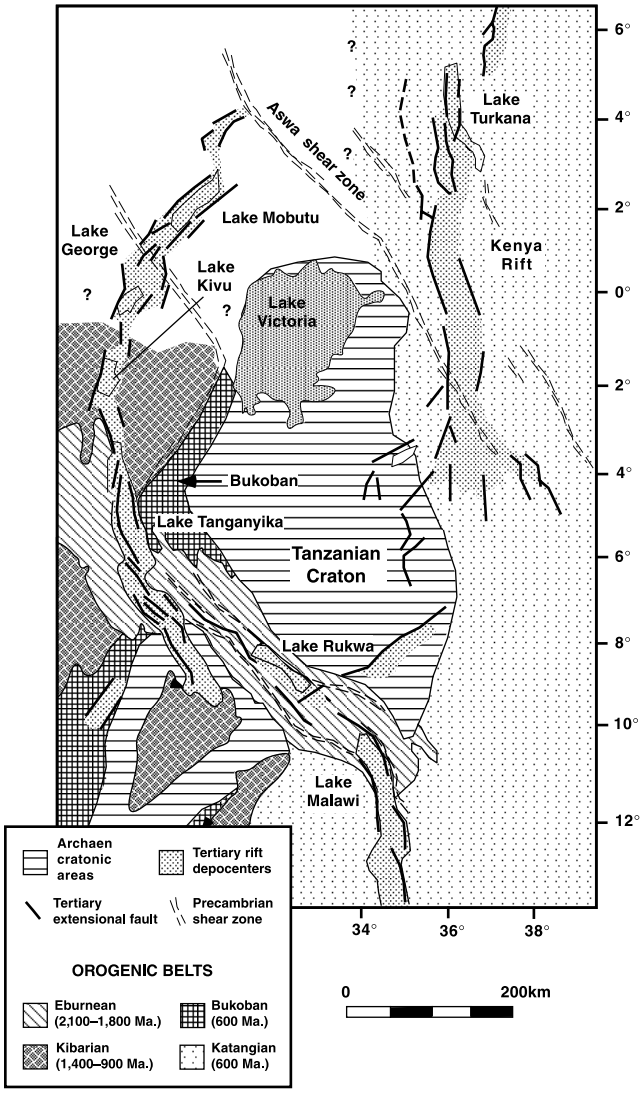


Figure 5. Map showing the main elements of the Precambrian basement terranes (compiled from Mondeguer 1991 and Smith and Mosley 1993).

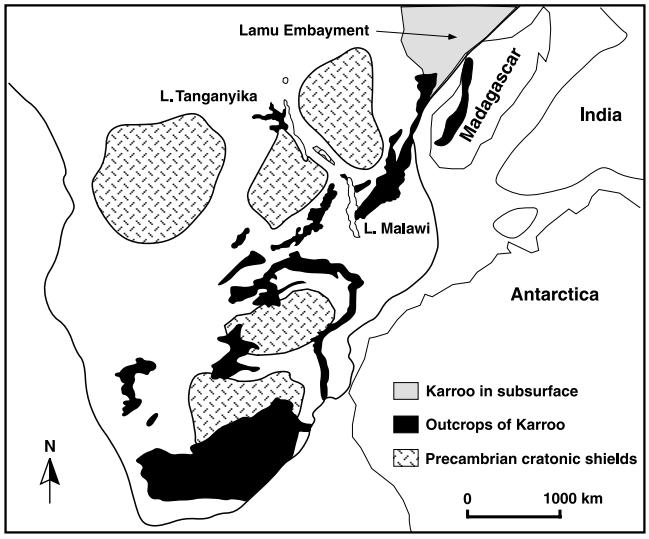


Figure 6. Map of the main Karroo outcrops in central and southern Africa. The pre-Jurassic rifting fit of Madagascar, India, and Antarctica with Africa is also shown (after Daly et al. 1989).

Cretaceous-Paleogene Rifting

An extensive rift system runs across central Africa, from the west coast in Nigeria to the east coast in Kenya. In places it may have a Jurassic history, but in general the rifts formed during the Cretaceous, with episodic active rifting lasting in some areas until the early Tertiary (e.g., Schull 1988, Genik 1992, Bosworth 1992, Bosworth and Morley 1994). This rift system affects the EARS in the northern Kenya area where the northwest trending Anza Graben terminates on the east side of Lake Turkana. The connection of the Anza Graben with the rift in Sudan is poorly known, but such a connection must lie in northern Kenya (Figure 7). Exploration in this area, discussed in Chapters 2 and 3, provides important information on the relationships between the older and younger rift trends. In most places the Cretaceous-Paleogene rifts are located in completely covered, low-lying areas, and are only known from subsurface studies (particularly gravity and seismic surveys and drilling). These rifts form deep basins (6–8 km) that in places have significant accumulations of hydrocarbons (e.g., Schull 1988, Genik 1992). The extension of the productive rift

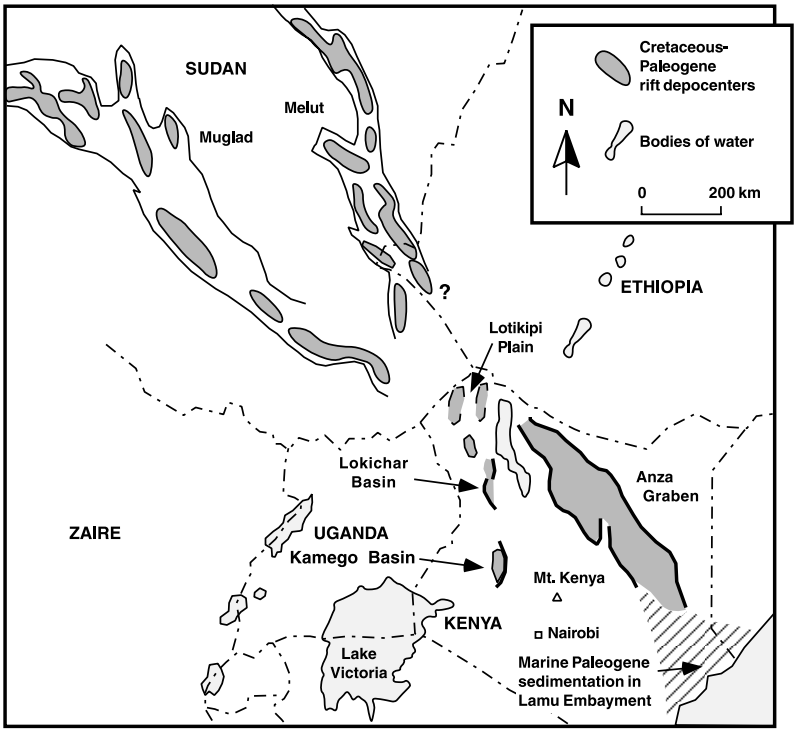


Figure 7. Map of the Cretaceous rift system in East Africa (after Ebinger and Ibrahim 1994 and Hendrie et al. 1994).

trends in Sudan were traced to the Anza Graben which was explored by several oil companies. Subsurface data for the Anza Graben is examined in Chapter 4.

The topographic low in northern Kenya between the East African and Afar Domes (Figure 2) coincides with the trend of Cretaceous-Paleogene rifting and may reflect a regional influence of the older rift trend on the younger one.

DEEP STRUCTURE OF THE RIFT AND TECTONIC MODELS FOR RIFTING

Significant variations in the thickness and structure of the crust in the Kenya Rift has been revealed by seismic refraction and wide-angle reflection surveys conducted by the KRISP working party (1991). The crust thins from about 35–40 km on the flanks to 30 km along the rift axis in the central part of the rift (Figure 4). Passing along the rift axis, the crust thins from about 35 km beneath the Kenya Dome to about 20 km in the Turkana area. A high velocity zone (6.8 km/sec) at the base of the crust has been interpreted as a zone of magmatic underplating (Green et al. 1991, KRISP working party 1991, Figure 4).

The potential zone of magmatic underplating represents material added to the crust during rifting, and implies that the amount of crustal stretching (Beta factor) is higher than the present thickness of crust would indicate (Latin et al. 1993, Macdonald et al. 1994). Macdonald (1994) suggested that the central and southern Kenya Rift may have undergone the same amount of extension as did the Turkana area, but since volcanism is more voluminous in the central and southern Kenya Rift, the crustal thinning was compensated by magmatic underplating. This would mean that progressive shallowing of the Moho to the north does not necessarily indicate an increase in the amount of extension (Figure 4). This conclusion has two flaws. First, there is no evidence for approximately 35–40 km of extension in the upper crust of the central Kenya Rift, as there is for the Turkana area. It does not appear possible for half graben systems, as extensive as those in the Turkana area, to be concealed beneath flows in the central and southern Kenya Rift. The surface rift simply is not wide enough. Nor are large negative gravity anomalies, indicative of buried basins, present. The second flaw is the assumption that volcanism diminishes in volume to the north. Karson and Curtis (1989) made a minimum estimate of the volume of extrusives in the Kenya Rift between 4°N and 2°S of 109,400 km^3. Yet for the Lotikipi Plain area (partly below and partly above 4°N) across to the east side of Lake Turkana, the volume of Oligocene-Miocene lavas was estimated at 450,000 km^3 by Morley et al. (1992). Consequently it is difficult to argue that the volume of lavas decreases to the north.

Despite the plausibility of the arguments for significant crustal inflation by magmatic intrusion there does not seem to be a major discrepancy between upper and lower crustal thinning estimates in the Kenya Rift. In fact, estimates of upper crustal extension fit well with extension estimates using the KRISP data. For both brittle and ductile extension in the Turkana area extension is estimated at about 40 km (Hendrie et al. 1994), while about 10 km is estimated for the central Kenya Rift. Hence it is difficult to explain why there appears to be a good correlation between upper and lower crustal thinning when magmatic underplating and

intrusion should result in a relative underestimate of lower crustal thinning.

This problem remains unresolved, and has important implications. One, or a combination, of the following factors could be involved:

1. the extension estimates are wrong;
2. assumptions about the relative volumes of intruded to extruded igneous rocks are incorrect;
3. a significant volume of crustal material is lost into the mantle when magma is intruded (Morley 1994); and
4. the simple shear/pure extension model for the Turkana area is incorrect.

Seismic refraction data and low rift topography suggest considerable ductile flow of the lower crust (Mechie et al. 1994). If ductile flow can move material laterally along the rift, or towards the rift flanks, then the material added during magmatic underplating may be widely redistributed by nonplanar ductile strain. A more detailed deep-crustal refraction or reflection survey is needed in the Turkana area to help resolve this problem.

Considerable debate has centered around whether the tectonic evolution of the East African Rift is active or passive (e.g., Morgan and Baker 1983). In the passive model, rifting is driven by regional stresses (the forces created by the processes of plate tectonics, such as subduction and ridge push) and the athenospheric mantle rises passively, controlled by the amount and geometry of extension in the lithosphere (e.g., McKenzie 1978). In the active rift model, the rise of the asthenospheric mantle is attributed to thermal plumes and is independent of extension in the lithosphere—the upwelling asthenosphere is the driving mechanism for rifting.

Many geophysical studies of the East African Rift System have been directed at trying to constrain the active and passive rifting models. There is strong evidence for mantle thermal plumes (active rifting), as indicated by the broad

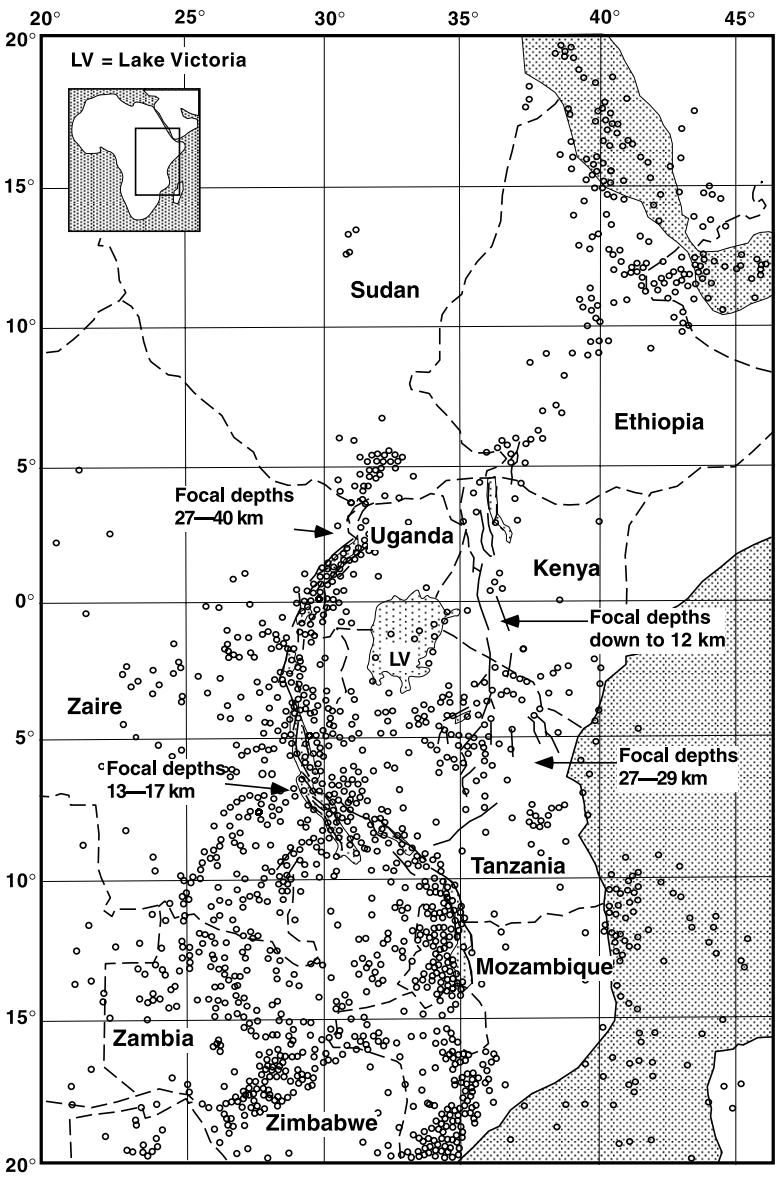


Figure 8. Map of the seismicity of East Africa from events recorded this century (from Nusbaum et al. 1993).

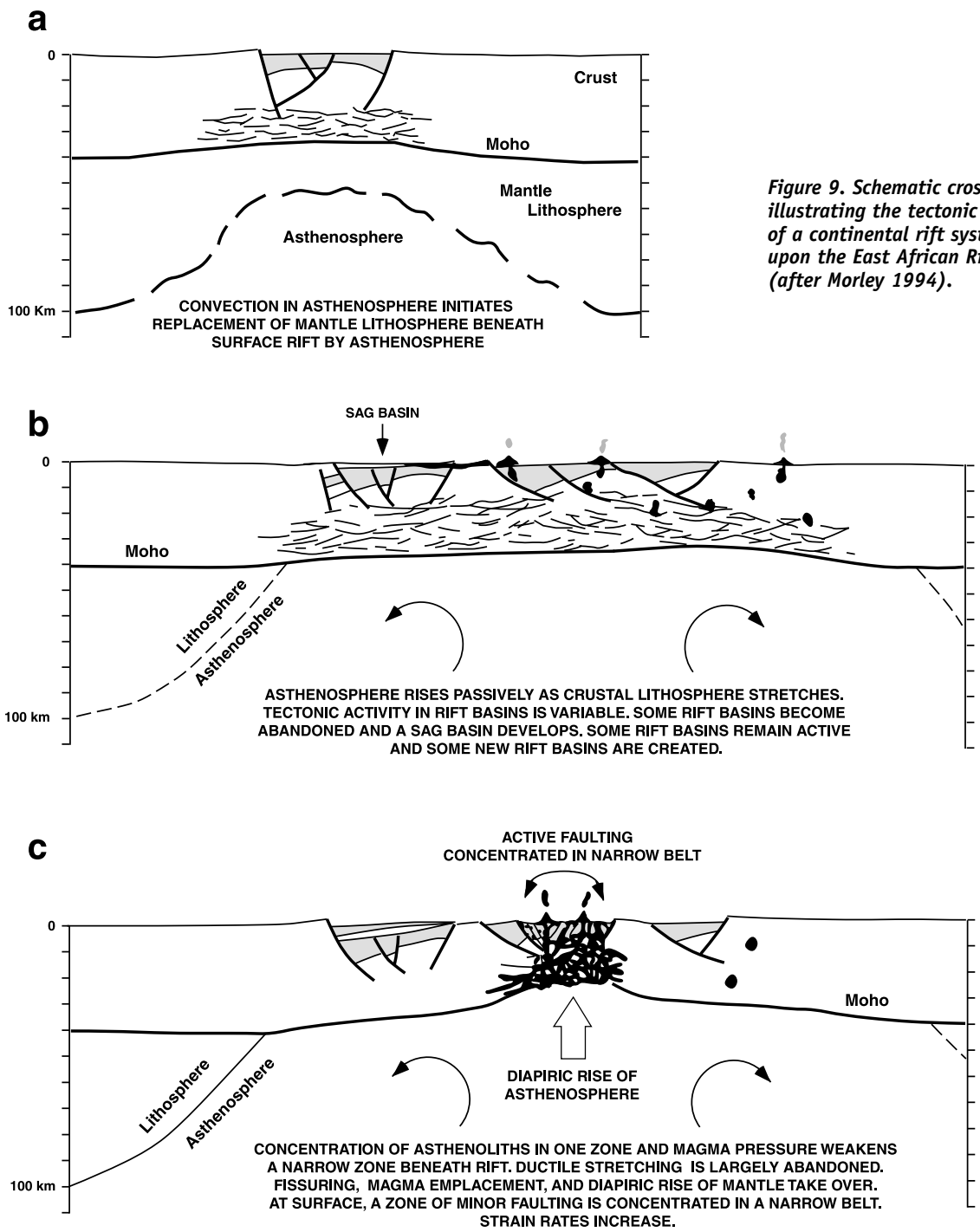


Figure 9. Schematic cross sections illustrating the tectonic development of a continental rift system based upon the East African Rift System (after Morley 1994).

topographic highs of the Afar and East African Domes (Ebinger et al. 1989, Figure 2). The KRISP seismic refraction project, building on earlier studies, identified velocity and density anomalies in the upper mantle underlying the Kenya Rift, interpreted as the presence of partial melts (Davis et al. 1993, Mechie et al. 1994, Keller et al. 1994). A steep sided low-velocity body confined to approximately the width of the surface rift, and extending from the base of the crust to at least 165 km depth has been identified in the southern Kenya Rift (Achauer et al. 1994, Slack and Davis 1994). Consequently, the asthenosphere is interpreted to lie very close to the base of the crust under the Kenya Rift and thermal processes within the mantle are thought to be important for affecting both the igneous and structural evolution of the rift (e.g., Baker and Wohlenberg 1971, Williams 1978, Morley 1994, Smith 1994). It is thought that the

observed crustal thinning of about 10 km is too little to explain the magnitude of mantle lithosphere thinning required to emplace asthenospheric mantle near the base of the crust, hence an active mantle plume has been proposed instead (Davis et al. 1993).

These data do not mean that a passive rifting model can be entirely written off. A passive rift system will tend to locate itself at the weakest regions of the continental lithosphere. If mantle plumes are responsible for the weakening, then a passive rift will be superimposed on active regions of the mantle. One problem with the passive rifting model is that the African Plate is regionally in a state of compression, which led Zoback (1992) to argue that the buoyancy force of low-density upper mantle beneath the Kenya Rift is (locally) more dominant than the ridge-push compression and this produces the present-day northwest-southeast extension. However, there is strong evidence for several significant changes in the paleostress field from the Miocene-Recent (see Bosworth et al. 1992 and Bosworth and Strecker 1997 for reviews, and Chapter 12). Such changes conflict with the long term process envisioned by Zoback (1992). The active rift model also fails to explain why the northwest-southeast oriented minimum compressive stress direction arose in the late Pleistocene, whereas the topographic domes developed much earlier (Bosworth et al. 1992, Bosworth and Strecker 1997).

SEISMICITY IN EAST AFRICA

Figure 8 is a map of earthquake epicenters in East Africa, it shows that the western branch of the rift is considerably more active than the eastern branch (Fairhead and Girdler 1971, Nusbaum et al. 1993). The western branch represents a continental rift in the early stages of development, having a duration of some 10 m.y. (e.g., Ebinger 1989). Strong seismic activity runs from Lake Mobutu in the north to Lake Malawi in the south, also a large area to the west of the rift is seismically active (Fairhead and Henderson 1977). One striking feature of the western branch is that many earthquakes are of large magnitude and deep seated. Between Lakes Tanganyika and Malawi, focal depths tend to be intermediate (13–17 km), (Shudofsky 1985). However, data from Lake Rukwa shows a large deep earthquake with a focal depth of 25 km (Forster et al. 1995) in this zone with a northeast-southwest to east northeast-west southwest minimum compressive stress direction. Seismicity in the Rukwa area ranges between depths of 6–30 km (Camelbeeck and Iranga 1996) and is associated with the boundary fault margin of the rift, with motion on planes dipping between 60° and 70°. Deep seismicity between 29–40 km affects the extremities of the rift branches and the Ruwenzori Horst area (Zana and Hamaguchi 1978, Shudofsky 1985, Shudofsky et al. 1987). The localization of deep earthquakes to the extremities of the rift branches suggests that the early stages of rift development in cold, brittle continental crust are characterized by deep faults that can traverse much of the crust (Morley 1989, Girdler and McConnell 1994).

The eastern branch is seismically relatively quiet (Figure 8). Nevertheless, numerous small earthquakes affect it (e.g., Tongue et al. 1994). Data from the central Kenya Rift (Tongue et al. 1994) shows a clustering of events beneath the central trough of the Kenya Rift at depths between 2–7 km, with the brittle-ductile transition zone lying at about 12–16 km. Sub-vertical fault plane mechanisms are thought to represent failure on dike-country rock interfaces and dike propagation, with minimum compressive stress directions ranging between east-west and northwest-southeast (Young et al. 1991, Tongue et al. 1994). The Kenya Rift is at a relatively advanced stage of structural evolution in contrast to the western branch. The early half grabens with large bounding normal faults have been abandoned in favor of a narrow zone of minor faults developed above a relatively high brittle-ductile transition zone. Young volcanic activity is concentrated along this central zone and sub-surface dike emplacement appears to be an important mechanism for accommodating extension.

A schematic model for the evolution of the rift system is illustrated in Figure 9, which combines the observations made from seismic refraction, seismological, seismic reflection, and surface geological studies. The asthenosphere rises underneath the rift by both active and passive mechanisms triggering partial melting of the upper mantle and the emplacement of magma bodies within the crust. Rifting in the crust begins by the formation of high-angle deep-penetrating, large normal faults, leading to the development of large half grabens, such as those found in the western branch of the EARS. As rifting progresses, the brittle-ductile transition zone within the crust rises and earthquake epicenters tend to rise. Somewhat lower-angle boundary faults may develop. Some rift basins become abandoned and new ones created. As extension progresses, a relatively narrow belt of igneous intrusions may form within the crust. Extension and igneous activity may become progressively more concentrated in this zone resulting in dike and minor fault swarms taking over accommodation of extension from the large half graben boundary fault systems. This evolutionary sequence could lead to oceanic rifting (Mohr 1982).

SUMMARY

The East African Rift System is formed primarily in the zones of Precambrian orogenic belts, and avoids the stable Archean cratonic areas. The most profound influence of pre-existing fabrics is exerted by Precambrian ductile fabrics. However, late predominantly extensional events during the Permo-Triassic, Jurassic, Cretaceous, and Paleogene have also influenced the location, orientation, and geometry of the Tertiary rift systems to some extent.

The two branches of the EARS have undergone different tectonic histories. In general, the western branch can be regarded as a good model for a young continental rift while the eastern branch is more representative of a mature continental rift system that has failed to produce oceanic crust. In comparison with the eastern branch, the western branch is younger (late Miocene-Recent), has only scattered, isolated areas of volcanic activity, and is more seismically active with deeper earthquakes (down to about 30–40 km).

The eastern branch was probably initiated in the Eocene and is at a structurally more advanced stage than the eastern branch. In the Kenya and Ethiopian Rifts, older half grabens have been abandoned in favor of a narrow rift zone characterized by minor fault swarms, dike intrusions, and volcanic centers. There is evidence for an important active rift component (very thin mantle lithosphere, large topographic domes), as well as the effects of passive rifting.

Recent research has shed considerable new light on the geological evolution and structure of the EARS and older rift basins, yet many problems remain. Some of these problems are as follows:

1. The complete paleostress evolution of the rift is not fully understood, although multiple stress changes have been suspected or recorded. The driving mechanisms for the stress changes remain a cause for debate.
2. The role of active and passive rift mechanisms and their impact on the generation of magmas is still an area of active investigation.
3. There is the problem concerning the amount of extension undergone by the rift system. Current estimates of upper crustal extension appear to tie well with the deep crustal extension estimates from seismic refraction data. However, the lower crust appears to contain a significant volume of intruded material, that perhaps should have inflated the crustal volume during extension. Consequently, the lower crust should actually display less extension than the upper crust. This issue needs considerably more investigation. More details about the deep structure of the crust would be very beneficial.
4. A comparison between deep profiles across the western and eastern branches would help our understanding of how rifts evolve, at present only data from the Kenya Rift is available.
5. The presence or absence of old, deep basins in the Ethiopian and southern Kenya Rifts needs to be further investigated. The Turkana area has provided some examples of the older half graben history of the rift, but large areas of the eastern branch have very little subsurface control.

Part 2—Rift Basins in the Kenya Tertiary Rift

INTRODUCTION

The Kenya Rift lies in the eastern branch of the East African Rift System (EARS). Major extensional faults in the Kenya Rift define separate sub-basins with distinctive structural and geological settings (e.g., Chapman et al. 1978, Morley et al. 1992). Different basin terminologies within different companies has resulted in multiple names for some basins as follows: Turkana (offshore) or Turkana Basin; Lokichar or Lodwar South; North Lokichar or Lodwar North; Kerio Basin or Kerio South; Turkana Basin (onshore) or Kerio North; and the Elgayo-Tugen Hills Basin or South Kerio Trough (Figure 10). Most of these basins are half grabens bounded on one side by major boundary faults (Figures 11–14). Sedimentation in many of these basins was episodically interrupted by volcanic pulses (Figure 15). Extensional activity in several of the basins was also episodic and significantly different timing of deformation can be identified even in adjacent basins.

A brief summary of the tectonic history and stratigraphy of the known half graben basins as determined by oil exploration activity to date is presented in Part 2 of this chapter.

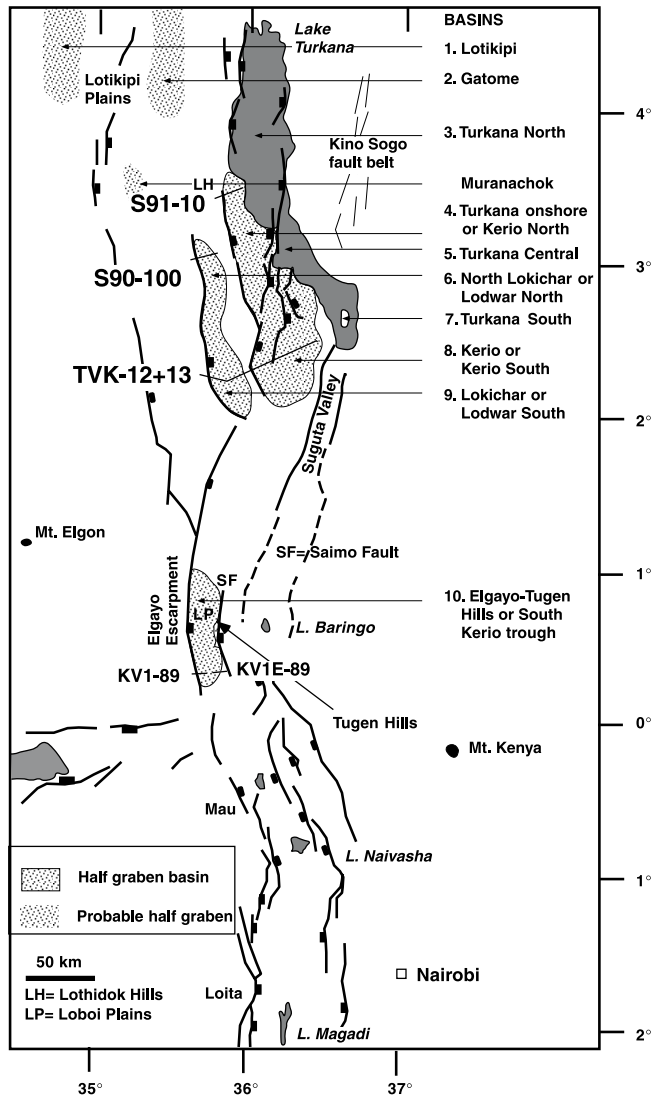


Figure 10. A map of the Tertiary Kenya Rift showing major faults, seismic lines, and basins. The basins shown above are: 1. Lotikipi; 2. Gatome; 3. Turkana North; 4. Turkana Central; 5. Turkana (onshore) or Kerio north; 6. North Lokichar or Lodwar North; 7. Lokichar or Lodwar South; 8. Kerio Basin or Kerio South; 9. Turkana South; and 10. Elgayo-Tugen Hills or South Kerio Trough.

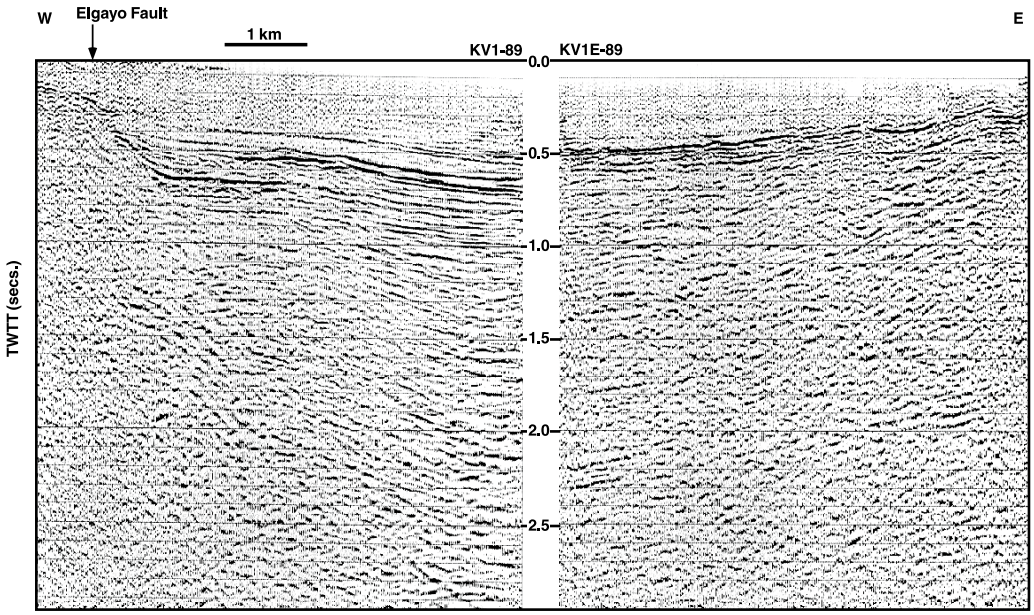
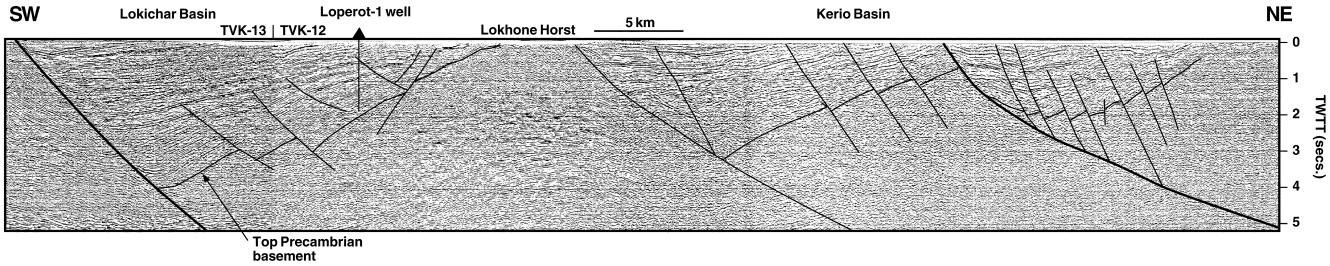


Figure 11. Seismic line across the Elgayo-Tugen Hills, or South Kerio Trough, KV1-89 and KV1E-89 (acquired by NOCK, see Figure 10 for location).



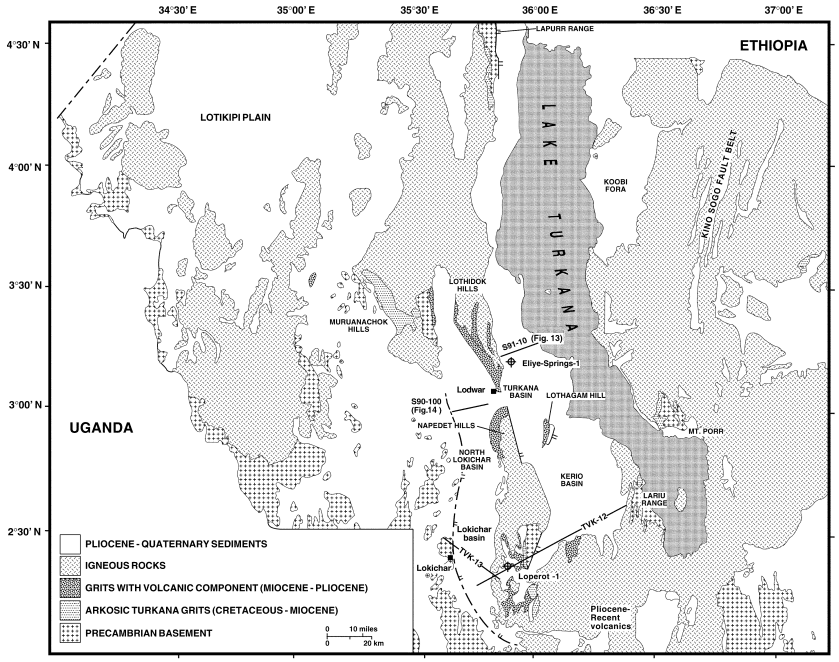


Figure 12. Regional composite of seismic lines TVK-13 and TVK-12 across the Kerio and Lokichar Basins (data acquired by Amoco).

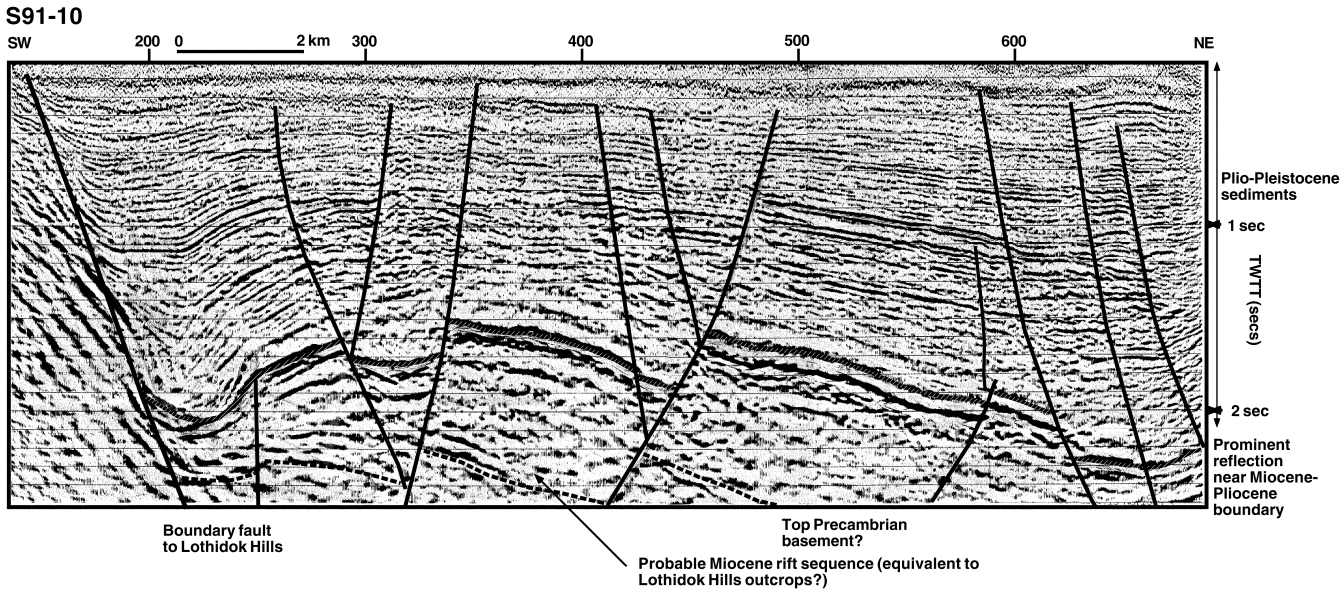


Figure 13. Seismic line S91-10 across Turkana (onshore) basin (data acquired by Shell, see Figures 10 and 12 for location).

SOUTH KERIO TROUGH OR ELGAYO-TUGEN HILLS BASIN

The South Kerio Trough covers the Kerio Valley and the Loboi Plains in the southeastern area of the central Kenya Rift (Figure 10). Seismic data acquired over the area by the National Oil Corporation of Kenya (NOCK) in 1990 has been processed and interpreted (Pope 1992 and Ngenoh 1993, Figure 11). There are no significant structures seen on the processed lines, although the seismic data does show a large boundary fault, the Elgayo Fault, which marks the western boundary of the trough (Figure 11). This fault curves towards the basin depocenter as defined by gravity data (BEICIP 1987). The seismic reflection data could not penetrate the thick volcanic cover in the shallow subsurface.

No wells have been drilled in the basin, but the stratigraphy is known from exposures in the Tugen Hills, which lie towards the flexural margin (e.g., Chapman et al. 1978). The flexural margin is terminated by the Tugen Hills and Saimo Faults which bound a basin to the east. A section from Precambrian basement up to the Pliocene has been mapped (Chapman et al. 1978). Much of the basin fill is comprised of middle Miocene-Pliocene volcanic flows and volcaniclastics. However, some formations contain coarse arkosic clastics, including the Tambach Formation arkosic sands (Figure 15) which are moderately-well sorted, fine- to medium-

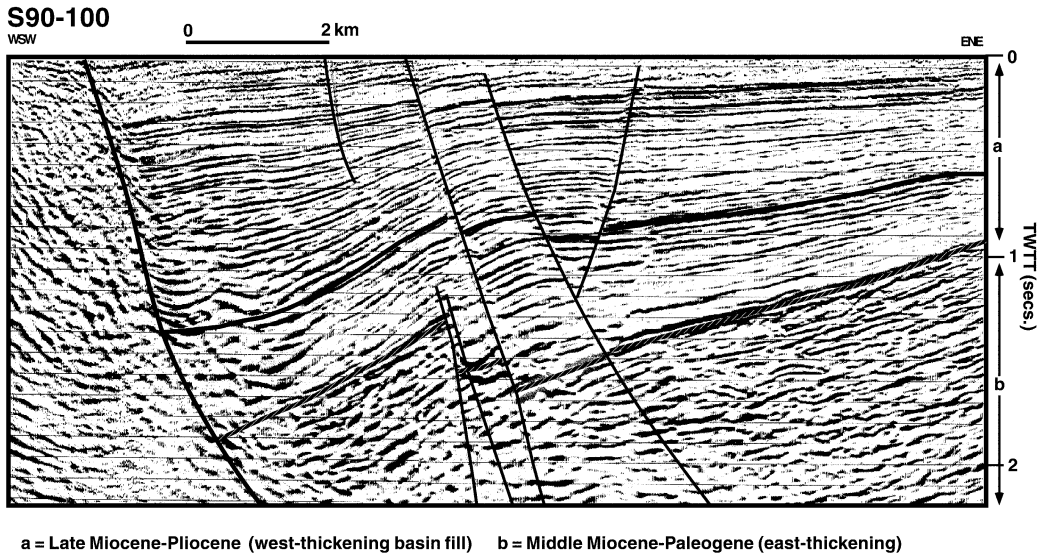


Figure 14. Seismic line S90-100 across the North Lokichar Basin (data acquired by Shell, see Figures 10 and 12 for location).

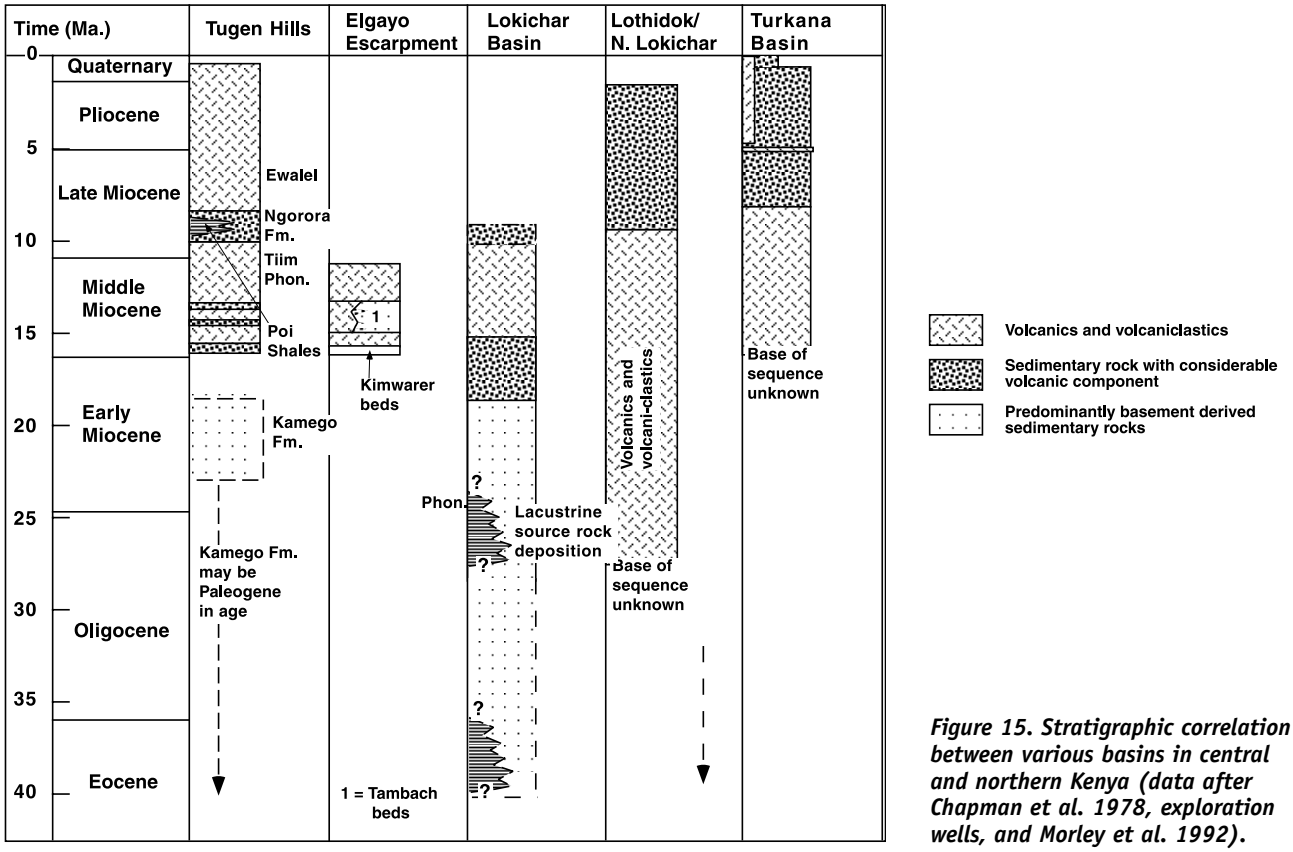


Figure 15. Stratigraphic correlation between various basins in central and northern Kenya (data after Chapman et al. 1978, exploration wells, and Morley et al. 1992).

grained, and cemented by clays (illite and smectites) and calcite. Primary porosity has been reduced by these cements from about 35% to 10% (Ego 1994).

Miocene syn-rift sedimentary deposits in the Tambach and Ngorora Formations contain good lacustrine source rocks. Outcrop samples from these formations (the Kapkiamo and Poi sub-basins within the Tugen Hills) yielded average TOC values of 3% in some intervals (NOCK unpublished report, 1989), of good to excellent oil-prone, Type-1 source rocks of lacustrine-algal origin. The Ngorora Formation was deposited on the Tiim Phonolites and capped by the Ewalel Phonolite (Bishop and Chapman 1970, Figure 15). It's stratigraphy has been extensively studied and displays a variety of fluvio-deltaic and lacustrine environments including tuffaceous sandstone filled channels, stunted stromatolitic carbonates, and fish-bearing paper shales (e.g., Bishop and Chapman 1970, Bishop and Pickford 1975). The latter weather to a white color, but fresh samples are high in TOC.

LOKICHAR OR LODWAR SOUTH BASIN

This basin lies in the southwest part of the Turkana region (Figures 10 and 12). Palynological data, from exposed sedimentary rocks and drilling, indicate that the deeper part of the basin is Eocene and early Miocene in age. Subsidence in the basin was controlled throughout its history by an east-dipping master fault with a maximum throw of about 7 km.

Basin fill was predominantly of fluvial and fluvio-lacustrine sediments with two marked pulses of lacustrine development. The lacustrine shales have good to excellent source rock potential and are of Types I and II. The porosities of potential reservoir sandstones decrease with depth from about 17% to approximately 10% (unpublished report by Shell E&P, Kenya B.V. 1993). Diagenetic pore filling clays (smectites and chlorite) and zeolites reduce the porosity and permeability—a common occurrence in volcanic-derived clastic sediments.

A regional cross section (composite of seismic lines TVK-13 and TVK-12) which crosses the Lokichar and Kerio Basins is shown in Figure 12. It shows two half grabens separated by a basement high, which in outcrop is called the Lokhone Horst. The density of secondary normal faults is higher in the Kerio Basin than the Lokichar Basin. Most secondary faults appear to penetrate basement.

TURKANA (ONSHORE) OR KERIO NORTH BASIN

The Kerio North Basin is the northern-most basin to be explored by drilling (Figures 10, 12, and 15), where the Eliye Springs-1 well was drilled to a depth of 2,970 m (see Chapter 3). The stratigraphy of the basin is not well constrained due to the poor recovery of palynofloras from the well; mainly red beds barren of palynomorphs were penetrated. However, radiometric dating of volcanics and a few

zones of palynoflora recovery suggest a late Miocene to early Pliocene age for rocks at total depth in the well. Fluvial sedimentation appears to have dominated the section. The sandstones in the well are predominantly composed of quartz, feldspar, and minor quantities of mica and heavy minerals.

The structural style of the basin is characterized by high angle, dominantly normal faults (Figure 13). Initially, a master fault lying on the eastern margin of the basin controlled sedimentation. Later development was partially controlled by an east-dipping fault.

NORTH LOKICHAR OR LODWAR NORTH BASIN

The North Lokichar Basin (Figure 10) is poorly known due to limited seismic coverage and poor outcrop data. A large high-angle boundary fault marks the western side of the basin (Figure 14) and gives the basin a half graben geometry. It is bounded to the east (flexural margin) by Miocene igneous complexes that form the Napedet Hills. Sedimentary deposition largely post-dates the middle Miocene volcanics, unlike the Lokichar Basin to the south, where most of the sedimentation occurred prior to volcanic activity (Figure 15).

REFERENCES CITED (Part 1)

Achauer, U., and the KRISP Teleseismic Working Group, 1994, New ideas on the Kenya Rift based on the inversion of the combined data set of the 1985 and 1989–1990 seismic tomography experiments: Tectonophysics, v. 235, p. 305–329.

Anhaeusser, C.R., R. Mason, M.J. Viljoen, and R.P. Viljoen, 1969, A reappraisal of some aspects of Precambrian shield geology: Geological Society of America Bulletin, v. 80, p. 2175–2200.

Baker, B.H., 1986, Tectonics and volcanism of the southern Kenya Rift Valley and its influence on rift sedimentation, in L.E. Frostick, R.W. Renaut, I.Reid and J.-J. Tiercelin, eds., Sedimentation in the African rifts, Geological Society of London Special Publication, v. 25, p. 45–57.

Baker, B. H. and J. Wohlenberg, 1971, Structure and evolution of the Kenya Rift Valley: Nature, v. 229, p. 538-542.

Baker, B.H., P.A. Mohr, and L.A.J. Williams, 1972, Geology of the Eastern Rift System of Africa: Geological Society of America Special Paper, No. 136, 67p.

Banks, R.J. and D. Beamish, 1979, Melting in the crust and upper mantle beneath the Kenya Rift: evidence from Geomagnetic deep sounding experiments: Journal of the Geological Society of London, v. 136, p. 293–311.

Barberi, F., and R. Santacroce, 1980, The Afar stratoid series and magmatic evolution of East African Rift System: Bulletin Society Geologique France, v. 7, p. 891–899.

Bosworth, W., 1992, Mesozoic and early Tertiary rift tectonic in East Africa: Tectonophysics, v. 209, p. 115–137.

Bosworth, W., and A. Maurin, 1993, Structure, geochronology and tectonic significance of the northern Suguta Valley (Gregory Rift), Kenya: Journal of the Geological Society, London, v. 150, p. 751–762.

Bosworth, W., and C.K. Morley, 1994, Structural and stratigraphic evolution of the Anza Rift, Kenya: Tectonophysics, 236, 93–115.

Bosworth, W., and M.R. Strecker, 1997, Stress field changes in the Afro-Arabian Rift System during the Miocene to Recent period: Tectonophysics, v. 278, p. 47–62.

Brewer, M.S., H.W. Haslam, P.F.P. Darbyshire, and A.E. Davis, 1979, Rb/Sr age determinations in the Bangweulu Block, Luapula Province, Zambia: Institute of Geological Science London, v. 79, 11p.

Camelbeeck, T. and M.D. Iranga, 1996, Deep crustal earthquakes and active faults along the Rukwa trough, eastern Africa: Geophysical Journal International, v. 124, p. 612–630.

Chapman, G.R., S. Lippard, and J. E. Martyn, 1978, The stratigraphy and structure of the Kamasia Range, Kenya Rift valley: Journal of the Geological Society of London, v.135, p.265–281.

Cohen, A.S., and C. Thouin, 1987, Nearshore carbonate deposits in Lake Tanganyika: Geology, v. 15, p. 414–418.

Coussement, C., 1995, Structures transverses et extension intracontinentale. Le role des zones de failles díAssoua et Tanganyika-Rukwa-Malawi dans la cinematique neogene du systeme de Rift Est-Africain, Ph.D. thesis, Universite de Bretange Occidentale: Brest, France, 222 p.

Crossley, R., 1979, The Cenozoic stratigraphy and structure of the western part of the rift valley in southern Kenya: Journal of the Geological Society of London, v.136, p.393–405.

Daly, M.C., 1984, Preliminary report on the foreland structure of the Irumide Belt of Northern Zambia, in J. Klerkx and A. Michot, eds., African Geology: Musee Royal Afrique Central, Tervuren, p. 67–74.

Daly, M.C., J. Chorowicz, and D. Fairhead, 1989. Rift basin evolution in Africa: the influence of reactivated steep basement shear zones, in M.A. Cooper and G.D. Williams (eds), Inversion Tectonics: Geological Society of London Special Publication, 44, 309–334.

Davis, P., P. Slack, H.A. Dahlheim, W.V. Green, R.P. Meyer, U. Achauer, A. Glahn, and M. Granet, 1993, Teleseismic tomography of continental rift zones, in H.M. Iyer and K. Hirahara, eds., Seismic Tomography: Theory and Practice: Chapman and Hall, London, p. 397–439.

Dunkleman, T.J., B.R. Rosendahl, and J.A. Karson, 1989, Structure and stratigraphy of the Turkana Rift from seismic reflection data: Journal of African Earth Sciences, 8, p. 489–510.

Ebinger, C.J., 1989, Tectonic development of the western branch of the East African Rift System: Bulletin of the Geological Society of America, v. 101, p. 885–903.

Ebinger, C.J., and A. Ibrahim, 1994, Multiple episodes of rifting in Central and East Africa: A re-evaluation of gravity data: Geologische Rundschau, v. 83, p. 689–702.

Ebinger, C.J., B.R. Rosendahl, and D.J. Reynolds, 1987, Tectonic model of the Malawi Rift, Africa: Tectonophysics, v. 141, p. 215–235.

Ebinger, C.J., T.D. Bechtel, D.W. Forsyth, and C.O. Bowen, 1989, Effective elastic thickness beneath the East African and Afar Plateaux and dynamic compensation of

the uplifts: Journal of Geophysical Research, v. 94, p. 2883–2901.

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: Journal of the Geological Society of London, v. 150, p. 99–108.

Fairhead, J.D. and R.W. Girdler, 1971, The seismicity of Africa: Geophysical Journal of the Royal Astronomical Society, v. 24, p. 271–301.

Fairhead, J.D. and N.B. Henderson, 1977, The seismicity of southern Africa and incipient rifting: Tectonophysics, v. 41, p. 19–26.

Forster, A.N., J.A. Jackson, and C.J. Ebinger, 1995, Source mechanisms and centroid depths of recent African earthquakes (abstract): Terra abstracts, Terra Nova, v. 7, p. 35.

Genik, G.J., 1992, Regional framework, structural and petroleum aspects of rift basins in Niger, Chad, and the Central African Republic (C.A.R.): Bulletin de la Centre du Recherches Exploration Production Elf Aquitaine, v. 13, p. 39–53.

Girdler, R.W., and D.A. McConnell, 1994, The 1990 to 1991 Sudan earthquake sequence and the extent of the East African Rift System: Science, v. 264, p. 67–70.

Green, W.V., U. Achauer, and R.P. Meyer, 1991, A three-dimensional seismic image of the crust and upper mantle beneath the Kenya Rift: Nature, v. 354, p. 199–203.

Gregory, J.W., 1896, The Great Rift Valley: London, John Murray, 424pp.

Gregory, J.W., 1921, Rift Valleys and the Geology of East Africa: London, Seeley Service, 479pp.

Griffiths, P.S., 1980, Box-fault systems and ramps: a typical association of structures from the eastern shoulder of the Kenya Rift: Geological Magazine, v. 117, p. 579–586.

Grimaud, P., J.-P. Richert, J. Rolet, J.-J. Tiercelin, J.-P. Xavier, C.K. Morley, C. Coussement, S.W. Karanja, R.W. Renaut, G. Guerin, C. Le Turdu, and G. Michel-Noel, 1994, Fault geometry and extension mechanisms in the central Kenya Rift, East Africa: A 3D remote sensing approach: Bulletin de la Centre de Recherches Exploration-Production Elf Aquitaine, v. 18, p. 59–92.

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 area: Tectonophysics, v. 236, p. 409–438.

Holmes, A., 1952, The sequence of Precambrian orogenic belts in south and central Africa: 18th International Geological Congress, London 1984, v. 14, p. 254–269.

Justin-Visenti, E., M. Nicoletti, L. Tolomeo, and B. Zanettin, 1974, Miocene and Pliocene volcanic rocks of the Addis Ababa-Debra Berhan area, Geo-petrographic and radiometric study: Bulletin of Volcanology, v. 38, p. 237–253.

Karson, J.A., and P.C. Curtis, 1989, Tectonic and magmatic processes in the eastern branch of the East African Rift and implications for magmatically active continental rifts: Journal of African Earth Science, v. 8 , p. 431–453.

Keller, G.R., J. Mechie, L.W. Braile, W.D. Mooney, and C. Prodehl, 1994, Seismic structure of the uppermost mantle beneath the Kenya Rift: Tectonophysics, v. 236, p. 201–216.

Kreuser, T., 1995, Rift to drift evolution in Permian-Jurassic basins of East Africa, in J.J. Lambiase, ed., Hydrocarbon habitat in rift basins, Geological Society Special Publication, No. 80, p. 297–315.

KRISP Working Group, 1991, The Kenya Rift: pure shear extension above a mantle plume: Nature, v. 345, p. 223–227.

Kroner, A., 1977, The Precambrian geotectonic evolution of Africa: plate accretion vs. plate destruction: Precambrian Research, v. 4, p. 163–213.

Kusznir, N., A. Roberts and C.K. Morley, 1995, Forward and reverse modeling of rift basin formation, in J.J. Lambiase, ed., Hydrocarbon habitat in rift basins, Special Publication of the Geological Society of London, v. 80 p. 33–56.

Latin, D., M.J. Norry, and R.J.E. Tarzey, 1993, Magmatism in the Gregory Rift , East Africa: Evidence for melt generation by a plume: Journal of Petrology, v. 34, p. 1007–1027.

Macdonald, R., 1994, Petrological evidence regarding the evolution of the Kenya Rift valley: Tectonophysics, v. 236, p. 373–390

Macdonald, R., L.A.J. Williams, and I.G. Gass, 1994, Tectonomagmatic evolution of the Kenya Rift valley: some geological perspectives: Journal of the Geological Society of London, v. 151, p. 879–888.

McConnell, R.B. 1972. Geological development of the rift system of eastern Africa: Geological Society of America Bulletin, 83, 2549–2572.

McKenzie, D., 1978, Some remarks on the development of sedimentary basins: Earth and Planetary Science Letters, v. 40, p. 25–32.

Mechie, J., K. Fuchs, and R. Altherr, 1994, The relationship between seismic velocity, mineral composition and temperature and pressure in the upper mantle—with application to the Kenya Rift and its eastern flank: Tectonophysics, v. 236, p. 453–464.

Mohr, P.A., 1982, Musing on continental rifts, in G. Palmason, ed., Continental and oceanic rifts, Geodynamics Series: Washington D. C., American Geophysical Union, no. 8, p. 293–309.

Mondeguer, A., 1991, Bassins sedimentaries en contexte extensif et decrochant: líexemple du ìcomplexe des Fosses Sud-Tanganyikaî, Rift Est-Africain. Morphostructures et sedimentation, Ph.D. Thesis, University of Bretagne Occidentale: Brest, 258 pp.

Morgan, P., and B.H. Baker, 1983, Introduction-processes of continental rifting: Tectonophysics, v. 94, p. 1–10.

Morley, C.K., 1988, Variable extension in Lake Tanganyika: Tectonics, v. 7, p. 785–801.

Morley, C.K., 1989, Extension, detachments, and sedimentation in continental rifts (with particular reference to East Africa): Tectonics, v. 8, p. 1175–1192.

Morley, C.K., 1994, Interaction of deep and shallow processes in the evolution of the Kenya Rift: Tectonophysics, v. 236, p. 81–91.

Morley, C.K., 1995, Developments in the structural geology of rifts over the last decade and their impact on hydrocarbon exploration: Geological Society of London Special

Publication, v. 80, p. 1–32.
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 Kenya Rift: Journal of the Geological Society of London, v. 149, p. 333–348.
Nusbaum, R.L., R.W. Girdler, J.R. Heirtzler, D.J. Hutt, D. Green, V.E. Millings, B.S. Schmoll, and J. Shapiro, 1993, The distribution of earthquakes and volcanoes along the East African Rift System: Episodes, v. 16, p. 427–432.
Rabinowitz, P.D., M.F. Coffin, D. Falvey, 1983, The separation of Madagascar and Africa: Science, v. 220, p. 67–69.
Reeves, C.V., F.M. Karanja, and I.N. MacLeod, 1987, Geophysical evidence for a Jurassic triple-junction in Kenya: Earth and Planetary Science Letters, v. 81, p. 299–311.
Rosendahl, B.R., D.J. Reynolds, P.M. Lorber, C.F. Burgess, J. McGill, D. Scott, J.J. Lambiase, and S.J. Dersken, 1986, Structural expressions of rifting: lessons from Lake Tanganyika, Africa, in Sedimentation in the African Rifts , L.E. Frostick, R.W. Renault, I. Reid, and J.-J. Tiercelin eds., Geological Society of London Special Publication, v. 25, p. 29–43.
Rosendahl, B.R., J.W. Versfelt, C.A Scholz, J.E. Buck, and L.D. Wood, 1988, Seismic atlas of Lake Tanganyika, East Africa (Project PROBE Geophysical Atlas Series, folio 1), Duke University: Durham North Carolina.
Schull, T.J., 1988, Rift basins of interior Sudan: petroleum exploration and discovery: AAPG Bulletin, 72, 1128–1142.
Shudofsky, G.N., 1985, Source mechanisms and focal depths of East African earthquakes using Raleigh-wave inversion and body-wave modeling: Geophysical Journal of the Royal Astronomical Society, v. 83, p. 563–614.
Shudofsky, G.N., S. Cloetingh, S. Stein, and R. Wortel, 1987, Unusually deep earthquakes in East Africa: constraints on the thermo-mechanical structure of a continental rift system: Geophysical Research Letters, v. 14, p. 741–744.
Slack, P.D. and P.M. Davis, 1994, Attenuation and velocity of P-waves in the mantle beneath the East African Rift, Kenya: Tectonophysics, v. 236, p. 331–358.
Smith, M., 1994, Stratigraphic and structural constraints on mechanisms of active rifting in the Gregory Rift, Kenya: Tectonophysics, v. 236, p. 3–22.
Smith, M. and P. Mosley, 1993, Crustal heterogeneity and basement influence on the development of the Kenya Rift, East Africa: Tectonics, 12, 591–606.
Specht, T.D., and B.R. Rosendahl, 1989, Architecture of the Lake Malawi Rift, East Africa: Journal of African Earth Science, v. 8, p. 393–413.
Spooner, C.M., J.V. Hepworth, and H.W. Fairbairn, 1970, Whole-rock Rb-Sr isotopic investigation of some East African granulites: Geological Magazine, v. 107, p. 511–522.
Tongue, J., P. Maguire, and P. Burton, 1994, An earthquake study in the Lake Baringo basin of the central Kenya Rift: Tectonophysics, v. 236, p. 151–164.
Wescott, W.A., W.N. Krebs, D.W. Englehardt, and S.M. Cunningham, 1991, New biostratigraphic age dates from the Lake Rukwa basin in western Tanzania: AAPG Bulletin, v. 75, p. 1255–1263.
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: Journal of African Earth Science, v. 16, p. 425–435.
Williams, L.A.J., 1978, The volcanological development of the Kenya Rift, in E.R. Neumann and I.B. Ramberg, eds., Petrology and geochemistry of continental rifts, D. Reidel: Dordrecht, Holland p. 101–121.
Williamson, P.G., and R.J.G. Savage, 1986, Early rift sedimentation in the Turkana basin, north Kenya, in L.E. Frostick, R.W. Renaut, I. Reid, and J.-J. Tiercelin, 1986, Sedimentation in the African rifts, Geological Society of London Special Publication, 25, p. 267–284.
Young, P.A.V., P.K.H. Maguire, N.d'A. Laffoley, and J.R. Evans, 1991, Implications of the distribution of seismicity near Lake Bogoria in the Kenya Rift: Geophysics Journal International, v. 105, p. 665–674.
Zana, N., and H. Hamaguchi, 1978, Some characteristics of aftershock sequences in the western rift valley of Africa: Scientific Report Tokyo University Series 5, Geophysics, v. 25, p. 55–72.
Zoback, M.L., 1992, First- and second-order patterns of stress in the lithosphere: the world stress map project: Journal of Geophysical Research, v. 97, p. 11703–11728.

REFERENCES CITED (Part 2)

BEICEP, 1987, Kenya Tertiary Rift study: report to National Oil Corporation of Kenya, unpublished.
Bishop, W.W., and G.R. Chapman, 1970, Early Pliocene sediments and fossils from the northern Kenya Rift Valley: Nature, v. 226, p. 185–192.
Bishop, W.W., and M.H. Pickford, 1975, Geology, fauna and paleoenvironments of the Ngorora Formation, Kenya Rift Valley: Nature, v. 254, p. 185–192.
Chapman, G.R., S.J. Lippard, and J.E. Martyn, 1978, The stratigraphy and structure of the Kamasia Range, Kenya Rift Valley: Journal of the Geological Society of London, v. 135, p. 265–281.
Ego, J.K., 1994, Sedimentology and diagenesis of Neogene sediments in the central Kenya Rift Valley, M.Sc. thesis (unpublished), University of Saskatchewan: Regina, Canada 148 pp.
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 Kenya Rift: Journal of the Geological Society of London, v. 149, p. 333–348.
Ngenoh, D.K.A., 1993, Hydrocarbon potential of South Kerio Trough basin (Kenya) from seismic reflection data, M.Sc. thesis (unpublished), University of Western Ontario: London, Canada 126 pp.
Pope, D.A., 1992, Analyses and interpretation of seismic reflection profiles from the Kerio Valley, Kenya Rift, M.Sc. thesis (unpublished), University of Leeds: Leeds, England 133 pp.