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The East African rift system

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Abstract

This overview paper considers the East African rift system (EARS) as an intra-continental ridge system, comprising an axial rift. It describes the structural organization in three branches, the overall morphology, lithospheric cross-sections, the morphotectonics, the main tectonic features—with emphasis on the tension fractures—and volcanism in its relationships with the tectonics. The most characteristic features in the EARS are narrow elongate zones of thinned continental lithosphere related to asthenospheric intrusions in the upper mantle. This hidden part of the rift structure is expressed on the surface by thermal uplift of the rift shoulders. The graben valleys and basins are organized over a major failure in the lithospheric mantle, and in the crust comprise a major border fault, linked in depth to a low angle detachment fault, inducing asymmetric roll-over pattern, eventually accompanied by smaller normal faulting and tilted blocks. Considering the kinematics, divergent movements caused the continent to split along lines of preexisting lithospheric weaknesses marked by ancient tectonic patterns that focus the extensional strain. The hypothesis favored here is SE-ward relative divergent drifting of a not yet well individualized Somalian plate, a model in agreement with the existence of NW-striking transform and transfer zones. The East African rift system comprises a unique succession of graben basins linked and segmented by intracontinental transform, transfer and accommodation zones. In an attempt to make a point on the rift system evolution through time and space, it is clear that the role of plume impacts is determinant. The main phenomenon is formation of domes related to plume effect, weakening the lithosphere and, long after, failure inducing focused upper mantle thinning, asthenospheric intrusion and related thermal uplift of shoulders. The plume that had formed first at around 30 Ma was not in the Afar but likely in Lake Tana region (Ethiopia), its almost 1000 km diameter panache weakening the lithosphere and preparing the later first rifting episode along a preexisting weak zone, a Pan-African suture zone bordering the future Afar region. From the Afar, the rift propagated afterward from north to south on the whole, with steps of local lithospheric failure nucleations along preexisting weak zones. These predisposed lines are mainly suture zones, in which partial activation of low angle detachment faults reworked former thrust faults verging in opposite directions, belonging to double verging ancient belts. This is responsible for eventual reversal in rift asymmetry from one basin to the next. Supposing the plume migrated southward, or other plumes emplaced, the rift could propagate following former weaknesses, even outside areas influenced by plumes. This view of rift formation reconciles the classical models: active plume effect triggered the first ruptures; passive propagations of failure along lithospheric scale weak zones were responsible for the onset of the main rift segments. Various other aspects are shortly considered, such as tectonics and sedimentation, and relationships of the 'cradle of Mankind' with human evolution. By its size, structure and occurrence of oceanic lithosphere in the Afar, the EARS can be taken as a model of the prelude of oceanic opening inside a continent.

Keywords: Africa; Rift; Transform fault; Human evolution; Volcanism; Tectonic heritage; Transfer fault; Lithosphere; Hot spot; Continental break up; Passive margin

1. Introduction

The concept of 'East African rift fracture' was established by Suess (1891), following the explorations and discoveries of Livingstone, Stanley, Fischer, Thomson, Teleki and Von Höhnels during the XIXth century. Gregory in 1896 named it the 'Great Rift Valley of East Africa', and in 1921 he described a system of graben basins including the Red Sea and Dead Sea systems, forming the Afro-Arabian rift system. The present paper is focused on the African part, trying to update major knowledge and

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concepts. Several synthetic papers have been written before, e.g., Willis (1936), Quennel (1960), McConnel (1967, 1972), Baker et al. (1972), Mohr et al. (1972), Mohr (1982), Girdler (1991), Morley (1995), Burke (1996), Frostick (1997), Schlüter (1997). This new overall presentation is from a structural geologist who, since 1979, has been carrying out fieldwork and remote sensing analysis along the entire 6000 km north–south stretch of the rift, and has been almost everywhere.

The history of geological researches dedicated to the East African rift system (EARS) comprises several steps. After the discovery expeditions, detailed local geological studies span from the beginning of the 1930s to the end of the 1960s. These essential works were focused on mapping (e.g., Smith, 1931; McCall, 1957; McConnel, 1959; Harkin, 1960). The second step began during the1960s, lasting in the 1970s, with more regional scale studies of some aspects of the geology, especially tectonics, volcanism and acquisition of geophysical data (e.g., Mohr, 1962; Harris, 1969; Wohlenberg, 1969).

Research was boosted in the 1970s by the introduction of the concept of global tectonics, considering that oceans and related accretion of oceanic lithosphere begin with an intra-continental rift stage (Baker, 1970; Girdler and Sowerbutts, 1970; Mohr, 1970; Searle, 1970; Vail, 1970; Baker et al., 1971; McKenzie et al., 1972; Fairhead and Girdler, 1972; Chorowicz and Mukonki, 1979). This was also the time of first acquisitions of space imagery, covering large scenes with good ground resolution, a powerful tool for structural analysis of large areas (Mohr, 1974; Chorowicz and Mukonki, 1979).

In the last step, since the 1980s, the EARS remains the archetypal continental rift and a widely analogue to the early stages of evolution of passive continental margins preceding oceanic opening. Numerous papers concern the various aspects of rift evolution (e.g., Mohr, 1983; Bosworth, 1985; Mougenot et al., 1986a,b; Chorowicz et al., 1987; Ebinger et al., 1987; Daly et al., 1989; Strecker et al., 1990; Kampunzu and Lubala, 1991; KRISP Working Party, 1991; Rosendahl et al., 1992; Tiercelin et al., 1992a,b). The 1980's also saw the first reflection seismic profiling of many of the rift basins (e.g., Ebinger et al., 1984; Le Fournier et al., 1985; Rosendahl et al., 1986; Ebinger et al., 1987; Morley, 1988, 1989, 1999; Morley et al., 1992a,b). The number of papers is still increasing, making it difficult for non specialists to have a clear view of the rift evolution, and justifying this synthesis.

The particularity of this new overview is to consider the East African rift system as an intra-continental ridge system, comprising an axial rift, prelude of oceanic opening. Section 2 describes the structural organization of the rift, including overall morphology, cross-sections, the morphotectonics, the main tectonic features that can be found, and volcanism. Section 3 focuses on the kinematics, with discussion of the directions of movements, in the frame of the significance of transform fault, transfer and accommodation zones. Section 4 is an attempt to reconstitute the rift system evolution through time and space, considering various aspects such as rift propagation, relationships with ancient structures, mechanism of formation, tectonics and sedimentation, and relationships of the 'cradle of Mankind' with human evolution.

2. Structure

2.1. Overall morphology

The East African rift system shows up at the surface as a series of several thousand kilometers long aligned successions of adjacent individual tectonic basins (rift valleys), separated from each other by relative shoals and generally bordered by uplifted shoulders (Fig. 1). Each basin is controlled by faults and forms a subsiding graben or trough, near one hundred kilometers long, a few tens kilometers wide, empty or filled with sediments and/or volcanic rocks.

The rift valleys form two main lines, the eastern and western branches of the EARS. A third, southeastern branch is in the Mozambique Channel. The eastern branch runs over a distance of 2200 km, from the Afar triangle in the north, through the main Ethiopian rift, the Omo-Turkana lows, the Kenyan (Gregory) rifts, and ends in the basins of the North-Tanzanian divergence in the south. The western branch runs over a distance of 2100 km from Lake Albert (Mobutu) in the north, to Lake Malawi (Nyasa) in the south. It comprises several segments: the northern segment includes Lake Albert (Mobutu), Lake Edward (Idi Amin) and Lake Kivu basins, turning progressively in trend from NNE to N–S; the central segment trends NW–SE and includes the basins of lakes Tanganyika and Rukwa; the southern segment mainly corresponds to Lake Malawi (Nyasa) and small basins more to the south. The south-eastern branch comprises N-striking undersea basins located west of the Davie ridge. Most of the great lakes of Eastern Africa are located in the rift valleys, except notably Lake Victoria whose waters are maintained in a relative low area between the high mountains belonging to the eastern and western branches.

These successions of graben basins are generally bordered on the two sides by high relief, comprising almost continuous parallel mountain lines and plateaus, and sometimes volcanic massifs. The map of Fig. 2 shows the areas uplifted above 1200 m in Africa. Those belonging to the EARS are comprised in two ellipses, one is the Ethiopian dome, and the other includes the Kenyan and Tanzanian domes. The longest axes of the two ellipses have a NNE trend: at this scale the main expression of the EARS is uplift, forming on the whole a NNE-trending intra-continental ridge, interrupted by the Omo-Turkana lows. The highest elevations in the EARS region, in addition to volcanoes, are the graben shoulders. Other uplifted areas in the region are due to belts, which may be recent (Atlas, Zagros belts) or ancient (Karroo belt), or to intracontinental hotspots (Hoggar, Tibesti plateaus). The overall geomorphology results from processes affecting the whole

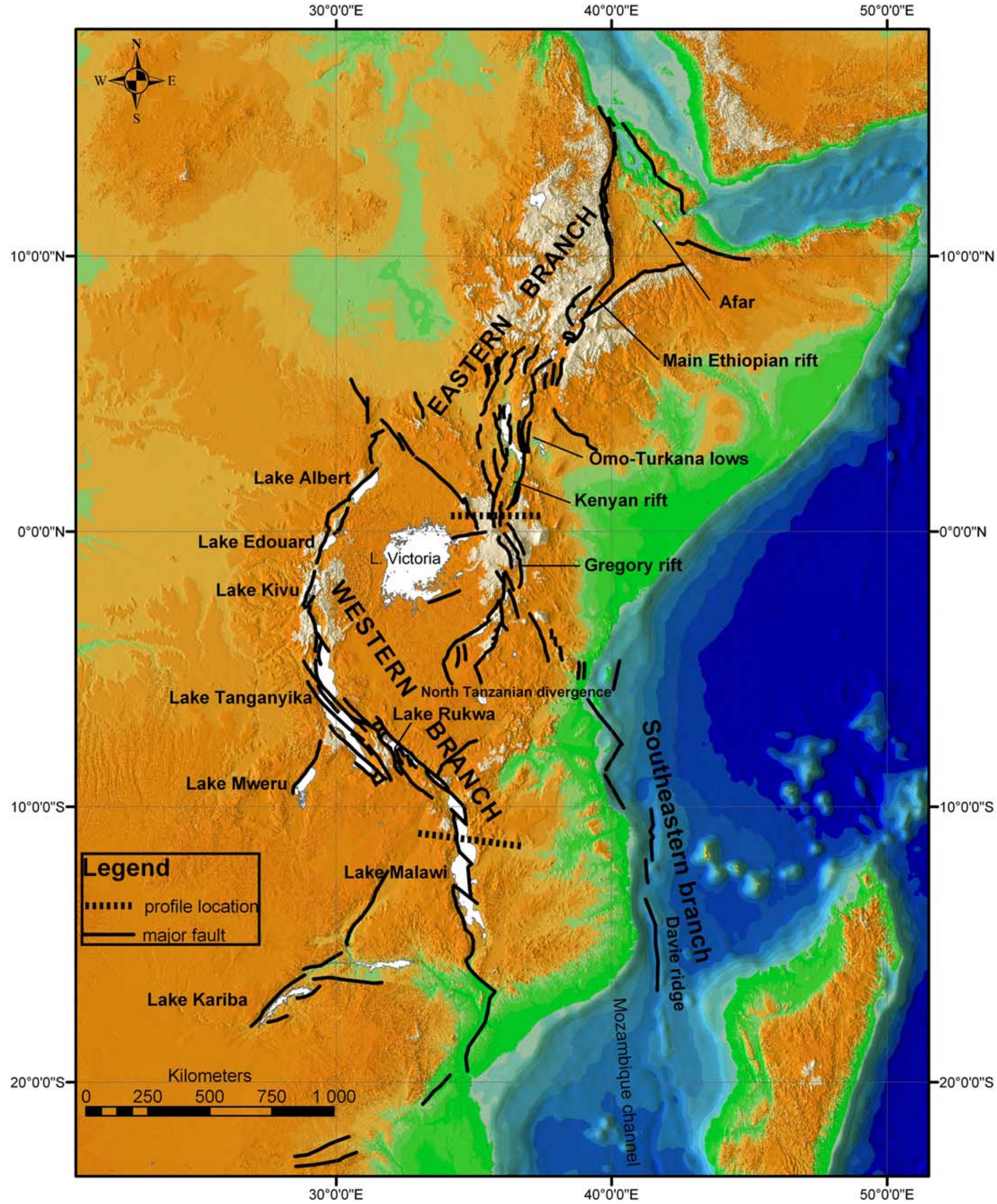

Fig. 1. Hypsographic DEM of the East African rift system. Black lines: main faults; E–W dotted lines: locations of cross-sections of Fig. 3; white surfaces: lakes; grey levels from dark (low elevations) to light (high elevations). The East African rift system is a series of several thousand kilometers long aligned successions of adjacent individual tectonic basins (rift valleys), separated from each other by relative shoals and generally bordered by uplifted shoulders. It can be regarded as an intra-continental ridge system comprising an axial rift.

lithosphere, and can be best described using examples of representative lithospheric cross-sections.

3. Lithospheric cross-sections

3.1. Example of the Malawi rift region

The cross-section of Lake Malawi (Nyasa) presents the main characteristics of most of the other graben basins (Fig. 3). This rift is at an early stage of development. The seismic reflection profile A (Specht and Rosendahl, 1989; Rosendahl et al., 1992) shows that thickness of the sediments is ~3000 m. The graben structure is asymmetric. (1) Almost all the tilted blocks dip in the same westward direction. (2) Accordingly, all the extensional faults have the same 65–70° eastward dip (antithetic faults), except a few synthetic faults in the west. (3) The throws of most of the faults do not exceed 600 m, whereas that of the westernmost border fault zone is approximately 7000 m if erosion is taken into account. This establishes the concept of a major fault bordering the one side, whilst on the other side the boundary fault is simply one of the many smaller faults.

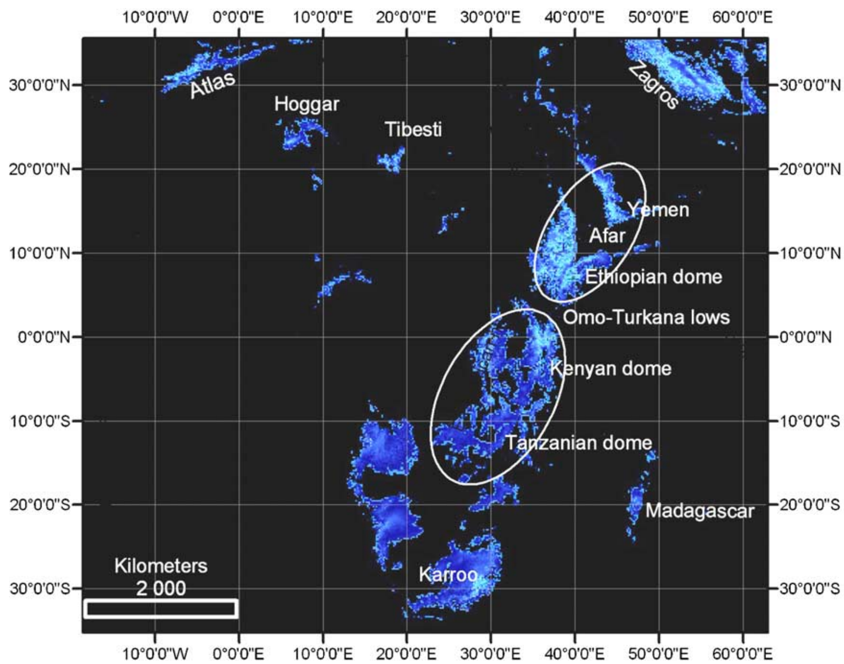

Fig. 2. Map of Africa showing in grey levels the elevations higher than 1200 m, evidencing the main Ethiopian and Kenyan–Tanzanian domes.

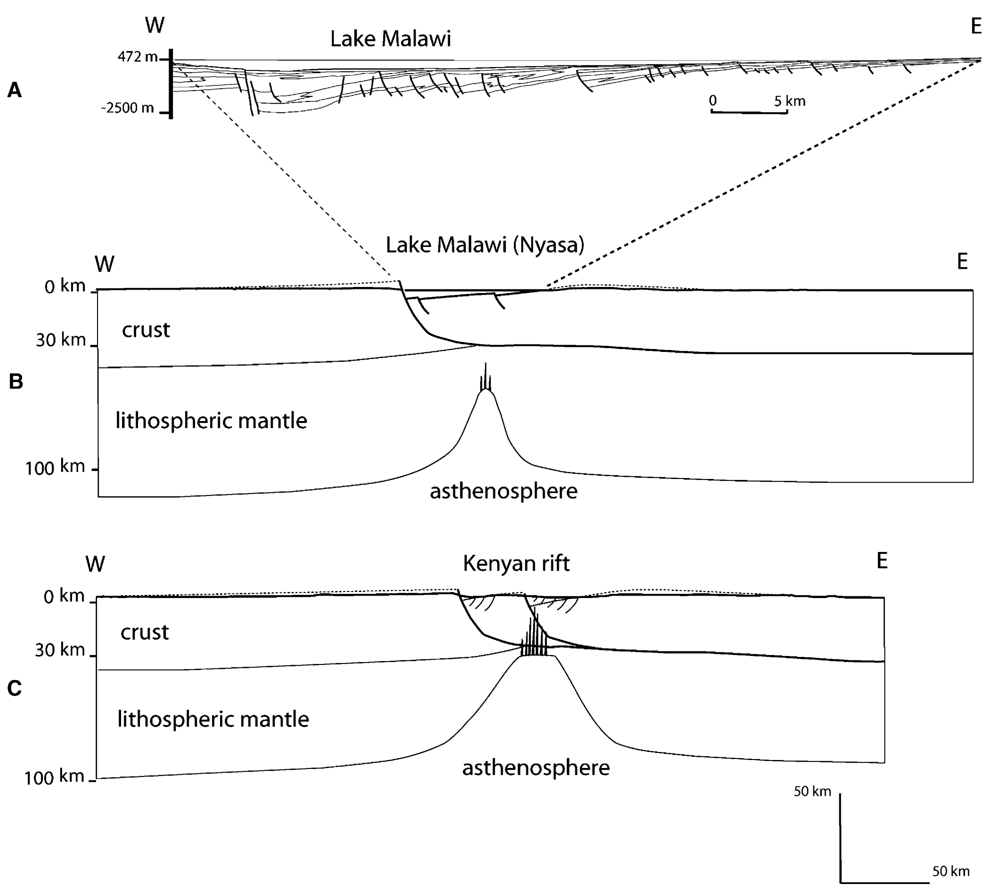

Fig. 3. Representative lithospheric profiles of the EARS. See location in Fig. 1. Vertical lines are inferred dyke intrusions. Dotted lines show the envelope surface before erosion. A. Seismic reflection profile of Lake Malawi (Specht and Rosendahl, 1989; Rosendahl et al., 1992). B. Inferred lithospheric cross-section of Lake Malawi. C. Inferred lithospheric cross-section of the northern Kenyan rift.

(4) The overall structure is that of a roll-over pattern dipping west, accompanying a unique major east-dipping fault along the western border.

Another important character of the seismic profile is the high density of faults in the earliest deposits (1 fault/km). In the younger layers, the number of faults decreases with time. Many faults have ceased to be active in turn, and their throws remain small (<100 m). Progressively, the higher in the sedimentary succession, the less numerous are the faults and the greater their throws. Considering the throw of the major bordering fault, with time this fault zone alone accommodates most of the deformation. It is high angle (dip 65°) near the surface, and is generally interpreted to be listric, abruptly becoming low angle downward (Fig. 3B) when it finally connects with sub-horizontal shears in the lower crust at depth to detachment that depends on crust thickness, between 15 and 30 km for instance in the western branch (Morley, 1989). The normal crust thickness in East Africa is 40 ± 5 km, according to Prodehl et al. (1997) from seismic-refraction surveys.

The W-E width of uplifted areas across rift direction is ~300 km. The profile takes into account rift shoulders that reach ~1300 m higher than the rift floor, after erosion. Uplift of the crust in the shoulders is related to ascent of an asthenospheric body, elongated in rift direction. This pattern in the mantle is almost symmetric. The rift consequently is asymmetric in the crust and symmetric in the lithospheric mantle.

However, the overall topographic surface is generally 500 m higher in the west that in the east of Lake Malawi, meaning the lithosphere is thicker in the west. This may correspond to a variation in lithospheric thickness inherited long after the Pan-African collision of two different lithospheric blocks.

3.2. Example of the Kenyan rift region

The Kenya rift region is a well documented area in terms of structure of the lithosphere. It has beneficiated from gravimetric studies (Wohlenberg, 1975a; Fairhead, 1976; Fairhead and Reeves, 1977; Girdler, 1978) and other geophysical studies (KRISP Working Party, 1991; Achauer et al., 1992), including seismic refraction-wide-angle reflection experiments (Mechie et al., 1997; Keller et al., 1994).

According to Achauer et al. (1992), crustal thickness beneath the graben valley is 30 km at latitude 0.5°, 40 km in western flank, 35 km in eastern flank. In the graben (Fig. 3C), sediments and volcanic layers are 3 km thick. Normal faulting and tilted blocks express rift asymmetry, with two major listric faults and associated roll-over structures.

The shoulders, partly covered by volcanic rocks, are ~2600 m high in the west, ~2000 m in the east. There is a regional negative Bouguer anomaly interpreted (Wohlenberg, 1975a; Fairhead, 1976) to express uplift of an asthenospheric body. Keller et al. (1994) have described a sharply defined lithospheric thinning, with low upper mantle velocities down to depths of over 150 km. A distinct relative positive residual anomaly in the middle of the graben is due to dyke injection, related to reservoirs of magma.

Focused extensional strain in the lithosphere is responsible for downward thinning of the crust—forming basins, and upward thinning of the lithospheric mantle (density 3.3) replaced by lower-density asthenosphere (density 3.2) in a state of partial melt. As is the case in Lake Malawi (Nyasa), the cross-section (Fig. 3C) expresses a pronounced rift asymmetry in the crust, but the lithospheric mantle has symmetric geometry.

The cross-sections are representative and show that the EARS is a lithospheric structure. Architecture is controlled by asthenospheric intrusion in the lithospheric mantle and by fault geometry in the crust. Pronounced variations in tectonics and related morphology all along the rift system are presented from north to south in the next section.

4. Morphotectonics

4.1. Eastern branch

4.1.1. Afar region

The Afar triangle (Fig. 4) is the triple junction between the African, Arabian and Somalian plates. It is floored by Mio-Pliocene "Red Bed" unit (Kazmin, 1976) dated from 24 to 5.4 Ma and Quaternary tholeiitic volcanic rocks (Barberi et al., 1972). Elevations are low, some under sea level (−176 m), the basement being made of high density oceanic and thin continental lithospheres. It is not yet clearly established which parts of the Afar depression are of oceanic type and which are thinned continental lithosphere (e.g., Tazieff, 1973; Barberi and Varet, 1975; Makris and Ginzburg, 1987; Souriot and Brun, 1992; Ebinger and Hayward, 1996; Hayward and Ebinger, 1996; Manighetti et al., 1997; Manighetti et al., 2001).

The eastern corner of the triangle is an oceanic rift vanishing west of the Gulf of Tadjura, forming the western end of the oceanic ridge of the Gulf of Aden. The northern corner is formed by the Erta Ale volcanic belt, a typical active oceanic ridge constituting the northwestern-end of the transitional Afar lithosphere (Tazieff, 1973; Barberi and Varet, 1975). Only the southern half of the Afar fully belongs to the EARS. It is an asymmetric rift comprising a series of east–southeast-dipping tilted blocks forming plateaus, bounded by normal fault scarps, which are parallel to the rift axis (Fig. 5).

To the northeast, the Afar triangle is bordered by the elongate Danakil block, comprising Pan-African basement (Fig. 4). The block is tangent to Africa at its northwestern end and to Arabia in the southeast. It has suffered anticlockwise rotation ('*bielette Danakil*' of Sichler, 1980).

The Somalian plateau, lying to the south of the Afar, is northwesterly bounded by a crest line at ~3000 m elevation. The plateau progressively downgrades to the southeast, to finally form plains at a few hundred meters elevation. This is expressed by the envelope surface in the

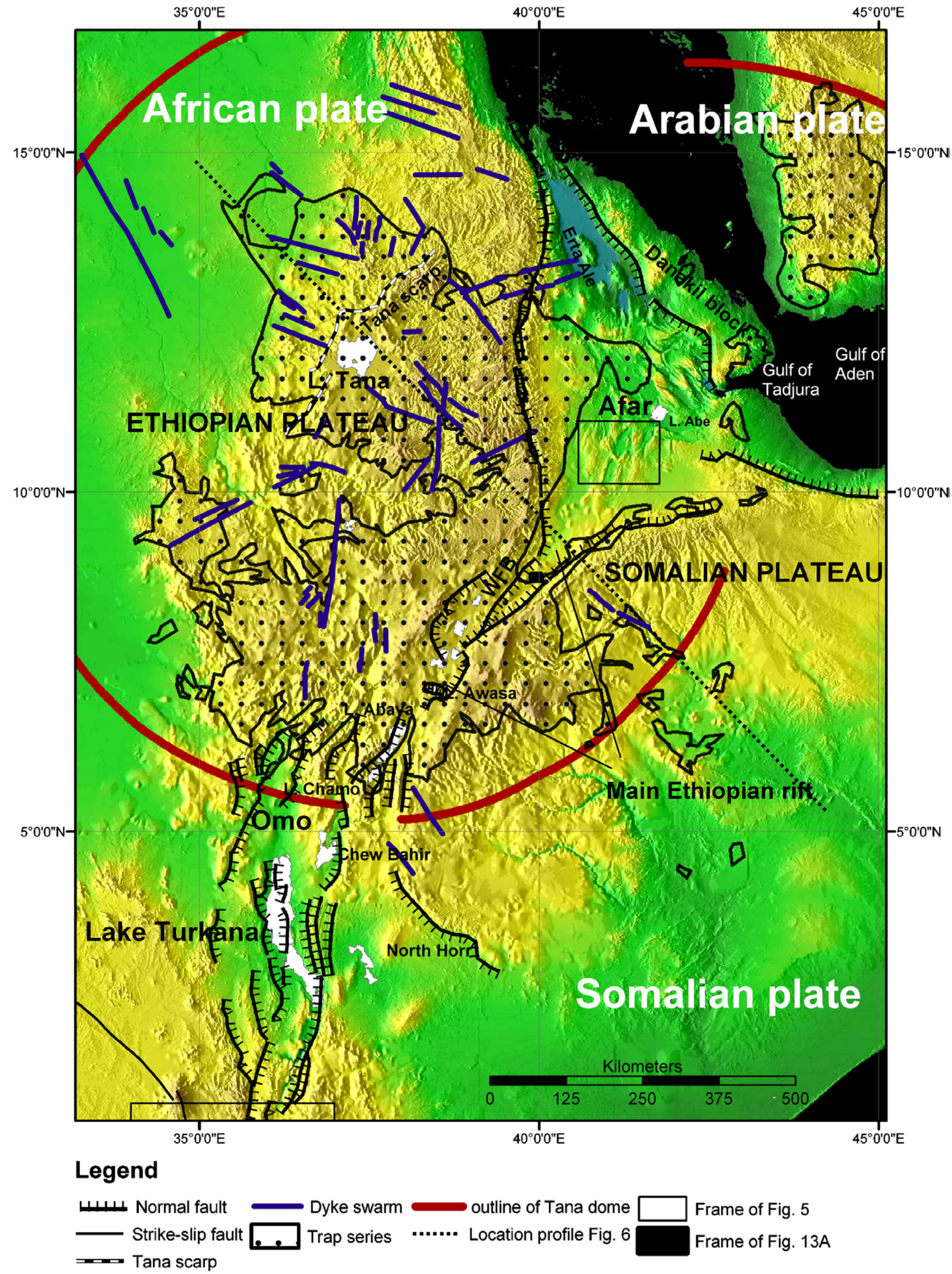

Fig. 4. Rift patterns in the Afar-Ethiopian region, overprinted on DEM (Gtopo30) shadow image.

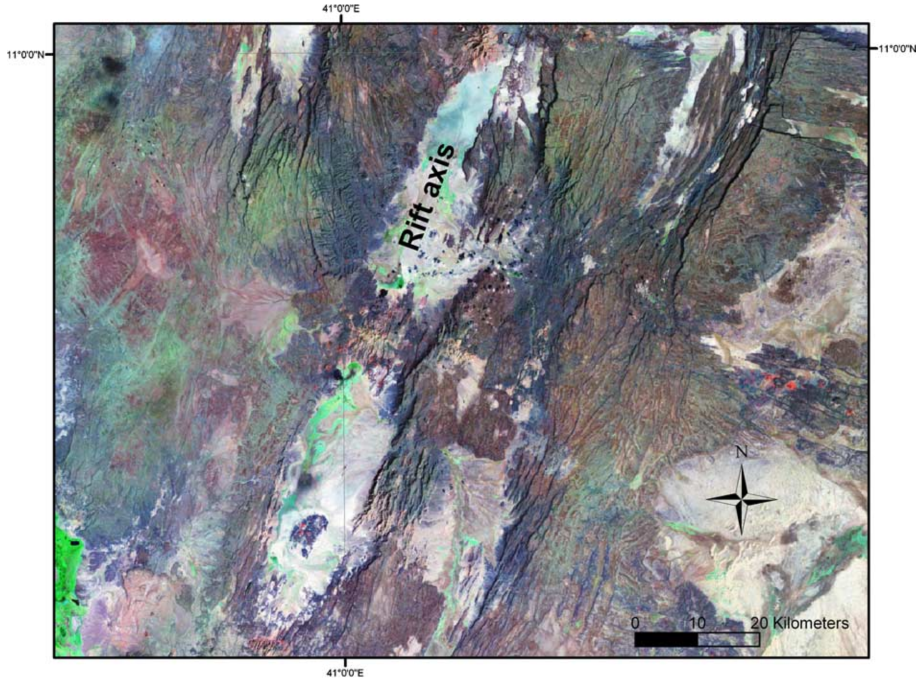

Fig. 5. Tilted blocks and fault scarps on black and white Landsat TM image of southern Afar. Location is shown (rectangle) in Fig. 4. All the blocks dip in the same east–southeastern direction.

topographic cross-section of Fig. 6 made from the GTopo30 DEM, forming a typical profile of a shoulder that results from tectono-thermal evolution in a rift.

The western side of the Afar is the Ethiopian plateau (Fig. 4), at elevations ~3600 m but reaching more than 4000 m by places. There is a sharp scarp transition zone over a distance of 70 km from 3700 m at the edge of the plateau, to 500 m in the depression, which comprises an almost continuous Plio-Quaternary rift following the 600 km stretch of the Afar western margin, only 10–15 km wide. This graben system is separated from the Afar low plains by an elongate continuous tilted block (600 km long, 30–60 km wide, 1000–2000 m in elevation), dipping gently eastward (Zanettin and Justin-Visentin, 1975). This feature can be compared to the Neotethyan passive paleo-margin of Jurassic age in the Western Alps, the large block being an equivalent to the Briançonnais zone or similar blocks known in passive margins (Chorowicz et al., 1999). This pattern can be taken as a model for the formation of complex passive margins, and results in the Afar from three successive tectonic events: (1) left-lateral strike-slip movement along a N–S trajectory of the western tip of the Danakil block during the early Miocene; (2) NW–SE extension during Miocene formation of the main Ethiopian rift; (3) eastward Plio-Quaternary gravity gliding of the tilted horst block, at the transition zone between the thick continental lithosphere and the thin continental or oceanic lithosphere.

In the Ethiopian plateau, considering Fig. 4 and cross-section of Fig. 6, the morphology appears very different from that of the Somalian plateau in the opposite side. Major shield volcanoes are well expressed and there is an important volume of volcanic rocks, especially the up to 1000 m thick trap series, dated 30 Ma (Chernet et al., 1998; Hoffman et al., 1997). But this volcanic pile does not enough explain the higher elevations, which are more

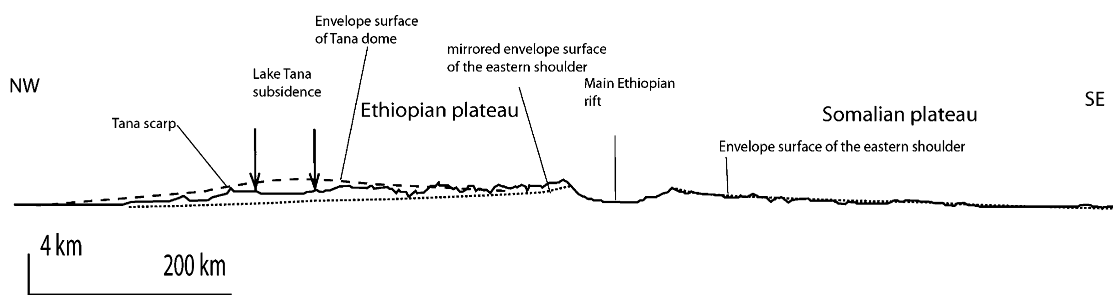

Fig. 6. Topographic profile with vertical exaggeration (location in Fig. 4), showing Tana dome added to 'normal' thermal envelope profile. 'Normal' thermal envelope profile is mirrored from that of the Somalian plateau.

than 3000 m in Tana area. Considering the envelope surfaces that can be drawn over the rift shoulders, which should be symmetrical if due only to asthenospheric intrusion, the Tana basin is clearly perched on an additional regional topographic swell. Subsidence of Lake Tana occurs at the centre of the swell. The elevations in the Ethiopian plateau are consequently related to another system of uplift, forming the Tana dome.

A consequence of this uplift is that the volcanic pile and dome are cut in their western part by a front of regressive erosion called the Tana scarp, which has taken away at least 1000 m thick of stratoid lavas west of the escarpment, exposing the basement and large dykes that had fed the trap pile at 30 Ma (Mège and Korme, 2004). The boundaries of the volcanic pile and dome are now separated into three parts (Fig. 4): mainly in the Ethiopian plateau, to be completed by the part that has been taken away by the erosion, but the feeding dykes now exposed by erosion show the extent of the former lavas; in the southern part of the Arabian peninsula; in the Somalian plateau. The original distribution of the mid-Tertiary trap basalts and the Tana dome, prior to opening of the Red Sea and Gulf of Aden, had an ovoid shape in plan view, with a longitudinal axis passing much closer to Lake Tana than to the middle of the Afar. The axis trends parallel to the Paleogene motion of Africa (Bonavia et al., 1995). These patterns are related to a hotspot plume occurring in Northeast Africa at 30 Ma (Schilling, 1973; Schilling et al., 1992; Keller et al., 1994; Hoffman et al., 1997). The core hit the African lithosphere at Lake Tana (Chorowicz et al., 1998).

4.1.2. Ethiopian rift region

The main Ethiopian rift (MER), striking NE, is connected to the southern corner of the Afar (Fig. 4). The rift floor is ~50 km in width, with a length of about 330 km. Its elevations are 750 m in the NE; they progressively rise south-westwards to 1700 m, and finally downgrade at 1100 m in the south. Lava flows and evaporites fill in great part the graben basins, but detritic sediments are few because the climate is dry since the Miocene. The MER is continuously bordered on the two sides by crest lines, with abrupt transitory scarp faces overlooking the rift valley floor. The eastern crest line elevations vary between 2500 m and 3000 m. On the western side, the elevations of rift shoulders are comprised between 1800 and 3500 m. The rift is segmented into graben basins, each one asymmetric, the graben fill dipping east, and the major fault lying along the eastern side. A narrow fault zone, the Wonji Fault Belt (WFB, Mohr, 1962) obliquely crosses the MER between northwestern border in the north and southeastern border in the south. The MER ends near Lake Awasa.

4.1.3. Omo-Turkana low lands

The elevations (in places less than 400 m) in the Omo-Turkana region are significantly lower than the nearby dome regions (Fig. 4). This low area has no pronounced rift and shoulders, but there are several N- to NE-striking half-graben basins (e.g., Omo valley, lakes Turkana and Chew Bahir). The North Horr basin has a NW strike, with the major fault at its NE border. It is located apart from the main region of half-grabens and can be considered as an appended basin. Each half-graben is filled with late Cenozoic sediments, some up to 7 km thick. The first basins were initiated west of lake Turkana during late Oligocene-early Miocene times, then faulting gradually shifted eastward with time (Morley et al., 1992a,b; Morley, 1999).

According to Mahatsente et al. (2000), thickness of crust in the western and eastern plateaus of the MER is 51 km and less than 31 km in Omo basin. About 20 km in the Turkana area, the crust thickens to about 35 km south of Lake Baringo in the Kenyan rift (KRISP Working Party, 1991; Mechie et al., 1994).

4.1.4. Kenyan (Gregory) rift region

The Kenyan (Gregory) rift region corresponds to the Kenyan dome (Fig. 2). Three major central volcanoes (Kenya 5200 m, Elgon 4321 m, Kilimanjaro 5964) are situated in its vicinity (Fig. 7).

The northern Kenyan rift is developed in the northern part of the dome. It is composed of two parallel rift valleys at ~1050 m elevation, both trending N10 °E, separated by the Kamasia block, at ~2500 m elevation by places (Fig. 8). The eastern one (northern Kenyan rift *s.s.*) prolongs the southern part of Lake Turkana half-graben, and continues to the south, to reach the area of Lake Bogoria (Fig. 7). The Kerio graben ends abruptly west of Lake Bogoria (Fig. 8). Both troughs are west-dipping roll-over half-grabens, with major border faults running along the western shoulders (Fig. 3C). These major faults are sometimes accompanied by hydrothermal venues as is the case for Lake Bogoria (Renaut and Owen, 1988). The western scarp border of Kerio graben is lightly eroded, attesting to more recent formation than the eastern scarp.

The central Kenyan rift is individualized by a sharp turn in rift direction at the southern end of the northern Kenyan rift valley, at the latitude of Lake Bogoria (Fig. 7). Its N150 °E trending segment is set in the middle of the Kenyan dome, with elevations reaching more than 3700 m in the eastern shoulder and more than 3000 m in the western one. Elevations of the rift floor rise from ~1050 m in the north, to ~2100 m in the middle, and steps down progressively southwards to ~600 m at Lake Natron. Several authors (e.g., Chorowicz and Vidal, 1987; Smith and Mosley, 1993) have explained this bend as intersection with a large NW-striking basement structure called the 'Aswa lineament'.

The N80 °E trending Nyanza half-graben branches with the central Kenyan rift near Lake Bogoria (Fig. 7), and disappears westward under Lake Victoria. The rift floor is generally at 1100 m, the immediate shoulders being at 1600 m. The lake Bogoria region is a triple junction between the northern Kenyan, central Kenyan and Nyanza rifts, almost at centre of Kenyan dome. Time evolution of Kenya rift comprises several periods (Hackman et al.,

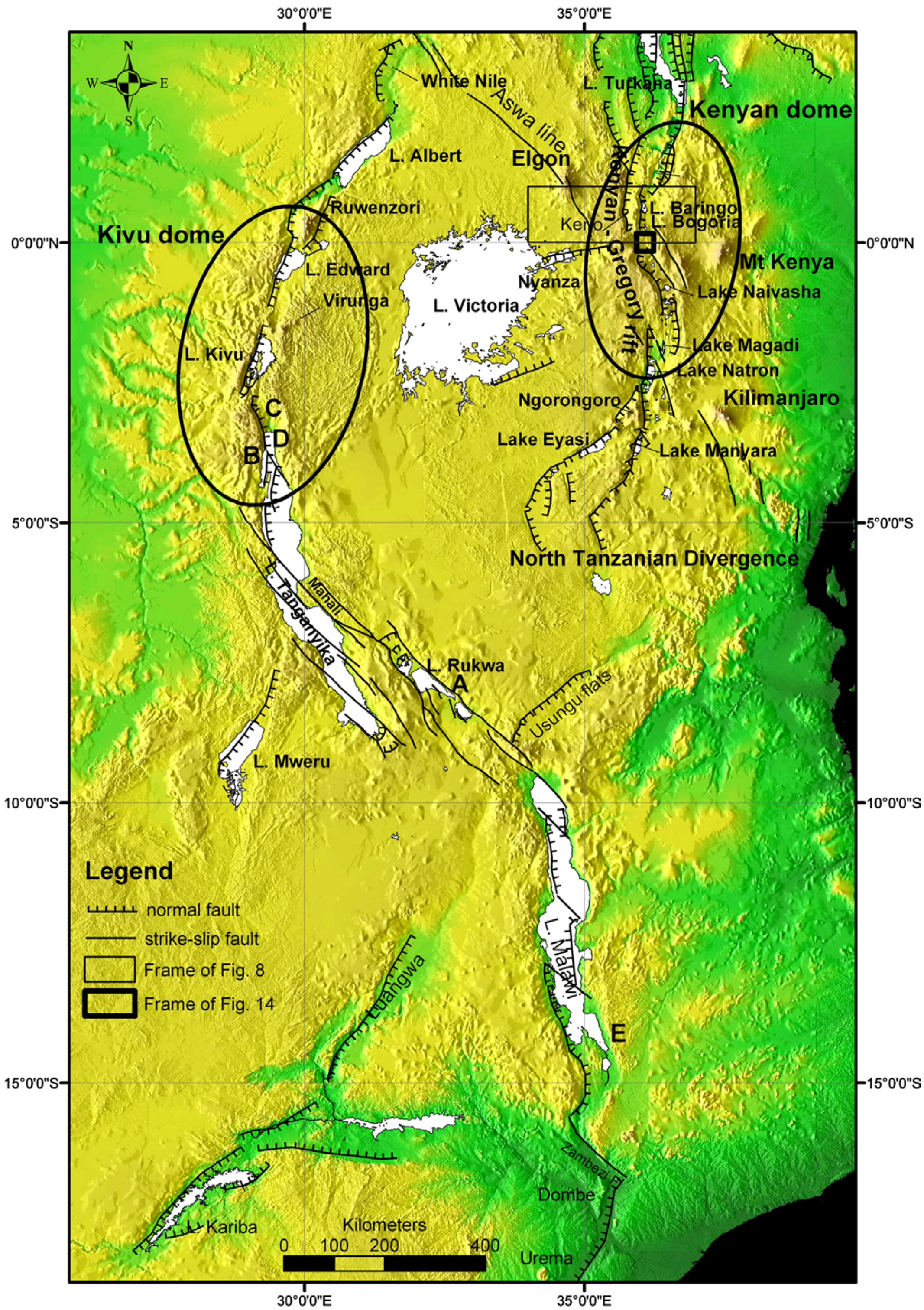

Fig. 7. Western branch and part of eastern branch of the East African rift system, on shadowed DEM. (A–E) location of the sketches of Fig. 12.

1990; Smith and Mosley, 1993): (a) 23–11 Ma, first volcanism and formation of a low area in the middle of the future dome; (b) 11–5.3 Ma, individualization of graben basins and uplift of shoulders; (c) 5.3–1.6 Ma, formation of the Kerio valley; (d) 1.6–0.01 Ma, concentration of activity in eastern part of northern Kenyan rift *s.s.* and central Kenyan rift.

4.1.5. North Tanzanian divergence

Near Lake Natron, the southern Kenyan rift splays out southward to form the North Tanzanian divergence, with three main branches: Lake Eyasi, Lake Manyara and elongated plains south of the Kilimanjaro. These diverging half-graben valleys are outside the Kenyan dome, and spread over a large width.

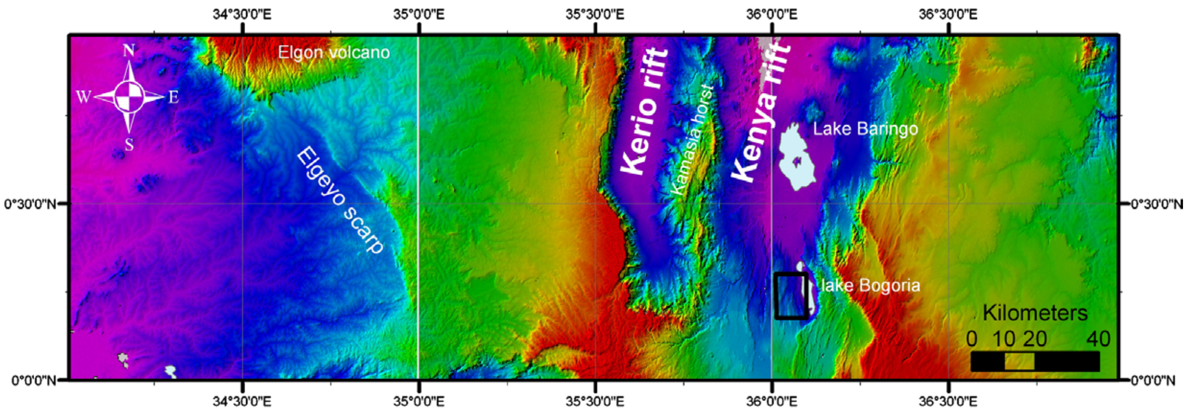

Fig. 8. Shadowed Shuttle Radar Topographic Mission DEM of Kenyan rift, Kerio valley and Elgon volcano area. See location in Fig. 7. Rectangle is location of Fig. 13C.

Pn wave velocities indicate that there are no broad (>100 km wide) thermal anomalies in the uppermost mantle beneath this area (Brazier et al., 2000). It is likely that these half-graben basins affect only the crust and are not underlain by asthenospheric ascent. The EARS does not exist significantly south of the Kenyan rift.

4.2. Western branch

4.2.1. Northern segment

The northern rift segment of the western branch is an arched succession of half-graben basins, most of them hosting a lake (Fig. 7). They are developed in the Kivu dome. The Virunga NE-trending group of large central volcanoes (e.g., Nyiragongo, Nyamulagira) lies in the centre of the dome.

Lake Albert (Mobutu) is at 618 m elevation with shoulders more than 2200 m high in the west and more than 1300 m in the east. To the north, there is a Neogene White Nile graben. Separated from Lake Albert by the Ruwenzori highs, Lake Edward (Idi Amin) has rift floor at 200 m, and shoulders at 2300 m (west) and 1600 m (east). The rift floor steps up progressively southward to reach Lake Kivu (1420 m), which is bordered by shoulders at ~3000 m elevation.

To the south of Lake Kivu, the rift floor rapidly steps down to northern Lake Tanganyika at elevation 773 m. The shoulders there are well developed, reaching 3400 m in the west and 2600 m in the east.

4.3. Central segment

4.3.1. Lake Tanganyika

The Tanganyika rift is more than 700 km long and up to 70 km wide, almost entirely covered by the 650 km long lake (Fig. 7). The water surface is 773 m above sea level, the maximum depth 1310 m and 1470 m in the northern and southern parts respectively, separated by the Mahali shoal. The maximum thickness of sediment fill is around 4000–5000 m (Sander and Rosendhal, 1989; Rosendahl et al., 1992), starting in the late Miocene or early Pliocene (Burgess et al., 1988; Rolet et al., 1991; Tiercelin and Mondeguer, 1991).

The northern basin is divided by NW-striking faults into three asymmetric sub-basins, from north to south, the Bujumbura (350 m deep), Rumonge (1150 m deep) and Kigoma sub-basins (1310 m deep). Each is bounded on the western side by a major listric normal fault with curved trace. The maximum vertical throw component is ~10000 m, erosion of the western shoulder being taken into account. This fault is accompanied by hydrothermal activity as is the case for the northern basin (Tiercelin and Mondeguer, 1991; Thouin and Chorowicz, 1993; Tiercelin et al., 1993).

The central and southern Tanganyika basins also are divided into sub-basins by NW-striking faults that are each more than 300 km in length. They are not always bordered by high shoulders but also frequently by plains (elevation, ~1000 m) and narrow hills that exceptionally reach elevations at around 2000 m.

4.3.2. Lake Rukwa area

The lake Rukwa basin is bordered by strike-slip faults that are connected to those of Lake Tanganyika (Chorowicz et al., 1983; Tiercelin et al., 1988; Wescott et al., 1991; Morley et al., 1992a,b; Kilembe and Rosendahl, 1992; Peirce and Lipkov, 1988). The main basin is more than 200 km long and 60 km wide, water surface at around 800 m. Sediments are of Plio-Quaternary age with thickness less than 1200 m (Ebinger et al., 1989; Wescott et al., 1991; Harper et al., 1999).

At right angle to the strike of the Rukwa basin, the Usungu flats form a half-graben appended to the main line of rift basins, without any uplifted shoulder (Fig. 7). To the west, the NNE trending Lake Mweru half-graben is situated apart from the main Tanganyika–Rukwa–Malawi alignment. There are no significant elevated rift shoulders.

4.3.3. Southern segment

The Malawi rift, 650 km long and 60 km wide, is mostly hidden by Lake Malawi (Nyasa), 500 km in length with water surface at 472 m and maximum depth ~700 m

(Fig. 7). Shoulders are at more than 2000 m in the central part, and ~1500 m in southern part. The rift is composed of four half-graben basins, separated by NW-striking faults (Ebinger et al., 1987; Flannery and Rosendahl, 1990; Chorowicz and Sorlien, 1992; Rosendahl et al., 1992). The northernmost basin (Livingstone basin) is late Miocene-Pleistocene to recent in age. In plan view the main border faults have a typical arc shape, and lie alternately along the eastern and the western borders.

West of Lake Malawi, the elongate Luangwa low lands have a NNE strike and belong to the Karroo rift system. This basin, appended to the main EARS line, may be due to reactivation of the Karroo normal faults, with slight subsidence, but the drainage of the basin is possibly responsible for lack of significant late Cenozoic deposits. There is no significant rift shoulder.

The Dombe and Urema half-graben basins form a line that connects to the southern tip of the Malawi basin by the NW-striking Zambezi fault. To the west, other isolated, less defined graben basins comprise Lake Kariba.

4.4. South-eastern branch

The south-eastern branch of the EARS is less developed than the others (Fig. 9). It comprises the Pemba, Mafia, Kerimbas and Lacerda basins (Mougenot et al., 1986a,b). These are more than 20 km wide half-graben zones of normal faulting. They are mostly parallel to the main trend of the margin and may have been favored by tendency to detachment of the margin sediments due to gravity forces. The Kerimbas and Lacerda basins are possibly related to reactivation of ancient strike-slip faults of the Davie Ridge.

5. Main tectonic features

The morphotectonics of the EARS are under control of divergent movements, inducing localized extensional strain in the continental lithosphere. The brittle crust has reacted by faulting and subsidence, forming elongate, narrow rifts, while the lithospheric mantle is subjected to sharply define ductile thinning, inducing ascension of asthenospheric mantle.

The most characteristic features in the rift system are then these narrow elongate zones of thinned lithosphere related to deep intrusions of asthenosphere in the upper mantle. This hidden part of the rift structure is expressed on the surface by thermal uplift of shoulders, and argued by geophysical data. There are not enough geophysical data to precisely map these narrow zones of thinned lithosphere. Map of Fig. 10 is mainly based on the

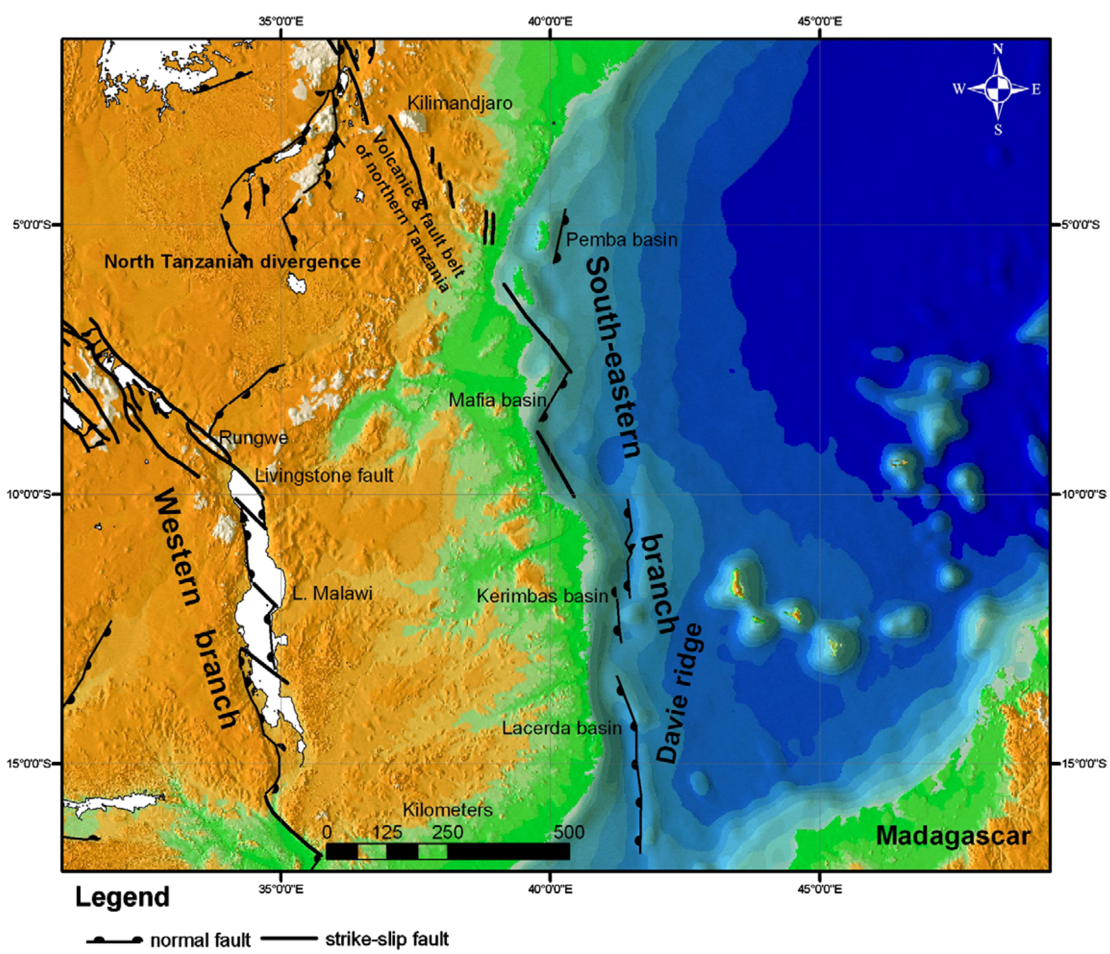

Fig. 9. Mozambique region of the EARS, showing the southeastern branch and volcanic and fault belt transform of northern Tanzania.

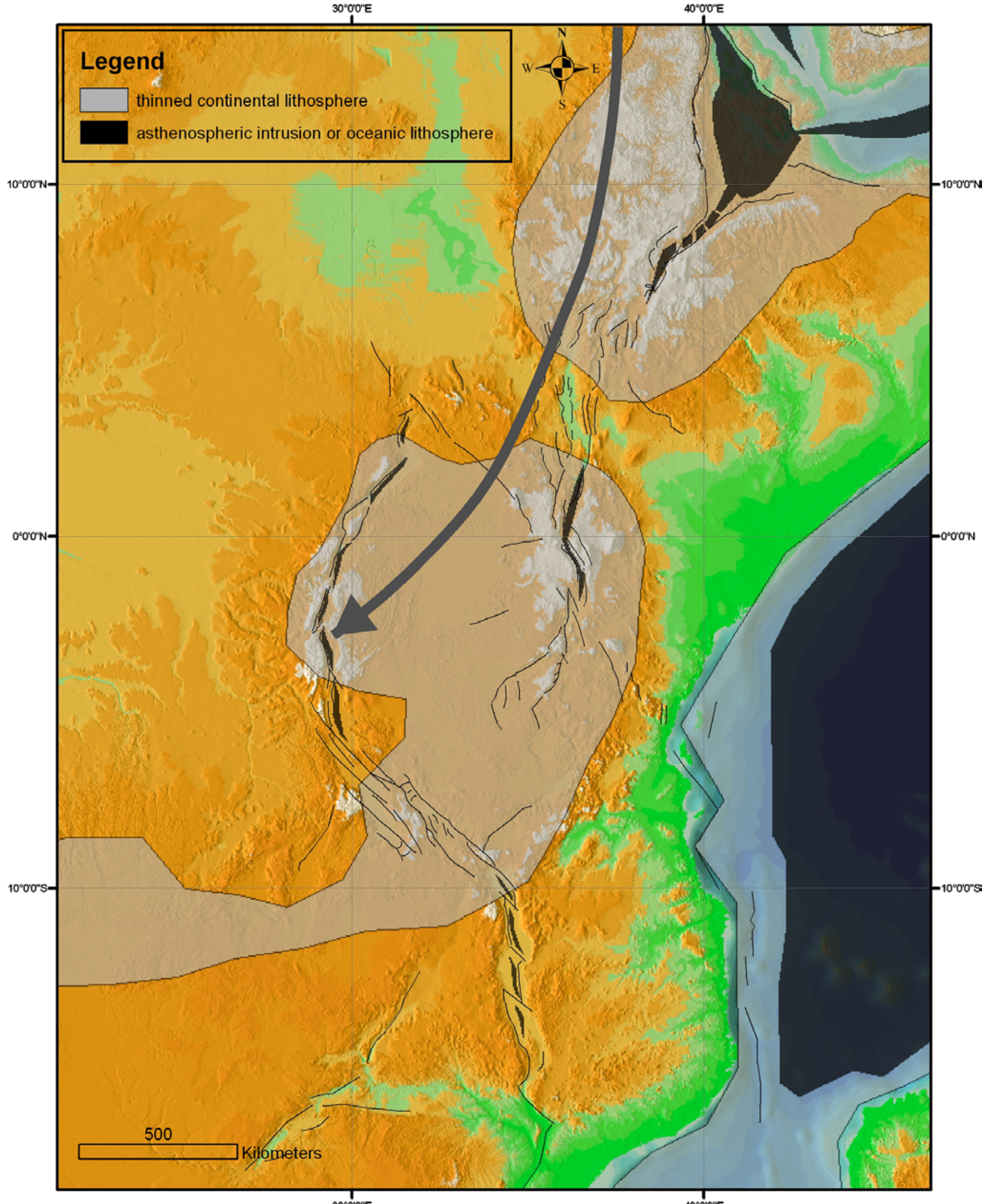

Fig. 10. Tentative lithospheric map of eastern Africa. Curved arrow presents the hypothesis of the late Oligocene to present-day trajectory of a hot-spot core (Bonavia et al., 1995).

consideration of the thermally uplifted shoulders and associated tectonically subsiding graben basins. The hypothetical pattern of asthenospheric intrusions or oceanic lithosphere is that of discontinuous bodies, far narrower than those related to the opening of the Red Sea, Afar and Gulf of Aden.

On the surface, the main tectonic features are normal faults, but there are strike-slip, oblique-slip and sometimes reverse faults. Extension produces widespread open fractures, comprising tension gashes. Most of the fractures are syn-depositional, and when volcanism occurs, it is closely related to the tectonics. Major fault characteristics are shown by an example, the 'split crater' (Jutz and Chorowicz, 1993), located in Kenya (Fig. 11). A non-eroded, Quaternary volcano is affected by faults. The main fault in the middle of the crater has three throw components: a vertical component, expressed by offset of the crater rim (v in Fig. 11B and C), showing that the fault is extensional with eastern compartment down; a horizontal strike-slip component (hsl in Fig. 11A); a horizontal transversal component (ht in Fig. 11A–C), showing that the fault is open with a volcanic plug in the opening (Fig. 11A). Almost all the rift faults have three throw components, with various proportions.

When the vertical throw component is dominant, the faults are considered to be normal. They generally are interpreted to be listric (Bosworth, 1985; Bosworth et al., 1986; Gibbs, 1990) and abruptly connect in depth with low-angle detachment levels (Fig. 12A). The largest faults are along one border of the graben basin. They have tens

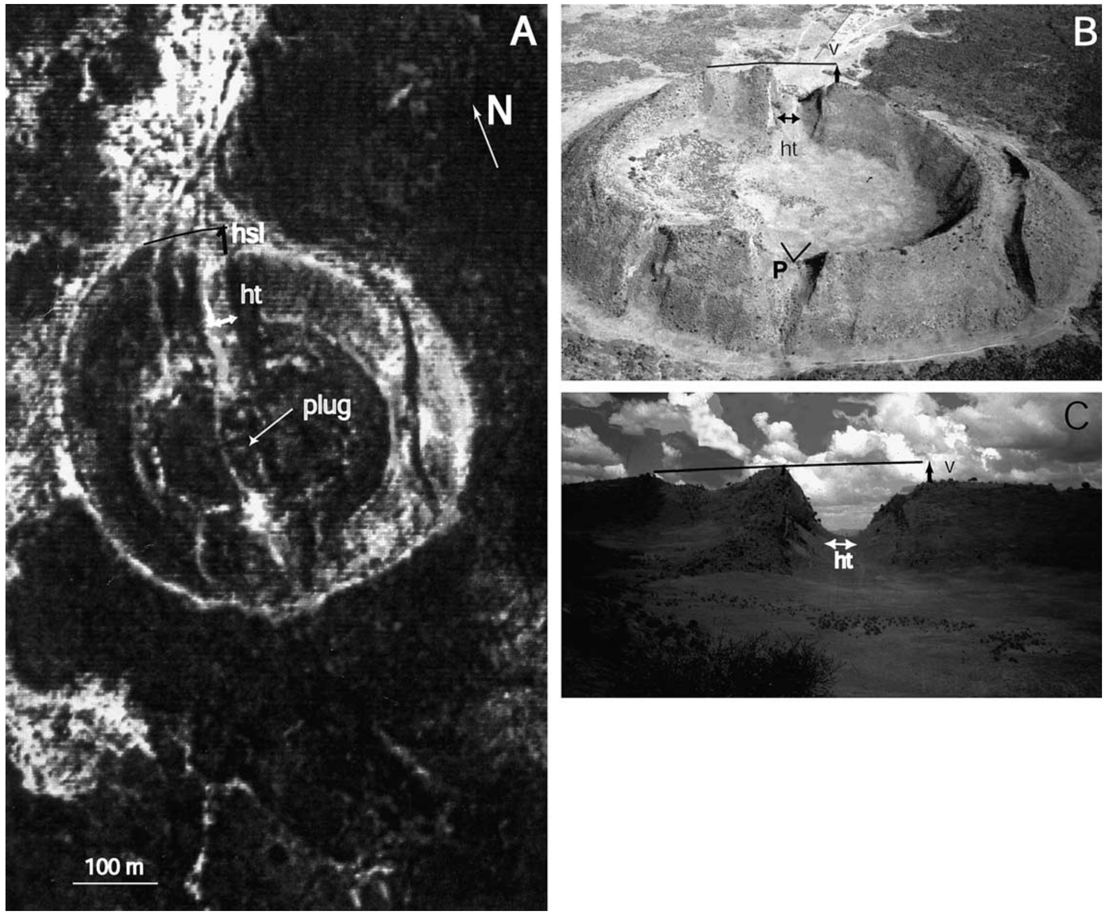

Fig. 11. Split crater, showing three components of fault throw: hsl in A, horizontal strike-slip component; ht (undetermined) in A, B and C, horizontal transversal component; v in B and C, vertical component. Location in Fig. 15.

of km in length and several kilometers of vertical throw (Fig. 3). They dip ∼65° at the surface, are generally curved in plan view, penetrate the middle-lower crust and connect with the low-angle ductile–fragile transition zone at depth of 20–30 km. Seismicity occurs throughout the depth range 0–30 km. For a given graben segment, there is only one of these major normal and detachment faults, which generates a general asymmetric roll-over structure. Throw of the other normal faults is generally much smaller (hundreds of meters or less). They bound tilted blocks whose width depends on the thickness of the series affected by the faulting. Rhomb box faulting and strike-ramps structures (Griffiths, 1980) are related to normal faults having noticeable oblique-slip component. Progressive syn-depositional block tilting is responsible for fan-like pattern in seismic profiles (Fig. 12B). Listric faults may be open in their upper part, and the gap is then filled with sedimentary breccias (Fig. 12C). Subsequent movement of the fault may transform the sedimentary breccias into tectonic breccias, similar to a mylonite. Volcanoes along normal faults are infrequent; it seems that normal faults do not in general offer openings that would permit magma ascension.

Transcurrent faults, with dominant strike-slip component, can run over long distances (more than 500 km), as is the case in central Tanganyika area (Fig. 7). In some places, folds occur related to the strike-slip faults (Fig. 12D and E) and, similarly to the extensional deformations, they are syn-depositional. Normal and strike-slip faults sometimes interact to form spoon faults (e.g., Livingstone fault, Fig. 9).

In the case of tension fractures, the horizontal transversal throw component predominates. These fractures are widespread in rift zones, but are sometimes difficult to identify because they look like valleys (Fig. 13), with floors made of breccias and olistolite fill (Chorowicz et al., 1994; Korme et al., 1997). Tension fractures are first to form during deformation, and can be immediately filled by breccias or magma.

Other types of open fractures are found in the EARS. Eruptions of the Rungwe volcano occurred in relationships with activity of NW-striking faults bounding the Rukwa basin (Ebinger et al., 1989). Updating the mapping from Landsat imagery (Fig. 9) suggests that the volcano formed in a releasing bend relay (accommodation) zone, belonging to the Tanganyika–Rukwa–Malawi dextral fault zone. In other places, tail-crack and horse-tail structures can be found at the termination of oblique-slip faults, where they are responsible for the occurrence of volcanoes and

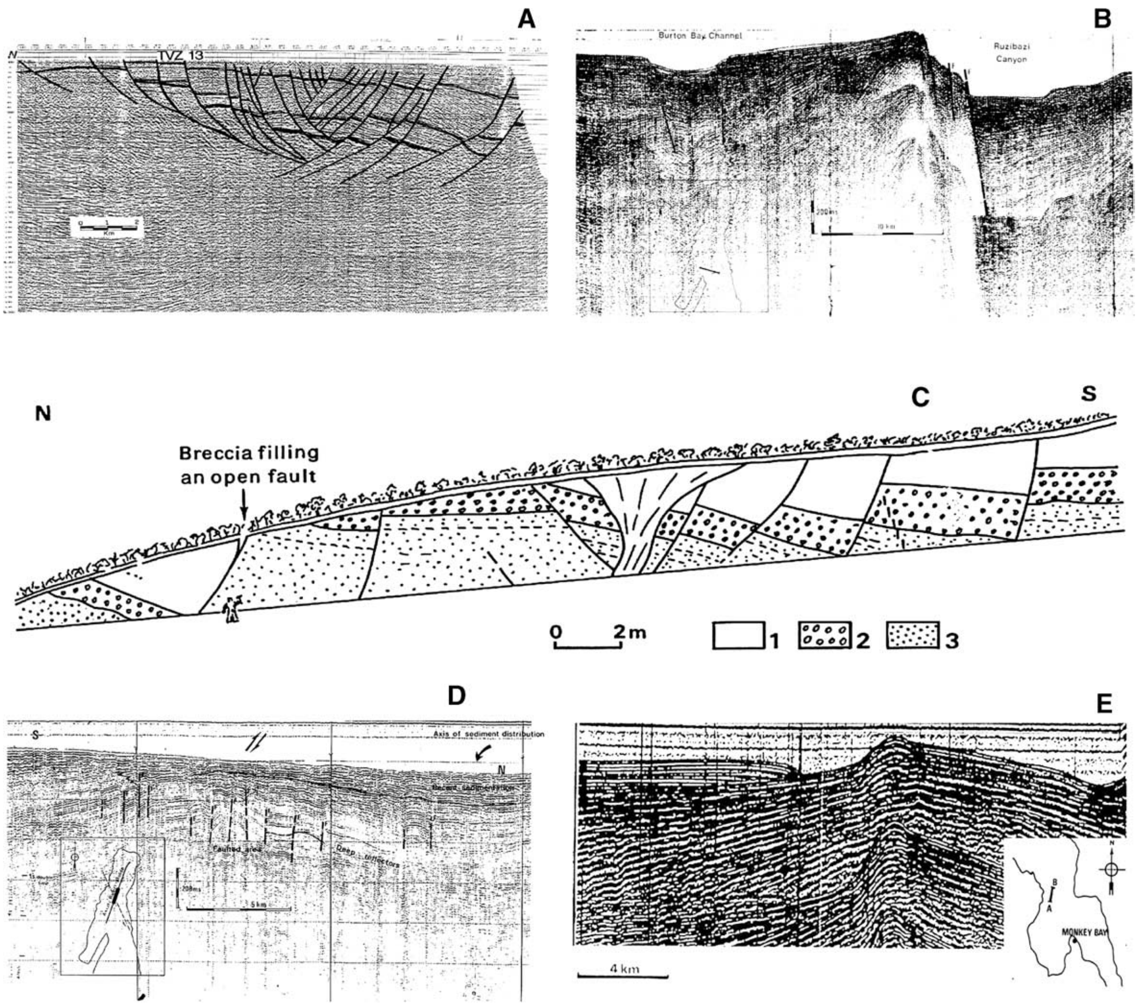

Fig. 12. Examples of fault patterns in the EARS. (A) Seismic profile across Lake Rukwa, showing listric faults that connect at depth into a decollement level (Kilembe and Rosendahl, 1992). (B) Seismic profile across Lake Tanganyika showing progressive block tilting (original, project PROBE, 1986). (C) Cut along road in Ruzizi valley of north Tanganyika plain, showing tilted blocks bounded by listric normal faults. One of the faults is open in its upper part, filled with breccias. (D) Seismic profile in northern part of Lake Tanganyika, showing syn-depositional anticline. This anticline is related to numerous strike-slip faults in central part (original, project PROBE, 1986). (E) Seismic profile in southern Lake Malawi (Nyasa), showing anticline with strike-slip mechanism and normal throw component in its central part (original, project PROBE). See locations of these sketches in Fig. 7.

calderas. A good example is the Menengai caldera, which is developed at the southern end of a complex N-striking fault zone (Fig. 14). An arched crack intruded by volcanic rocks is developed west of the fault end, indicating, in the frame of the tail-crack model, that the N-striking zone of normal faults has a dextral strike-slip throw component. Another example is that of the Awasa calderas (Chorowicz et al., 1994) that have formed at the southern end of the main Ethiopian rift, demonstrating a right lateral throw component along the NNE-striking eastern bounding fault (Fig. 4).

6. Volcanism

6.1. Distribution and types of volcanism

Cenozoic volcanism in the EARS is widespread in the north—especially eastern branch, but sparse in the south (Fig. 15). In other rift systems of similar length, such as the West European rift, volcanism is much scarcer. Abundant volcanism in Northeast Africa is related to plume occurrence (Schilling, 1973; Schilling et al., 1992; Keller et al., 1994). The scenario of north to south migration of plume activity (Bonavia et al., 1995) is consistent with the thin lithosphere shown by a large N–S trending gravimetric negative Bouguer anomaly in eastern-central Africa (Wohlenberg, 1975b; Girdler, 1978; Simiyu and Keller, 1997), and explains the differences in volcanism between western and eastern rifts, and north or south.

The volcanoes are rooted on open fractures such as tension joints, tail-cracks or tensional releasing bends (Korme et al., 1997). Off-axis central volcanoes, such as Mount Kenya are not easily explained (Bosworth, 1987), except perhaps if the opening of large tension fractures happened by reactivation of Precambrian zones of weakness, in some places apart from the main line of the rift valley. This may

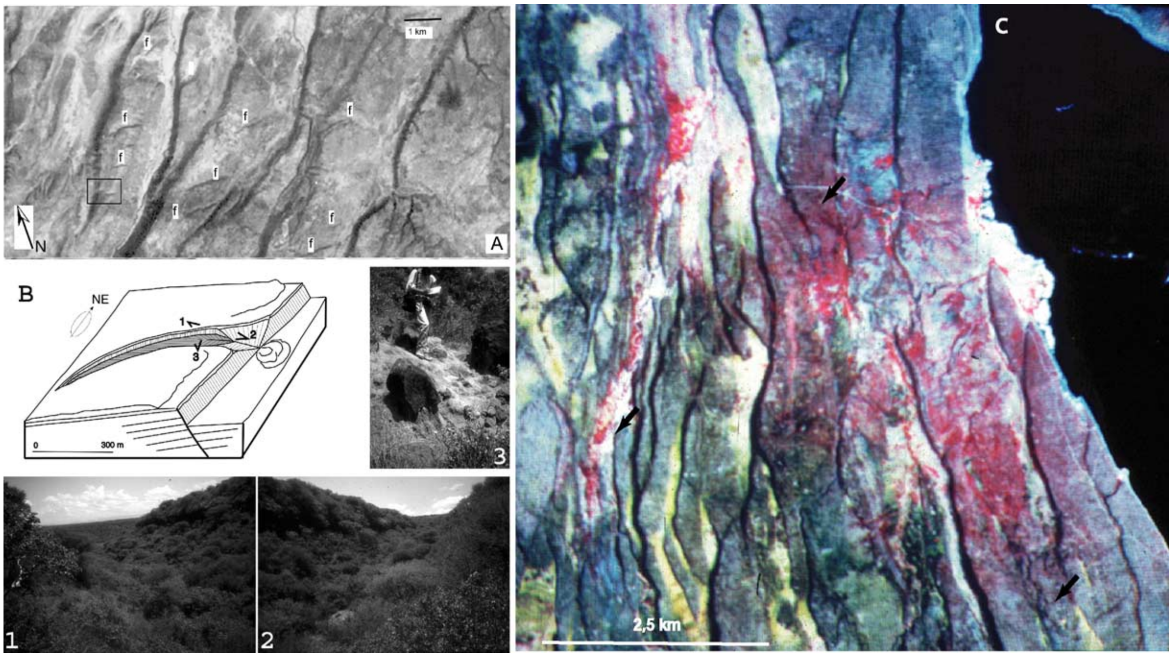

Fig. 13. Extract of a SPOT image acquired over Ethiopia (location in Fig. 4), showing phonolite blocks tilted to the west, bounded by normal faults and affected by tension fractures (Chorowicz et al., 1994). (A) E-striking small valleys (f), oblique to the main downgrading slopes, are tension fractures. One of them (small rectangle) is drawn in B. (B) Three-dimensional sketch of a tension fracture valley. 1, 2 and 3 are field photos of the valley. (1) V-shaped downstream valley, (2) flat-bottomed upstream valley, (3) sand, conglomerate and block fill of the valley in location 3. (C) in Kenya, along shore west of Lake Bogoria (location in Fig. 8), blocks tilted to the east, bounded by normal faults, and ENE-striking tension fractures (arrows).

be the case for the Sagatu ridge, east of the Ethiopian rift, a 2 km wide fissure system (Mohr and Potter, 1976). Mounts Elgon and Kilimanjaro (Downie and Wilkinson, 1972) are situated in strike-slip zones where releasing bends of pull-apart openings are likely to form.

The magmas are alkaline to hyperalkaline, and typically evolved from continental tholeiites rich in incompatible elements, through alkalic products to transitional magmas relatively low in incompatible elements (Mohr et al., 1972). There is a predominantly alkali basaltic source with contributions from a lower crustal protolith (Hay et al., 1995). According to Macdonald (1994), all major sequences show evidence of extensive polybaric fractionation within the upper mantle and lower crust. Inter-crustal fractionation has commonly been accompanied by assimilation and the development of silica-(sometimes over-)saturated liquids. Fractionation at high crustal levels has generated a wide range of mugearitic, trachytic and phonolitic magmas. Crustal anatexis has locally resulted in the formation of peralkaline rhyolites.

Magmatism is linked with asthenospheric ascent, either under regional plume conditions, or local rift-related lithospheric mantle thinning. According to Kampunzu and Lubala (1991) and Mohr (1992), during lithospheric extension, largely related to dyking, diapiric ascent of a lherzolitic asthenospheric wedge leads to the generation of basaltic melts. Partial melting may occur at all levels within the diapir, producing magmas that are tholeiitic at shallow level, transitional at intermediate levels and alkaline at deeper levels.

6.2. Age of volcanism

The plume responsible for this volcanism occurred first in the lake Tana region and may have subsequently migrated southward together with northward drift of Africa. The rate of migration of the onset of magmatism from Ethiopia to Tanzania is similar to the rate of migration of the African plate over the same period (George et al., 1998). This model is still questionable and other views have been proposed, e.g., (1) timing of plume evolution may have to take into account plume and related trap volcanism in southern Ethiopia at 45–35 Ma (Ebinger et al., 1993a,b); (2) Africa may have been essentially stationary during the past 30 Ma (Burke, 1996).

Volcanism began during the Oligocene in the Ethiopian zone. $^{40}Ar/^{39}Ar$-isotope geochemistry indicates that Ethiopian flood basalt erupted very rapidly (1 Ma) at 30 Ma (Hoffman et al., 1997). In southern Afar, other volcanic events occurred at 14–11, 11–10, 9–7, 5–4 and post 1.6 Ma (Mohr, 1978; Berhe, 1986). In the Afar–MER transition, volcanism was renewed at 10 Ma; bi-modal basalt-rhyolite volcanism began at 7 Ma, and continued until recent time (Chernet et al., 1998).

More to the south, in the northern Kenyan rift, basalts and rhyolites were emplaced at 33–25 Ma (Oligocene), then nephelinites and phonolites at 26–20 Ma, and highly hyperalkaline basalts at 15 Ma in eastern periphery of the region (Zanettin et al., 1983). At beginning of the Pliocene, trachytic, phonolitic, nephelinitic rocks, and basaltic volcanism were accompanied with some rhyolitic

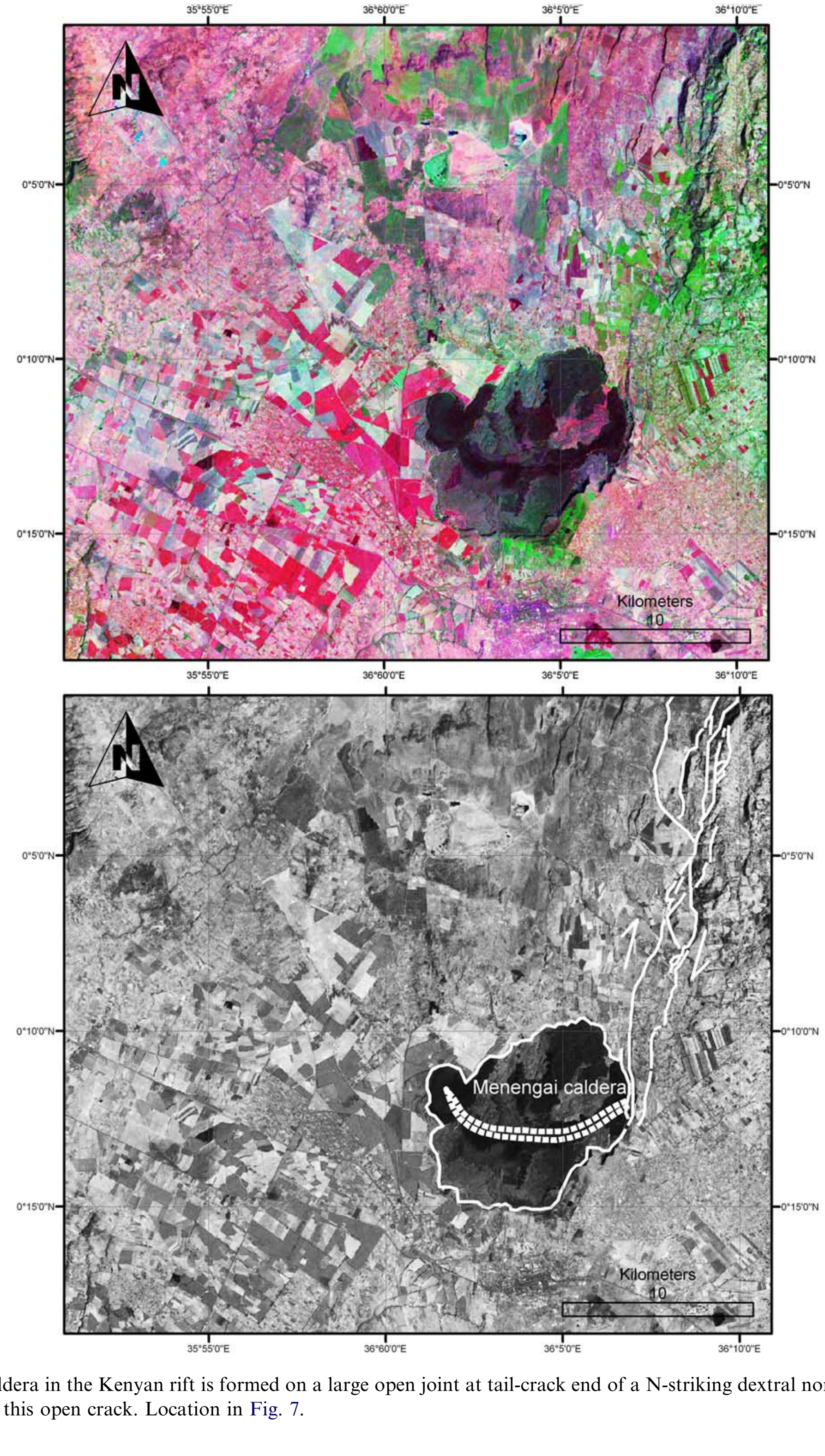

Fig. 14. The Menengai caldera in the Kenyan rift is formed on a large open joint at tail-crack end of a N-striking dextral normal oblique-slip fault zone. The lavas are issued from this open crack. Location in Fig. 7.

activity. In the late Pliocene and Quaternary, volcanism was trachytic in the rift valley floor, basaltic to the east (Baker et al., 1971; Woldegabriel et al., 1990). In the central Kenyan rift, according to Hackman et al. (1990) and Smith (1994), 1 km thick Samburu flood basalts erupted between 20 and 16 Ma, followed by large volumes of trachyte and phonolite between 5 and 2 Ma, and carbonatite and nepheline-phonolite volcanoes around 1.2 Ma. However, as there are not yet much $^{40}Ar/^{39}Ar$ dates, it is still necessary to be careful about timing.

Data concerning the western branch were summarized by Kampunzu et al. (1998). In the north of the branch

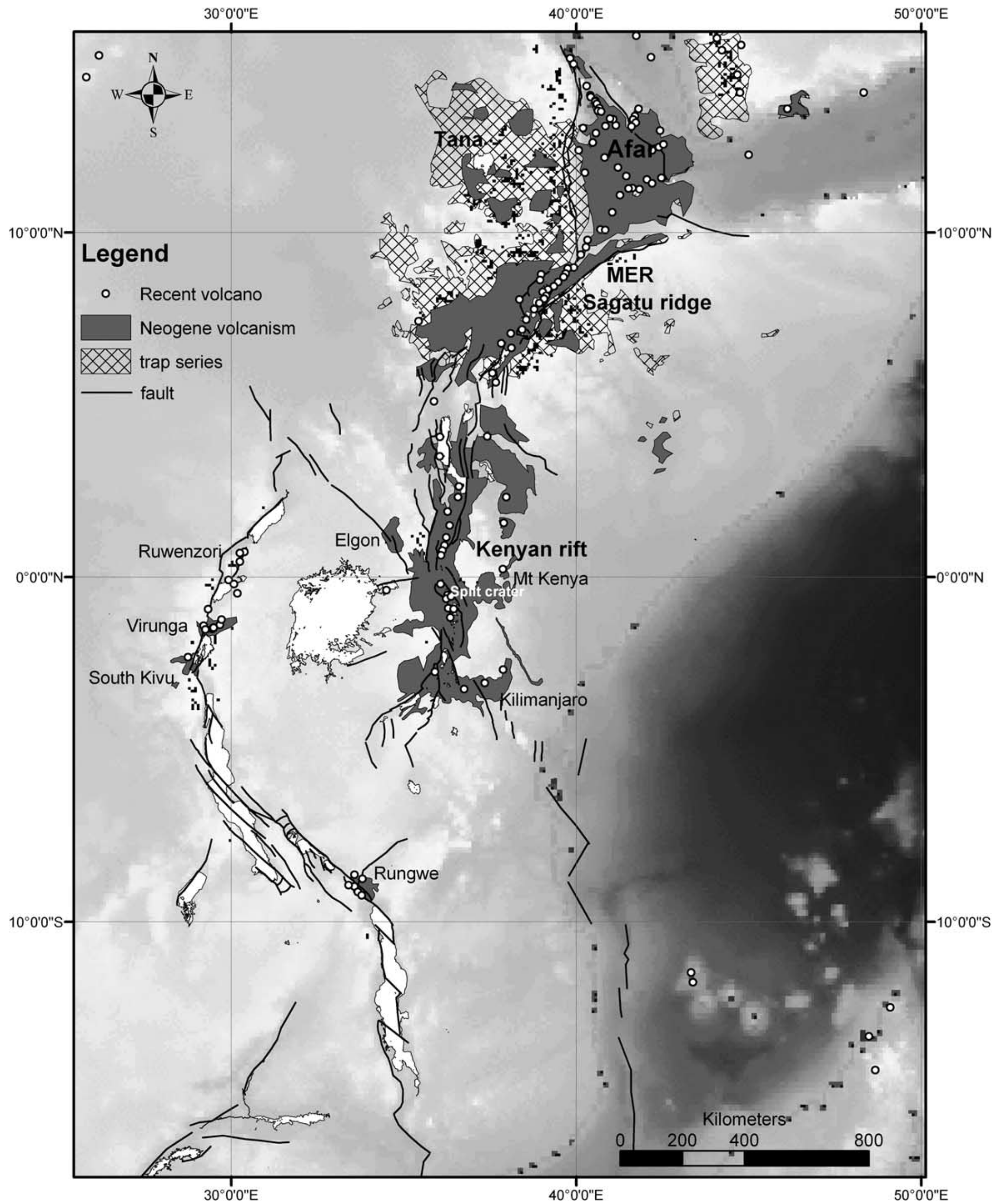

Fig. 15. Time and spatial distribution of volcanism in the EARS (source, USGS database).

(Ruwenzori, Fig. 15), volcanism began at 12 Ma. In the Virunga, massif fissure volcanism started between 11 and 9 Ma, and after a long period of quiescence (9–3 Ma), large Pliocene-Pleistocene central volcanoes were formed, made of highly under-saturated, potassic, ultra-alkaline lavas. In the south Kivu region, volcanism commenced at 8 Ma with tholeiites, followed by sodium alkaline lavas. The southernmost volcanic massif in the western branch is the Rungwe volcano, whose activity began at 8.6 Ma (Ebinger et al., 1989).

7. Kinematics

7.1. Directions of movements

The EARS is a lithospheric opening in the African continent, which in terms of plate tectonics results from the divergence of large, regional-scale blocks. Movements can be deduced from focal mechanisms of earthquakes, borehole breakout in exploratory wells, structural analysis, strike of dykes, elongate volcanoes, volcanic chains, tension fractures and neo-formed normal faults (e.g., Shudofsky, 1985; Morley, 1988; Bosworth et al., 1992; Chorowicz et al., 1994; Haug and Strecker, 1995; Korme et al., 1997). The first structural analysis data published (Chorowicz and Mukonki, 1979) argued for NW–SE movements in the western branch. Other complementary data collected over the whole rift system were coherent with this interpretation (e.g., Chorowicz et al., 1994; Chorowicz and Sorlien, 1992). This direction of extension was also argued by Sander and Rosendhal (1989), Specht and Rosendahl (1989), Versfelt and Rosendahl (1989), Morley et al. (1992a,b) and Scott et al. (1992).

However, different analyses and interpretations concluded to NE (Katz, 1987) or E–W extension (Ebinger, 1989a; Bosworth et al., 1992). From focal mechanisms of earthquakes Fairhead and Girdler (1972) have proposed E–W to NE–SW extension. A two directions model of extension was presented for the Ethiopian rift, with NW–SE extension before the beginning of the Quaternary, and E–W extension after (Boccaletti et al., 1998). A more complex pattern was defended by Bosworth and Strecker (1997), from analysis of the paleostress fields, with rift extensions in Kenya varying from E–W to ENE–WSW and NE–SW direction between 12 and 0.6 Ma, to turn since 0.6 Ma to the NW–SE direction. According to Delvaux et al. (1992), Strecker et al. (1990), Bosworth et al. (1992) and Van Der Beek et al. (1998), several changes occurred in the kinematic regime of the EARS during the late Cenozoic, including an episode of compression, which they explained by changes along the boundaries of the African plate. In the whole, these various interpretations are based on local paleostress analysis, but generalization of paleostress data to the overall plate movements is not obvious. It is necessary to make a clear distinction between local paleostress fields, and movements at regional scale (tectonic transport direction). There are many opportunities for a main regional stress field to suffer local deviations, e.g. transcurrent faults, local gravity effects, local volcanoes, plume effects. The semi-automatic deciphering of different paleostress phases (Delvaux, 2001) may be hazardous. In fact, the main deformations in the EARS being along the major faults, it is necessary to privilege what has happened along the major, mapped faults (Chorowicz et al., 1998) rather than drawing conclusions from minor movements on small fractures, in blocks that may have rotated, or suffered their own strain depending on local conditions. This question of directions of movements has not reached a general agreement. It is not yet clear what paleostress fields tell about regional movements, and more observations are needed all over the EARS.

It occurs that two types of movements affect the same faults, alternating through time and space (Chorowicz, 1990). (1) NW–SE drifting movement of large continental blocks is coherent with the overall geometry of the rift system. In general, the moving apart responsible for the rift opening is then oblique to rift trend, except for rift segments that strike NE, the main Ethiopian rift and the northern end of the western branch. (2) Locally, eventual movements can be triggered by high relief that is responsible for gravity gliding effects, especially along the major border faults, and then they tend to occur at right angle to the fault strike (Chorowicz, 1990). E–W local extension is described in many places especially in the eastern branch that comprises the N-striking Kenyan rift (e.g., Strecker et al., 1990; Bosworth et al., 1992; Bosworth and Maurin, 1993). Smith and Mosley (1993) have proposed that body forces should be taken into consideration in the Kenyan rift. Interference between two types of fault movements is also a conclusion of Clifton et al. (2000), showing with clay models that tensional stresses at the rift margin are modulated and reoriented by a secondary stress field related to a change in boundary conditions, resulting in the formation during oblique rifting of two distinct sub-populations of faults. According to these views, there is a changing pattern of stress and deformation within a constant kinematic scheme. Local varying movements may alternate or interfere with stable drifting movements, with a tendency of the larger faults to be more influenced through time by the local gravity-driven deformations, giving the impression of different tectonic phases with different directions of extension. It would be surprising in areas affected by important thermal uplift, oblique extension and major rift collapse forming large border faults, that body forces would not be responsible for noticeable part of the local deformation.

8. Transform and transfer fault zones

8.1. Transform fault zones

NW-striking large faults linking rift basins are parallel to the main extensional movement as interpreted in the preceding chapter, and consequently can be regarded as intracontinental transform fault zones belonging to the rift system (Fig. 16). First presented by Chorowicz and Mukonki (1979) and Kazmin (1980), this model was also supported by Tiercelin et al. (1988), Chorowicz (1989), Daly et al. (1989), Kilembe and Rosendahl (1992), Wheeler and Karson (1994). The model predicted that these NW-striking fault zones have a major strike-slip component, right- or left-lateral according to their place in the model. Observations along these fault zones, presented in the next section are significantly in agreement with the model.

The Tanganyika–Rukwa–Malawi fault zone connects two main segments of the western branch (Chorowicz et al., 1983; Chorowicz, 1984). In terms of plate tectonics, this is an exemplary right-lateral intracontinental transform. It forms a complex left-stepping en echelon fault and fold zone, including a releasing-bend opening responsible for the occurrence of the Rungwe volcano (Figs. 9 and 16). Folds affect the basement and are developed in narrow stripes between the dextral strike-slip faults (Fig. 17). This explains why some parts of the border lands of the Tanganyika–Rukwa–Malawi zone form low plains (synclines)

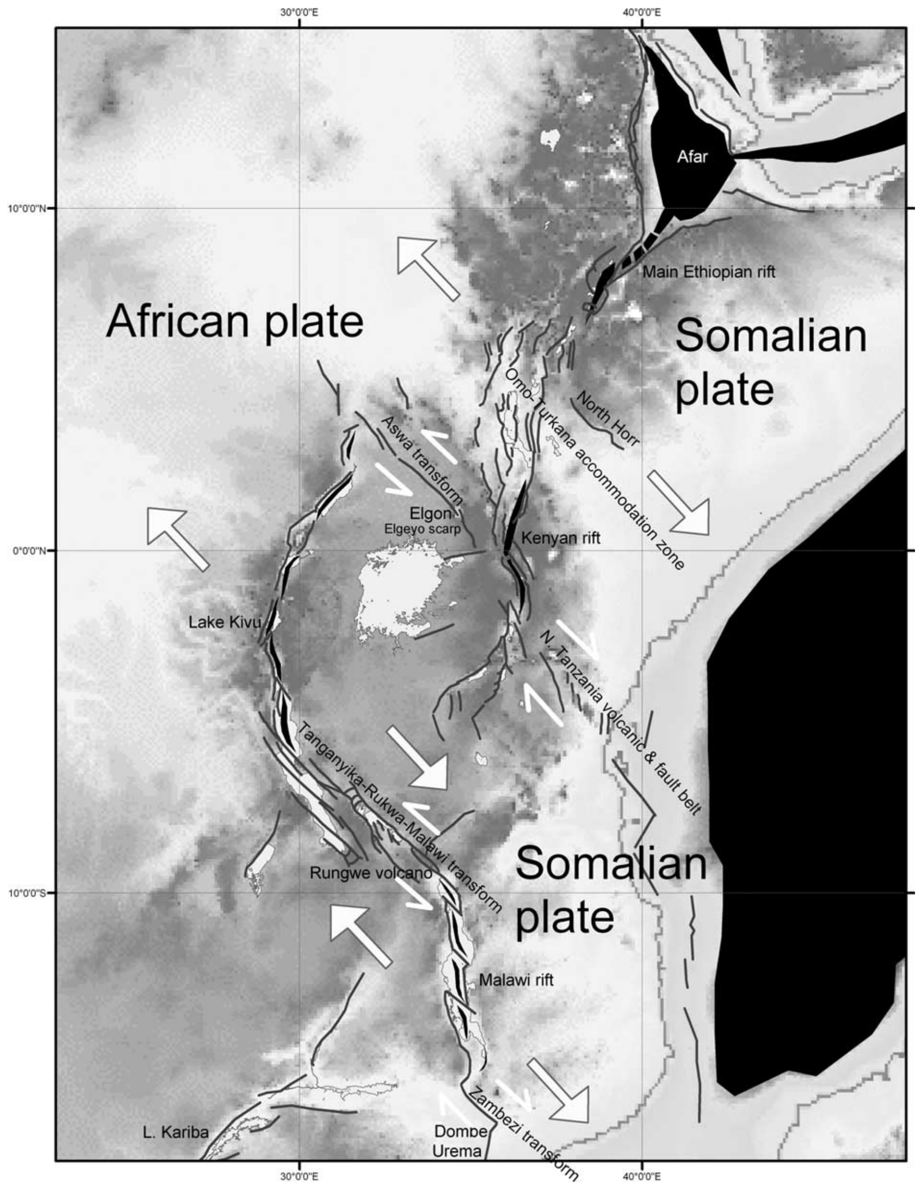

Fig. 16. On-going individualization of the Somalian plate in Eastern Africa. Asthenospheric intrusions (black polygons) show already open lithosphere. White arrows show direction of relative divergent movement.

instead of the usual high shoulders, and other parts are high relief, such as the Malahi highs (anticlines). In this interpretation, the Rukwa basin, interpreted by Kilembe and Rosendahl (1992) to be a pull-apart basin, is rather a syncline.

The other NW-striking zones are not all so typical, and can be considered as partly diffuse transforms, in which strain is distributed over wide and long NW-trending areas, with numerous fractures in the crust, including strike-slip faults (Fig. 16). They can be regarded as links between the narrow elongate zones of asthenospheric material intruded into the continental lithosphere.

The Aswa transform is partly diffuse but comprises several large left-lateral NW-striking faults, among which some are responsible for the strongest earthquake in Africa (Gaulon et al., 1992). The transform includes also the Elgon volcano, probably in a left-stepping fault relay zone, and the left-lateral strike-slip (Chorowicz et al., 1987) Elgayo fault scarp (Fig. 8). At the intersection of the transform zone and the eastern branch, the Kenyan rift turns in strike to align with the transform and constitutes the central Kenyan rift (Fig. 16). The transform zone continues SE-ward to form the volcanic and fault belt of northern Tanzania, linking the eastern and the southeastern branches of the EARS. According to Fairhead (1980), the gravimetry, characteristics of the faulting, seismicity and geothermal activity of the belt represent the early stages of a right-lateral transform fault. The difference between the crustal extension to the north and south of the volcanic and fault chain is considered to be taken up along the 200 km length of the chain by en echelon faulting and fissuring, the latter providing routes for magma to reach the surface.

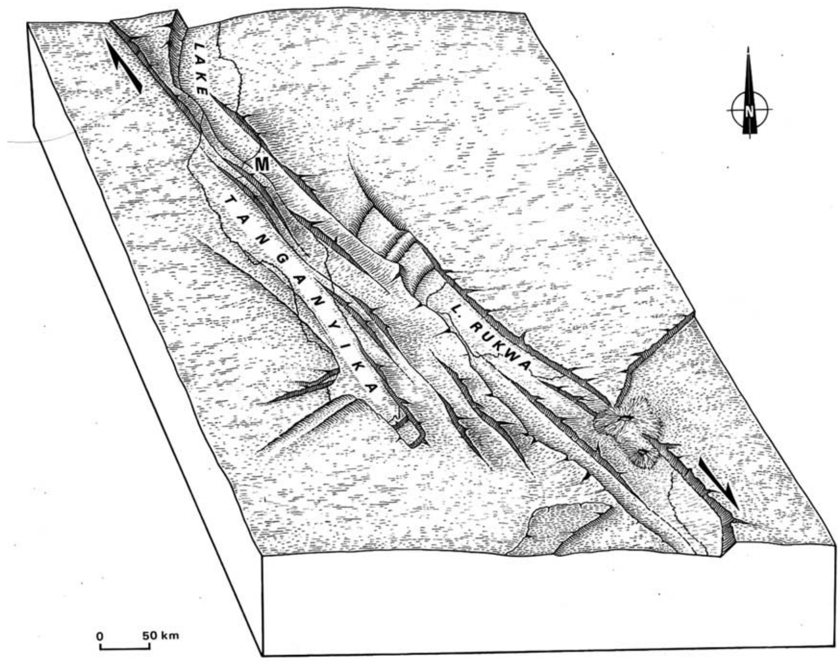

Fig. 17. Fault and fold zone of the Tanganyika–Rukwa–Malawi segment of the EARS. Folds are developed in stripes between left-stepping en echelon dextral strike-slip faults. This pattern of folds explains why some segment border areas of the Tanganyika rift form low plains instead of the usual high shoulders.

The Zambezi transform links the southern end of the Malawi rift with the line formed by the Dombe and Urema graben basins (Chorowicz and Sorlien, 1992). This structure remains insufficiently studied; according to the model, it might be right-lateral.

In this model, the major rift segments, namely the eastern branch and the two main segments of the western branch, linked by intracontinental transform zones, form a unique breaking line in Eastern Africa that resembles oceanic ridges linked by oceanic transform faults (Fig. 16). The EARS can then be taken as the beginning of the opening of an ocean (Chorowicz, 1983), between two large continental blocks drifting apart in the NW–SE direction. But this is an immature plate boundary, separating the main African plate and the not yet well individualized Somalian plate. These continental transform fault zones have been supposed to continue to propagate and become young oceanic transforms faults, as the rift develops until oceanic lithosphere appears (Daly et al., 1989), but however there is little evidence in passive margins that syn-rift transform and transfer faults (or accommodation zones) commonly evolve directly to oceanic fracture zones (Bosworth, 1994; Fantozzi and Sgavetti, 1998). The south-eastern branch seems a secondary process, which links the EARS opening with major structures to the east. The cumulate quantity of movement concerning the eastern and western branches is badly known and variable from 40 km in the Kenyan rift (KRISP, 1991; Morley et al., 1992b; Hendrie et al., 1994) to 13 km in the western branch (Morley, 1988; Ebinger, 1989a). This is small for a more than 6000 km long rift system, and consequently the Africa/Somalia Eulerian pole is badly constrained. For Delvaux (2001), the late Quaternary Eulerian pole is in the Afar, and for Asfaw (1992) at 1.5 °S, 29.0 °E (northern tip of Lake Kivu). In fact, the Eulerian pole likely lies in the south near Limpopo valley (see Section 9.1).

8.2. Transfer faults and accommodation zones

The transform fault zones run mainly outside the rift basins and link distant rift branches. The concept of transfer fault (Gibbs, 1984) applies to a transverse fault zone inside a rift basin, and corresponds to a change in fault geometries of the same age across, extension being transferred from one fault system to the next. The concept of accommodation zone was introduced by Bosworth (1985) and applies to an area where a change of fault geometry occurs, which may be of different ages. These two concepts are quite similar, and the consideration of the ages is not critical. Trying to make clear the difference, it can be said that a transfer fault is parallel to the movement of extension in a basin, whilst an accommodation zone is made of faults segments that are mostly not parallel to the movement. Both concepts can be illustrated in the EARS.

The concept of transfer fault was applied in the EARS first by Chorowicz (1989). In the Malawi rift case (Fig. 9), the main boundary fault changes side across transverse NW-striking faults (Chorowicz and Sorlien, 1992; Scott et al., 1992). This is accompanied by reversal in basin asymmetry. The result is a succession of different basins, each about 70 km long, separated by buried basement shoals or topographic highs corresponding to strike-slip faults paralleling the movement of divergence. These structures do not extend outside the 40–60-km-wide rift basins across the uplifted flanks. A transfer fault zone accompanied with reversal in rift asymmetry can be theoretically represented in Fig. 18. Such transfer faults characterized by their NW-strike and location inside the rift basins, are frequent in the EARS, and more generally separate segments that differ in the geometry of tilted blocks.

The concept of accommodation zone can be illustrated by the southern end of Lake Kivu (Fig. 7). It had been previously considered as a south-Kivu short transform zone (Chorowicz et al., 1979). NE-striking Pan-African faults were reactivated as reverse faults by NW-trending movement that produce local compression. Ebinger (1989a,b) confirmed the importance of these NE-striking faults, and interpreted them to be submitted to strike-slip deformation, because inferring NE-trending movement.

Another example of accommodation zone is the Omo-Turkana faulted area, 550 km long and 350 km wide, which develops between the southern end of the main Ethiopian rift and the northern end of the Kenyan rift (Fig. 16). Many half-graben basins strike NNE (de Heinzelin, 1988), but there is no distinct rift in the area, and consequently no narrow elongate intrusive asthenospheric body is expected. This area can also be regarded as a diffuse transform zone (Chorowicz, 1992), with only one fault striking NW, the North Horr fault (Atnapu and Bonavia, 1991), reactivating the Cretaceous-Paleogene Anza rift (Bosworth and Morley, 1994).

The transform and transfer fault zones, together with the accommodation zones segment the EARS along its length into 50–100 km-long individual basins, suggesting a relationships with regularly segmented oceanic rifts.

9. Time and space evolution

9.1. Rift propagation

First rift manifestation is the formation of open fractures at ~30 Ma in the Afar and Ethiopian plateau, due to hot spot activity, responsible for emission of trap lava flows (Hoffman et al., 1997; Mège and Korme, 2004), accompanied by initiation of three converging graben structures, forming a triple junction at lake Tana (Chorowicz et al., 1998). Rifting in the Gulf of Aden began at 29.9–28.7 Ma, and in the southernmost Red Sea at 27.5–23 Ma (Hughes et al., 1991). The Afar depression seems to have formed later in the Miocene. At 20 Ma, first volcanism appeared far more to the south at the future triple junction between the northern Kenyan, central Kenyan and Nyanza rifts (Pickford, 1982; Fitch et al., 1985). The main Ethiopian rift, located between these two nucleations, commenced after 11 Ma (Wolfenden et al., 2004), later than early Miocene volcanic events in the Turkana region (Bellieni et al., 1981).

In the western branch, the first lavas erupted in the Virunga at 12.6 Ma (Bellon and Pouclet, 1980). First S-Kivu sediments where deposited at 10–8 Ma (Ebinger, 1989b). To the north of Lake Kivu, in Lake Albert (Mobutu), after shallow downwarp at ~12 Ma, first major rifting episode was at 8 Ma, and rift shoulders were climatically significant at 4 Ma (Pickford et al., 1993). To the south of Lake Kivu, in Lake Tanganyika (Cohen et al., 1993), the central basin started to subside between 12 and 9 Ma (middle to late Miocene), the northern basin between 8 and 7 Ma (middle late Miocene)—there is a 5 Ma old fauna in Ruzizi

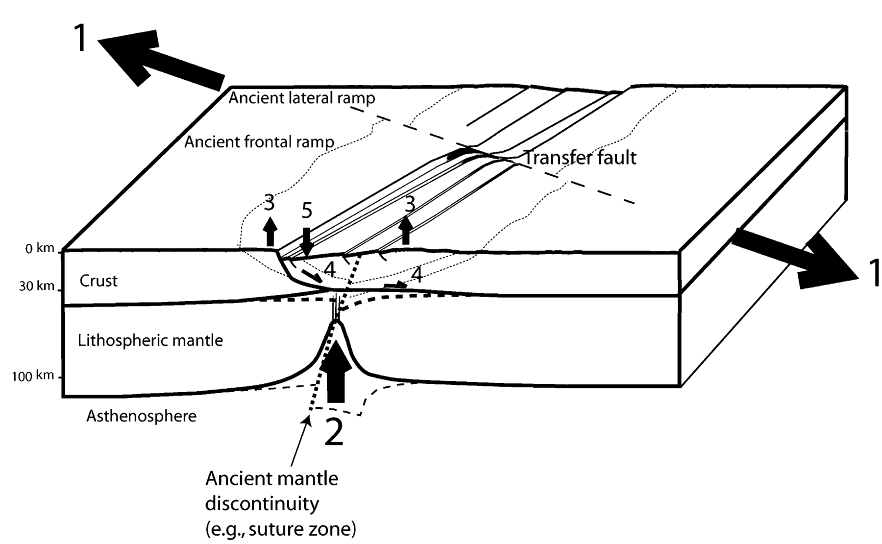

Fig. 18. Theoretical block diagram showing the dynamics of formation of a rift segment including a transfer fault. Geometry of fault patterns changes across transfer fault, with frequent reversal in asymmetry. The transfer fault is inherited from coincidence of divergent motion with predisposed lateral ramp or tear fault in ancient belt. Dotted lines: ancient faults. Striped lines: lithospheric suture structure of collisional belt. (1) divergence of continental blocks; (2) strain focused along former weak suture zone induces local thinning of the lithospheric mantle, and subsequent intrusion of elongate asthenospheric diapir; (3) induced heating of the lithosphere, responsible for wide (hundreds of kilometers) uplift of the area; (4) in the crust, a unique major normal fault abruptly linked in depth to a low-angle detachment fault reactivates deep part of former frontal ramp of ancient belt, and accommodates horizontal extensional movement; (5) this deformation induces a roll-over structure, and related subsidence.

basin—the southern basin between 2 and 4 Ma (Pliocene). The Rungwe volcanic complex is dated 8.6 Ma (Ebinger et al., 1989). The Malawi rift began to subside during the late Miocene at 8–9 Ma (Harkin, 1960; Ebinger et al., 1984, 1987, 1989, 1993a). Major rifting was in northern Lake Malawi (Nyasa) at 6–5 Ma, and the rift increased in length southwards (Flannery and Rosendahl, 1990).

The EARS on the whole propagated southward at a mean rate between 2.5 cm/year (Oxburgh and Turcotte, 1974) and 5 cm/year (Kampunzu and Lubala, 1991; Kampunzu et al., 1998). Rift fracturing and subsidence nucleated at different locations during the late Miocene, first in the Afar and Kenyan rift, in the Virunga and central Tanganyika (late Miocene). From each of these places it propagated along-axis to the north and south to link originally isolated basins (Ebinger, 1989a; Chorowicz, 1992). At present day, the EARS is still propagating southward as shown by distribution of the seismic activity (Fairhead and Henderson, 1977; Fairhead and Stuart, 1982) (Fig. 19). According to Kebede and Kulhanek (1992), the rift activity concerns areas in the south that are not yet more than gentle lows in southern Africa but are zones of weakness in the lithosphere. Several lines of propagation are defined, near Lake Mweru, at Lake Kariba, along the Mozambique coast and more to the south along the Limpopo suture (Brandl, 1986). From this overall pattern, the Eulerian pole between Africa *s.s.* and Somalia is still badly constrained, but should be at least at Limpopo valley or more to the south.

The various rift segments are at different evolutionary stages, depending on the age of rift initiation and distance to the Eulerian pole. For a given age, length of normal faults, width of rift basins, relief of the uplifted rift flanks vary depending not on the age but on the stage of a rift segment. The following stages of rift evolution can be defined on the basis of precise criteria (Chorowicz et al., 1987; Mondeguer et al., 1989; Hayward and Ebinger, 1996). (1) *Pre-rift stage* is characterized by dominance of horizontal

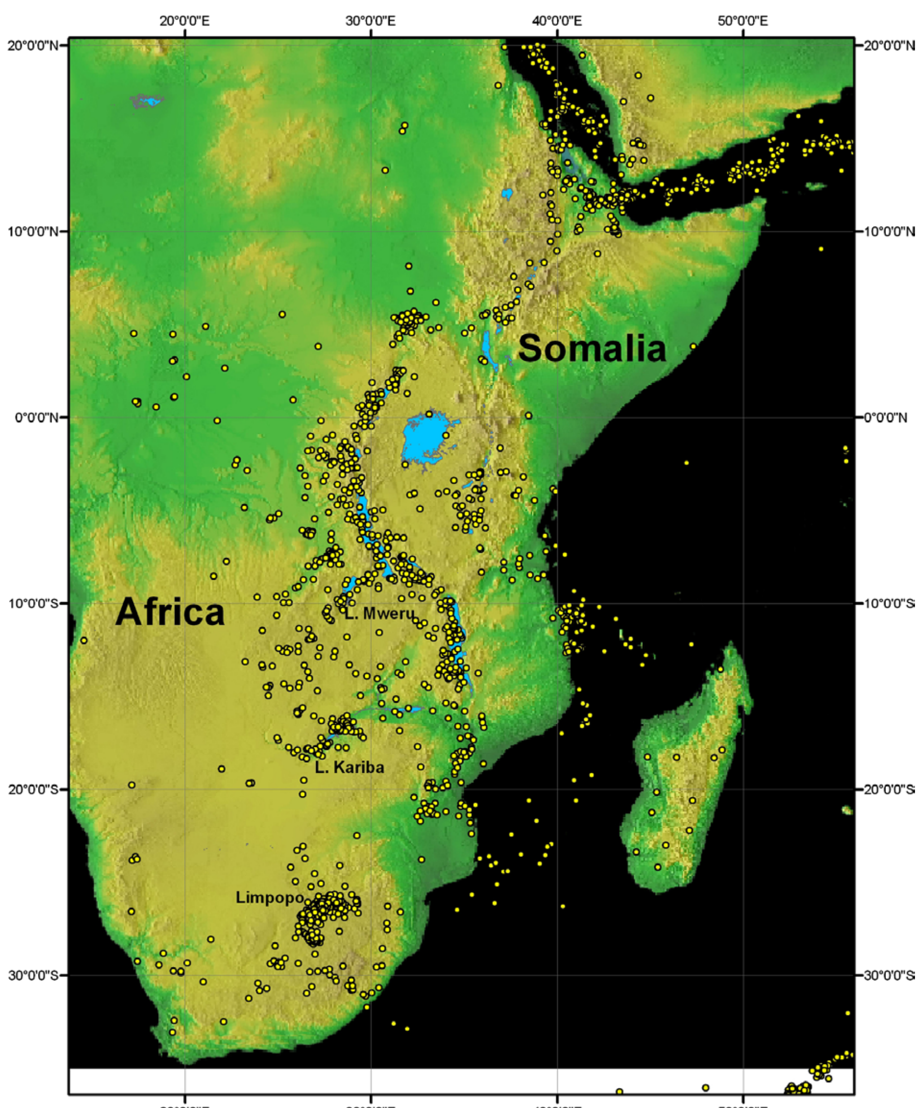

Fig. 19. Distribution of earthquake epicentres (source, USGS).

movements (along a given fault more ancient striations are generally more horizontal according to Chorowicz et al., 1987, and in the youngest propagating southern EARS the focal mechanisms of earthquakes are distinctly strike-slip according to Shudofsky, 1985), responsible for high energy seismicity, formation of en-echelon tension gashes and dense distribution of small-throw strike-slip and oblique-slip faults, with a strike-slip stress-field ($\sigma 1$ and $\sigma 3$ horizontal). If volcanism occurs, it is mainly tholeiitic hyperalkaline. The morphology is that of topographic depressions characterized by swamps and shallow lakes, without noticeable uplifted shoulders. There is not yet a true graben, and the asthenospheric intrusions are discrete. The Limpopo area is at this stage. (2) *Initial-rift stage* is related to noticeable subsident and divergent movements, with frequent earthquakes, lesser fault density, mostly extensional oblique-slip faults, and stress field being tensional ($\sigma 1$ vertical, $\sigma 3$ horizontal). Volcanism, if existing, comprises contaminated alkaline magma. The basins are well defined with roll-over geometry in cross-section, pronounced shoulders, and are composed of box-like half-graben troughs. They are some tens of kilometers in length, 30–40 km in width, and separate from each other by transform, transfer or accommodation zones forming basement highs. The asthenospheric intrusions in the lithosphere are pronounced, responsible for negative Bouguer anomaly. A good example of this stage is the Malawi rift. (3) *Typical-rift stage* has a very well defined rift valley and thick graben deposits, with frequent low-energy earthquakes. Tectonics is dominated by the effects of normal faulting, but most of the deformation is concentrated along the major faults bordering one side of each graben. Regional extensional movements are combined with local tension related to gravity induced deformation. The former shoals are buried; the originally separated individual basins now connected forming longer basins, often occupied by a great lake. The shoulders are high, the negative Bouguer anomaly is steep and narrow, due to asthenospheric intrusion. The northern Tanganyika rift is representative of this stage. (4) *Advanced-rift stage* differentiates from the preceding by a larger regional negative Bouguer anomaly, with a contrasting narrow positive gravity anomaly in the middle, indicating the intrusion of dense material into the crust and considered as the first apparition of oceanic-type material (Baker and Wohlenberg, 1971). There are many large tension fractures giving way to differentiated alkaline volcanism. This situation essentially corresponds to the main Ethiopian rift. (5) *Oceanic rift stage* is that of the Afar, with oceanic crust.

9.2. Relationships with ancient structures

It is well established that break-up of the continental lithosphere is controlled by its preexisting structures, and this applies for the evolution of the EARS (McConnel, 1972; Versfelt and Rosendahl, 1989; Chorowicz, 1992). The EARS being of lithospheric scale, tectonic predisposition forming weaknesses in the lithosphere would be lithospheric scale structures such as former sutures linked to belts, and transform or rift zones.

The EARS avoids Archaean cratons and is mainly incised in the frame of the N-striking Mozambique belts that affected Eastern Africa during the Pan-African events at the end of the Neoproterozoic. Some of the structures had already been active at several time intervals before (McConnel, 1972). The Western Afar margin and the main Ethiopian rift are parallel to two NNE-trending Pan-African lithospheric sutures located in west-central Ethiopia. In the Kenya rift, gravity anomalies attest to rift but also to suture structure in the lithosphere (Nyblade and Pollack, 1992). In north-western Kenya, a Neoproterozoic suture strikes north (Ries et al., 1992). The central Kenyan rift is part of the Aswa transform, reactivating a NW-striking shear and thrust zone (Smith and Mosley, 1993), which was previously described as the Yatta line, an ancient continental margin (Chorowicz et al., 1988). The Tanganyika–Rukwa–Malawi segment has suffered multiple reactivations, summarized by Delvaux (2001) as follows: (1) construction of a NW-striking Paleoproterozoic Ubende Shear Belt (1950–1850 Ma), along the western margin of the Archaean Tanzania craton (Chorowicz et al., 1990; Chorowicz, 1992; Theunissen et al., 1996; Klerkx et al., 1998), which had likely reactivated an earlier Paleoproterozoic suture (2100–2025 Ma); (2) during the Meso- and Neoproterozoic, formation of sedimentary basins; (3) during the Pan-African event, strike-slip movements; (4) during the late Paleozoic-early Jurassic, formation of Karroo rifts (Daly et al., 1989); (5) during the Eocene-early Miocene interval, an inferred rifting stage (Damblon et al., 1998). The Malawi rift is forming in the frame of a N-striking Mozambiquian belt, and more to the south, the Mweru area is superimposed to a Middle Proterozoic (Irumides) belt.

9.3. Mechanism of formation

The mechanism of formation of the EARS is still a subject of discussion. Since Suess (1891) and Gregory (1921), it is generally admitted that rift evolution is related to extension. However, rifting may be triggered by collision, as suggested for the formation of the West European rift system (Chorowicz and Deffontaines, 1993), but in Africa the Karroo compression in the south ceased during the Jurassic, and in the north the Mediterranean has suffered a complex evolution during the late Cenozoic, comprising subductions, E–W strike-slip movements, and collisions that are too much local and started too far in the past (Eocene) to be taken into account. It is unavoidable (McConnell, 1977; Richardson, 1992; Zoback, 1992; Delvaux, 2001) to examine the influence of far-field compressional stress due to oceanic ridge pushes in the east (Indian Ocean) and the west (Atlantic Ocean). This hypothesis is difficult to defend because the Central Indian oceanic ridge formed at 38 Ma (Patriat et al., 1982), and the Atlantic ridge was initiated even earlier. At more local scale, Bhattacharji and Koide (1987) have suggested from theoretical and experimental studies, the development of compressive stress adjacent to and around the active rift zones, due to mantle upwelling and penetrative magmatism. Finally in terms of plate tectonics, block movements in East Africa are divergent, and tension might be considered the major factor (McClusky et al., 2003).

But plates are rigid and do not easily suffer in their interior pressure release (Zoback, 1992). Rheology of the lithosphere has been estimated by Fadaie and Ranalli (1990) on the basis of geothermal data. The uppermost brittle layer varies from 10 ± 2 km in the eastern rift, to 18 ± 5 km in the western rift and 26 km in the shields. Deformation of the continental lithosphere, leading to rupture, has been theoretically conceived to occur by two possible ways (Burke and Dewey, 1973; McKenzie, 1978), considering the role played by the asthenosphere. (1) *Active rupture* would result from mantle convection (Pavoni, 1993) and plume movements in a dynamic asthenosphere that forcibly intrudes and deforms the overlying lithosphere. (2) *Passive rupture* model sees the asthenosphere uplift playing an entirely responsive role in filling the gap produced by lithospheric extension, itself a reaction to stresses generated elsewhere at plate boundaries due to external forces.

Models of tectonic rupture range from whole lithosphere simple shear (e.g., Wernicke and Burchfield, 1982; Lister et al., 1986) to combination of upper brittle layer simple shear, lower crustal delamination and lithospheric mantle pure shear (e.g., Lister et al., 1991). Gibbs (1984, 1990) has pointed out that stretching of the crustal brittle zone implies low-angle shallow detachment faults. According to Hendrie et al. (1994), the upper and middle crust experience simple shear by planar faults, while the lower crust and upper mantle are extended by pure shear and plastic deformation. These models consider an initial rather homogeneous lithosphere, but preexisting fabric is evidently frequent in the crust and also likely in the upper mantle. Mechanism of reactivation of former structures has to be considered differently in the crust and in the lithospheric mantle.

In the crust, ancient belts comprise fault systems of frontal ramps verging in opposite directions, and lateral ramp or tear faults which form several families by strikes, some being almost by places parallel to Cenozoic divergence (Fig. 18). Reactivation of a lateral ramp or tear fault forms a transfer fault. Reactivation of the deep, ductile part of a frontal ramp results in depth in low-angle detachment fault linking up with major border normal fault. Across a former lateral ramp, the detachment reactivation can change side, and the border fault then frequently changes vergence, with reversal in rift asymmetry. An argument favoring this view is the transfer fault system in EARS segments such as the Malawi rift, separating 70 km long sub-basins with half-graben structure, the asymmetry changing polarity across the transfer faults (Chorowicz and Sorlien, 1992). It is unlikely that the ancient frontal ramp shear vergence changed every 70 km along its strike in an ancient belt. This leads to the concept that tensional detachment faults may reactivate opposing reverse frontal ramp shear faults belonging to a former double verging belt, whose axis is in the middle. The low-angle tensional detachment faults may indifferently rework frontal ramp shears of either sides of the belt. Consequently, the rift segment is superimposed right over a former belt, likely linked to a suture or a transform zone whose root lies deeper than crust.

In the lithospheric mantle, along ancient sutures, transforms or even rift zones, there is a strong tectonic fabric due to preservation of a lattice preferred orientation of olivine crystals, inducing a large scale anisotropy, which may result in directional softening, leading to heterogeneous deformation (Tommasini and Vauchez, 2001). This mechanical anisotropy may induce strain localization, mainly producing tension fractures, and propagation of this initial instability, which will follow the old structural trend. But such lithospheric scale weaknesses are quite numerous in the African lithosphere, due to its complex history during the Precambrian. Only some of them are reactivated, and this shows that another factor of weakening should intervene, at least for the initiation of the continental rupture.

It is clear that the initial lithospheric failure appeared in the north, in relationships with plume onset. But the plume combined its effects with the presence of a Pan-African suture zone. Moreover, the core of the plume was not in the middle of the future Afar, but more to the west under Lake Tana (Chorowicz et al., 1998). The panache of the plume, arriving at ~30 Ma in the horn of Africa, was almost 1000 km in diameter, heating, thinning and consequently weakening a large area, in which the preexisting Pan-African suture zone was predisposed (Jepsen and Athearn, 1961). According to Mohr (1971), the dikes feeding the flood basalts of the trap series would have been injected along vertical fissures during the Palaeogene, faulting in the rift occurring later. This was also evidenced by west to east migration of Oligocene to Miocene volcanism (32–16 Ma) from the Ethiopian Plateau to the Ethiopian rift (Zanettin et al., 1974). This conjunction of two weakening effects (ancient suture + plume), under tensional environment, is likely to have been responsible for first formation of tension fractures and tensional faults in the Afar, long after plume emplacement (Wolfenden et al., 2004). When this happened, the Red Sea–Gulf of Aden opening was already active. This conclusion is in agreement with Bosworth et al. in another section of this volume. The result is the Afar triple junction, which looks very similar to a theoretical triple junction of rifts forming over a plume core (Burke and Whiteman, 1973).

The plume moved southwards due to possible drifting of Africa (Bonavia et al., 1995; Chorowicz et al., 1998). Another view in that Africa has been essentially stationary during the past 30 Ma (Burke, 1996), and in this case it would be necessary to hypothesize the emplacement of more than one plume head. According to George et al. (1998), the rate of migration of the onset of volcanism from southern Ethiopia to Tanzania is similar to the rate of migration of the African plate over the same period. In the Kenyan dome, arrival of a mantle plume prior to 20 Ma was followed by crustal up-warping at 15 Ma, and flood volcanism from partial melting at the base of the lithosphere (Smith and Mosley, 1993). The influence is not that of the centre of the plume, but lateral effect of the panache, because the uplift was moderate (1 km) and limited volume of lava was produced. In the Kenyan rift, thermal thinning of the lithosphere has to be invoked in order to explain the discrepancy due to more thinning of the mantle lithosphere than in the crust (Morley, 1994).

After weakening by plume (Zeyen et al., 1997), a failure initiated in the Afar and propagated southwards along the main Ethiopian rift, following the suture zone. The failure that nucleated more to the south in the Kenyan dome, at another suture location, propagated away from this zone (Smith, 1994). The main Ethiopian and northern Kenyan rifts propagating respectively southward and northward were following different Pan-African suture zones (Chorowicz, 1992), could not link, and instead spread out laterally forming a relay pattern of splayed graben basins in the Omo-Turkana region (Chorowicz, 1992; Ebinger et al., 2000). This is also where the Mesozoic Anza graben crosses beneath the EARS and links with the Sudan rifts (Bosworth, 1992). To the south, the failure encountered thicker, colder and more competent lithosphere, splayed in the north Tanzanian divergence, and stopped. Along the western branch, after arrival of the plume and related panache, failure nucleated likely in Kivu-northern Tanganyika area, and rifts propagated in opposite directions, the link with the eastern branch being accommodated by the Aswa transform zone. New nucleations are now occurring in the south (Lake Kariba, Limpopo valley). It is clear that large segments of the rift system developed far from the locations of plume impacts, as is the case for all the southern segment of the western branch. The rift system can propagate outside the area influenced by plume. On the whole (Fig. 10), the plume and panache trajectory in the west, or successive plume emplacements, have produced lithosphere thinning, initiated breakings and related asthenospheric intrusions. These breaking lines have propagated further following ancient weak lines, sometimes outside the zone of thinned lithosphere.

In the graben basin segments, the structural evolution is under effects of the divergence of continental blocks (1 in Fig. 18). Strain may be focused on a preexisting belt, and develops differently in the lithospheric mantle and in the crust. In the lithospheric mantle, mantle pure or simple shear is centred on a preexisting fabric of two different lithospheric plates, which had been formerly assembled after ancient subduction-collision. This fabric is the memory of the lithospheric mantle. After occasional wide thinning due to plume panache, focused thinning of the lithospheric mantle allows the rise of an asthenospheric elongate diapir (2 in Fig. 18). Heating of the lithosphere and rise of isotherms is responsible for wide (hundreds of kilometers) uplift of the area (3 in Fig. 18). In the crust, for a given graben basin, a unique low-angle detachment fault, abruptly connected to lisric major border fault, accommodates horizontal extensional movement (4 in Fig. 18). This deformation induces a roll-over structure that produces local subsidence (5 in Fig. 18). In the whole, focused (tens of kilometers wide) subsidence in the crust is superimposed to wide rise of the lithosphere. Extension is also taken by normal faulting, block tilting and production of numerous tension fractures, some of them reaching melt conditions in depth and permitting ascension of the magma. With time, the rift becomes more magmatically active. Dyke intrusions in depth are a process of crust extending further, and this process seems to become with time more important for the rate of extension than normal faulting. The final stage of this process of dyke intrusions, accompanied by high angle normal faulting, is a symmetrical oceanic rift with asymmetric margins inherited from previous history (Bosworth, 1994).

9.4. Tectonics and sedimentation

The sedimentation patterns in the EAR system are controlled by structures, with strong influence of climatic environments and occurrence of great lakes (Crossley, 1984; Le Fournier et al., 1985; Tiercelin et al., 1992a,b; Frostick and Reid, 1990; Lambiase and Bosworth, 1995). In the vicinity of the graben basins, preexisting structures and recent faults focus drainage and input place of sediments, while regional uplift of rift shoulders induces high energy erosion, but part of the drainage is deflected away from rift.

Detritic sediments accumulate at piedmont of major border faults (Fig. 20), forming alluvial cones, alluvial and delta systems, down slope fan deltas and deep fan deltas. There is sometimes development of littoral platforms with fan deltas, prograding deltas, underflows from canyons, carbonate deposits, stromatolites. The platforms and monocline tilted blocks are dissected by gorges. Landslides, rock fall deposits and deep fans are located at foot of the major boundary faults. In axial deep basins there are "sheet drape" sequences of homogeneous or laminated organic-rich mud, in anoxic conditions, and distal turbiditic sedimentation. The predominance of rift parallel fault blocks facilitates axial sediment transport, at the expense of down-dip, lateral transport. The transfer and transform faults separate several depocentres, which asymmetry is related with the roll-over patterns. Organic sediments are preserved in deep and sometimes shallow basins. Evaporitic sediments are under semi-arid climate, in shallow basins. Metallogenic deposits are related to hydrothermalism along major border faults (Tiercelin et al., 1992a,b; Thouin and Chorowicz, 1993).

The earliest stages of rift evolution (pre-rift and initial stages), are characterized by slightly subsident, marshy small graben basins. Sedimentation is a combination of

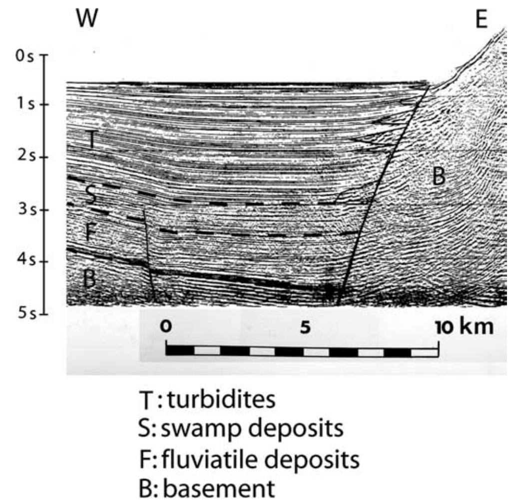

Fig. 20. Typical sedimentation sequence in rift lake as seen from seismic profile (project PROBE, extract of E–W seismic profile across Lake Tanganyika, at latitude 05 °S). B: basement; F: sand-dominated fluviatile deposits; S: clay and silt deposits with high organic content, corresponding to swamp environment; T: turbidites.

organic and detrital deposits in shallow lakes. The first sediments resting on the basement are sand-dominated fluviatile deposits (F in Fig. 20). They are followed by clay and silt deposits with high organic content corresponding to swamp environment (S in Fig. 20). In late stage of rifting, sedimentation changes and is characterized by thick deposits of deep lacustrine detrital sequences forming turbidites (T in Fig. 20).

9.5. Cradle of mankind

The concept of the East African rift being the 'cradle of Mankind' was expressed by Leakey (1973). This implies close relationships between the evolution of the EARS and the occurrence of the Homo species. Scenarios are many and all hypothetical. According to the structural data, the formation and evolution of the cradle is controlled by tectonics at lithospheric scale. The uplift of plume domes, subsidence of graben basins, asthenospheric diapirs and subsequent uplift of shoulders, together with the construction of volcanoes was slow. These processes progressively modified the environment of the Hominoids, creating barriers and downthrown basins, likely frequently isolated from outside main Africa. Induced climatic changes at regional and local scale are responsible for the onset of a savannah environment on isolated rift floors, succeeding to forest. These savannah depressions, surrounded by high relief, had the particularity to include freshwater lakes.

Denys et al. (1985, 1986) have found close correlations between the Mio-Pliocene evolution of rodents (massive extinctions and subsequent radiative evolutions) and the tectonic history of the eastern branch of the East African rift. Isolation of savannah basins seems to have played a major role in forcing rapid evolution of the endemic rodent populations. This suggests that the same structural and climatic history has favored the birth of savannah-adapted hominoids and Homo species. Since 2 Ma, disappearing of barriers in the eastern branch is almost coeval with mobility and homogenization of the rodent populations—and migration of Homo species.

10. Conclusions

The East African rift system can be regarded as a unique succession of graben basins linked by intracontinental transforms and segmented by transfer zones and accommodation zones. This is a lithospheric scale structure, characterized by uplift of hot asthenospheric elongate diapirs across the upper mantle, responsible for shoulder uplifts forming an intracontinental ridge system, several hundreds of kilometers wide, several thousands of kilometers long, quite equivalent to an oceanic ridge system. It clearly appears that the inscribed continental graben valleys and basins are organized over a major failure, related in the crust to a main border fault and low angle detachment fault, inducing asymmetric roll-over structure, eventually accompanied by smaller normal faulting and tilted blocks. External tensional stresses have caused the continent to split along lines of preexisting lithospheric weaknesses marked by old tectonic patterns that focus the extensional strain.

By its size, structure and occurrence of oceanic lithosphere in the Afar, the EARS can be taken as a model of the beginning of oceanic opening inside a continent. The opening propagates on the whole from north to south—including local nucleations of lithospheric failures—due to SE-ward relative divergent drifting of a not yet well individualized Somalian plate.

The role of plume impacts is determinant in EARS history. The plume that had occurred at around 30 Ma was in Lake Tana region (Ethiopia), its almost 1000 km diameter panache weakening the lithosphere and preparing the first rifting along a Pan-African suture zone bordering the future Afar region. The rift propagated afterward along the suture zone, activation of low angle detachment fault reworking former opposing faults belonging to the double verging ancient belt associated to suture. Supposing the plume migrated southward, or other plumes emplaced, preparing other nucleations of failure along former suture or transform zones, the rift system propagated and is still developing southwards. The main phenomena are formation of domes related to plume effect, weakening the lithosphere and, long after, upper mantle failure inducing asthenospheric intrusion and related thermal uplift of shoulders. The rift can propagate outside the area influenced by a plume. This new view of rift formation reconciles the classical models (Ziegler and Cloetingh, 2004): active plume effect triggered the first ruptures; passive propagations of failure along lithospheric scale weak zones were responsible for the onset of the main rift segments.

A lesson from this overall consideration of the EARS is that basins are of different types. (1) Most of them are rift graben basins, with asymmetric roll-over structure and thermally uplifted shoulder expressing asthenospheric intrusion. Generally these basins are more or less filled with volcanic or sedimentary deposits, but if not deep and drained by river they may form only a rift valley devoid of sediments, as is the case for the southernmost basin of Lake Malawi rift. (2) Others are found along NW-striking transform zones, as is the case typically for southern Tanganyika and Rukwa basins, and then they trend almost parallel to the differential movement of lithospheric blocks (oblique extension), and comprise major strike-slip faults. In this case shoulder uplift is not systematic, and when it occurs it corresponds to folds of the basement accompanying the transcurrent deformation. (3) The half-graben basins that lie in the Omo-Turkana relay zone between the Ethiopian and Kenyan rifts are of smaller dimensions, with no uplifted shoulders. Tensional strain spreads over a large area and thinning of the lithosphere is insignificant. The lack of uplifted shoulders indicates that the basins are not associated to ascending asthenospheric bodies. (4) Some graben basins are not situated along the main breaking rift line, and can be considered as appended basins. This is the case for instance for the North Horr basin in the Turkana region, Lake Mweru south of Lake Tanganyika and Usungu lows north of Lake Malawi (Nyasa). They do not have uplifted shoulders. They have to be considered as essentially crustal structure, with no significant upper mantle particularities.

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This synthesis is compiled from abundant literature but no attempt is made to present a comprehensive bibliography. I tried to favor the first mentions of new observations and understandings.

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