Presented at SDG Short Course IV on Exploration and Development of Geothermal Resources,
organized by UNU-GTP and KenGen, at Lake Bogoria and Lake Naivasha, Kenya, Nov. 13 – Dec. 3, 2019.

UNITED NATIONS UNIVERSITY
UNU-GTP
Geothermal Training Programme

KenGen

THE INFLUENCE OF PRE-EXISTING CRUST IN THE EVOLUTION OF EARS AND ITS IMPACT ON THE GEOTHERMAL EXPLORATION STRATEGY OF THE WESTERN RIFT

Ermias Yohannes
Ministry of Energy and Mines, Department of Mines
P.O. Box 272, Asmara
ERITREA
ermias_yohannes@yahoo.com

ABSTRACT

The East African Rift System (EARS) is a well developed continental rifting that splitted the Somalian Plate from Africa (Nubian). Volcanism initiated in the north in Mid Oligocene, in Kenya at 25 Ma followed by episodic rifting. The western rift is young, in which volcanism began at about 12 Ma in the north and at about 7 Ma in the south. Volcanism in the south is only confined to four volcanic provinces, the Tore-Ankole, Virunga, South Kivu and Rungwe.

The lithospheric and crustal structure is addressed through P and S wave anisotropy. The East African Rift and related plume extending from Malawi to the Red Sea illustrate the geodynamics of the mantle below Africa and the underlying mantle convection. Studies of lithospheric thickness along the northern rift, Main Ethiopian Rift indicate that, under the very northern most sections of the rift near Afar, where the extension factors are highest, and rifting has progressed furthest, the lithosphere has been thinned appreciably from its pre-rift, steady-state thickness. It becomes thicker towards Tanzania and around Rukwa and Lake Tanganyika with crustal thickness of 40-44 km.

Pre-existing structures have controlled the location and rifting in the East African Rift System. The Paleoproterozoic Ubendian orogeny occurred between 2100 and 1800 Ma is prominent in the south is a high grade metamorphism and interpreted as a product of collisional orogeny. Whereas the MesoProterozoic Kibaran belt is a short lived but prominent at about 1375 Ma tectonomagmatic event. The Western Branch of EARS follows preferentially these two belts and avoids Archean. The EARS easily penetrated through the Juvenile NeoProtorozoic East African Orogeny.

The less evolved western rift is characterized by thick crust, less volcanism and thick sedimentation makes it to define a different approach than the Eastern Rift. The region from Ruwenzori through Kivu to Rusizi and finally to south should be handled distinctly separable and rifting related to transfer zone should be considered to outline strategy for geothermal exploration.

1. INTRODUCTION

The East African Rift System (EARS) is a major geodynamic feature which has been splitting the African continent since the late Eocene-Oligocene (e.g. Macgregor, 2015). The EARS (Figure 1) consists of a series of connected continental rifts separating the main African (Nubian) Plate from the Somali Plate, and the Indian Ocean. The rift valleys form two main lines, the eastern and western branches of the EARS. A third, southeastern branch is in the Mozambique Channel (Chorowicz, 2005).

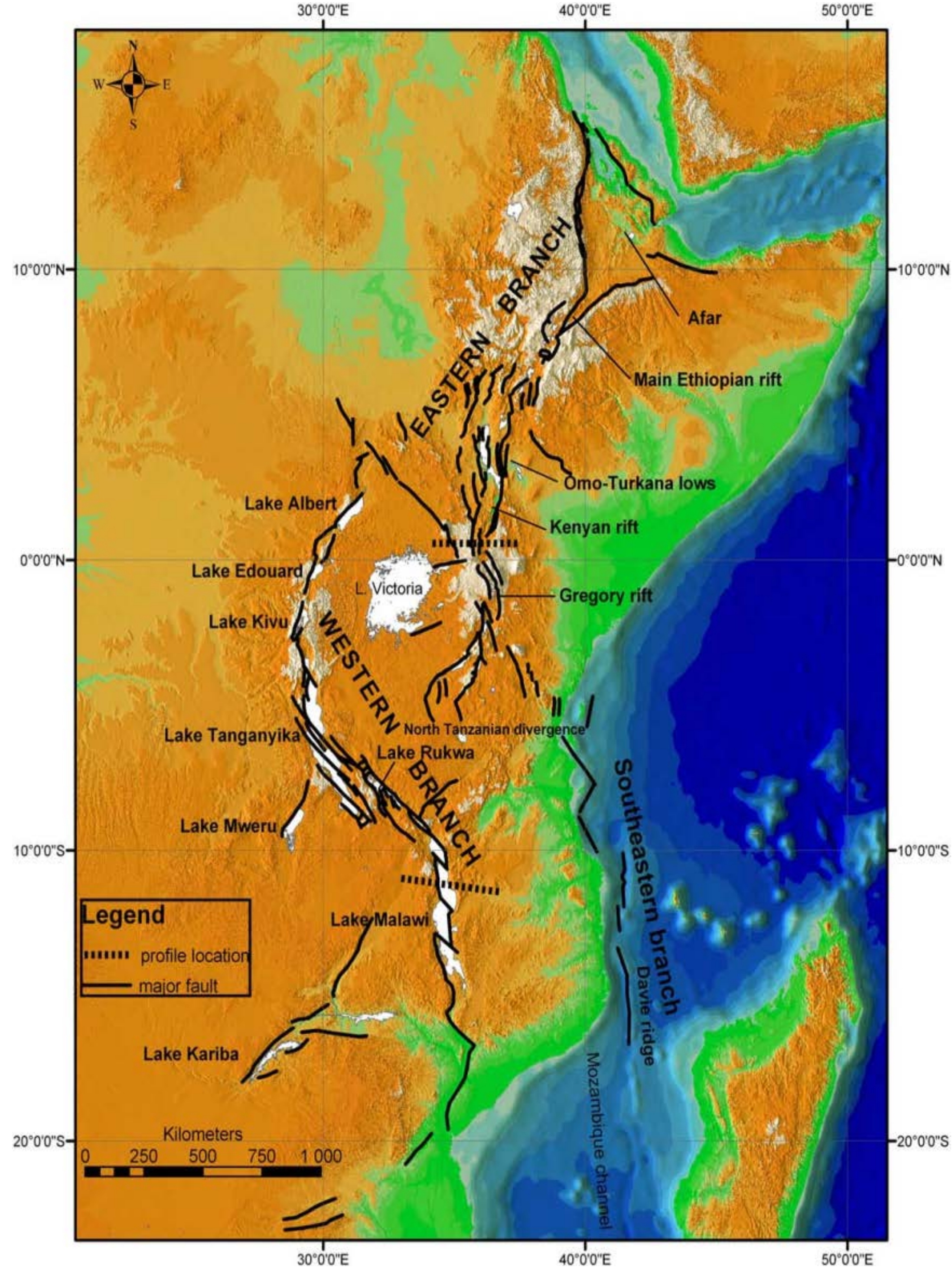


FIGURE 1: Location of the East African Rift System. The EARS is a series of several thousand kilometers long aligned successions of adjacent individual tectonic basins, separated from each other by relative shoals and generally bordered by uplifted shoulders.

The EARS propagates from the northern Afar Depression and Main Ethiopian Rift to the southern diverging branches (Figure 1): The Eastern branch from the Kenyan rifts to central Tanzania and, the western branch runs through Lake Albert, Lake Edward, Lake Tanganyika, and Lake Malawi, and ends in southern Mozambique.

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. 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.

2. EVOLUTION OF EARS

Rift-related volcanism, faulting and topographic relief are well-exposed in the EARS, this makes it to be a classic continental rift. Variety of tectonic styles, reflected in grabens and half-grabens, transverse fault zones and accommodation zones, and influences of pre-existing geologic structures, are present in the rift system.

The evolution of the rift system can effectively be considered in three parts: Afar and the Main Ethiopian Rift in the north, the Eastern Branch and the Western Branch.

2.1 Northern Rift

The Northern Rift comprises the Afar and the Main Ethiopian Rift. It is characterized by bi-modal magmatism, transitional basalts that straddle the tholeiitic-alkaline boundary and younger (Upper Miocene-Pliocene to Recent) trachyte-pantellerite ignimbrites and lavas (Figure 2). The volume of volcanism of the Northern Rift collectively reaches about 300,000 km^3.

The Afar part extreme north - Beginning about 4 Ma, extensive basaltic volcanism covered most of Afar and obscured the geologic record of events (Braile et al., 2006). Since that time, the history of northern and eastern Afar is more a story of sea-floor spreading than continental rifting.

In the Main Ethiopian Rift and its flanking plateaus, volcanism began in the mid-Oligocene or earlier. WoldeGabriel et al. (1990) found evidence for six episodes of volcanism but no evidence for lateral migration of volcanism from the flanks to the rift valley such as is observed in Kenya.

2.2 Eastern (Kenya) Rift

A general migration of volcanism from north to south and from west to east and progressions from strongly alkaline to less alkaline magmatism and from mafic to more evolved and felsic compositions are noteworthy aspects of Kenya rift magmatic evolution. The volume of volcanism is

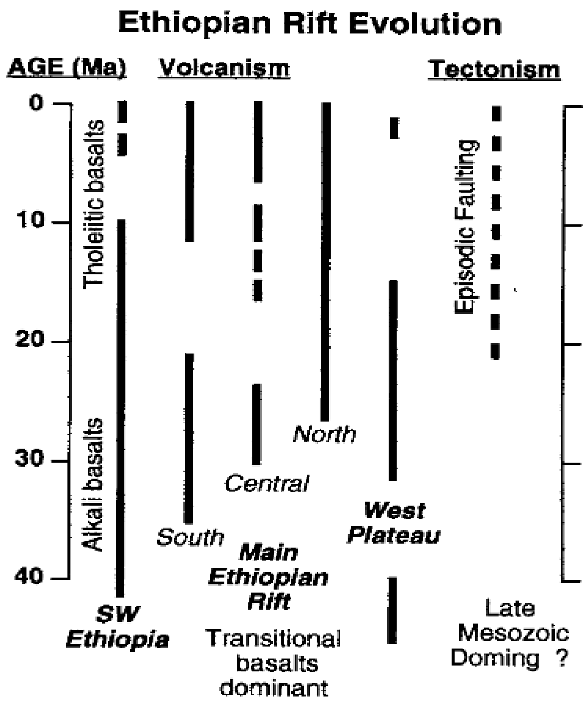


FIGURE 2: Schematic diagram illustrating the volcanic and tectonic evolution of the Ethiopian Rift during the last 40 million years

230,000 km^3. The main phase of uplift took place during the period of 40-21 Ma (Figure 3). A pre-rift depression formed in the early Miocene, but clear development of half grabens did not occur until about 12 Ma (Smith, 1994).

2.3 Western Rift

The Western Rift generally appears to be younger than the Kenya rift. Doming began at about 20 Ma (Figure 4) and volcanism began at about 12 Ma in the north and at about 7 Ma in the south (Ebinger, 1989).

The volcanism has continued to the present. Volcanism in the Western rift occurs in four isolated Miocene to recent volcanic provinces including, from north to south, the Toro-Ankole, Virunga, S. Kivu, and Rungwe regions. The development of the many basins in the Western rift began concurrently with or prior to the volcanism. Through time, the border faults propagated to both the north and the south resulting in the linkage of many basins which were initially separate features. The basins have tended to narrow with time due to hanging wall collapse (Ebinger, 1989). As in the area of the Kenya topographic dome, the main phase of tectonism occurred in the last 5 Ma.

Apart from the graben of the western rift the transverse structure and faults play an important role in shaping the rift. The main troughs are linked or cut by transverse structures, generally striking NW-SE. There are two types of NW-SE transverse structures (Chorowicz et al., 1987): (1) large lineaments linking the main segments of the Rift; and (2) fault zones dividing the main rift segments into elementary parallelogram-shaped basins.

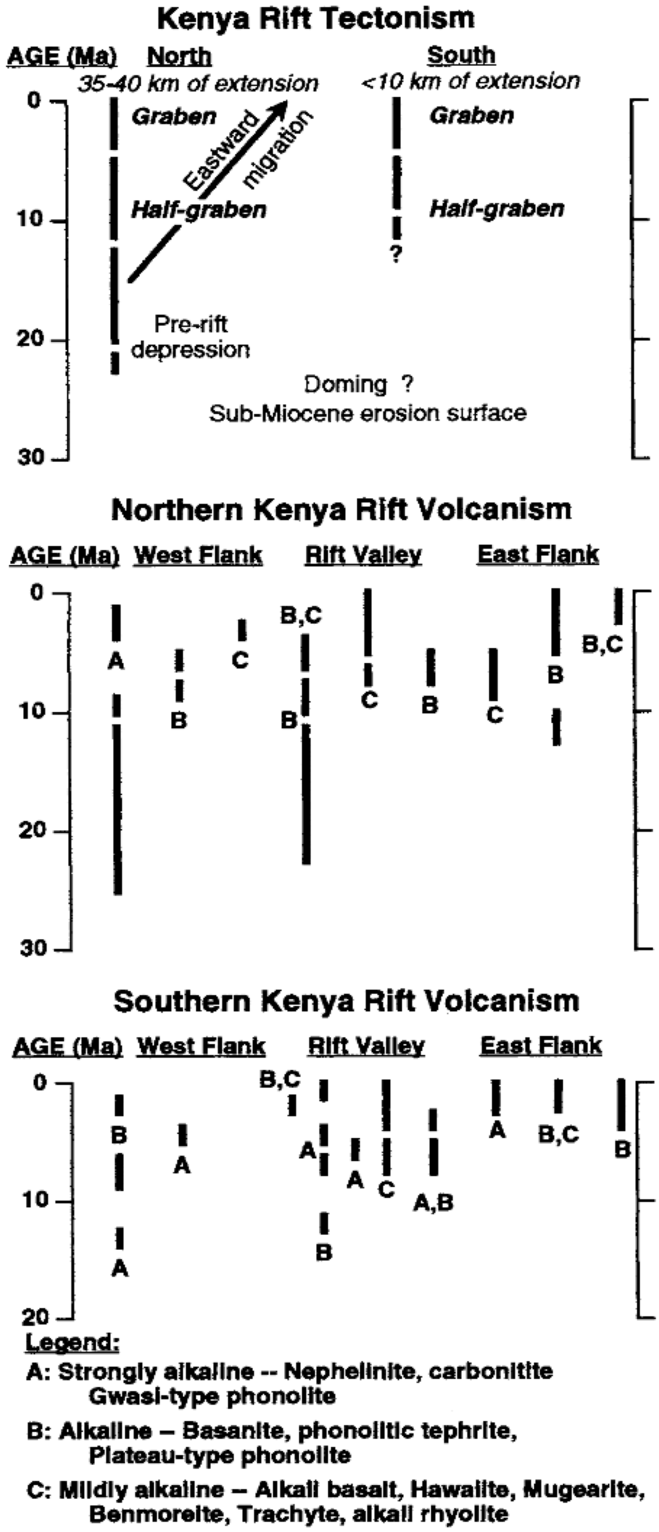


FIGURE 3: Tectonic evolution of the Kenya Rift during the last 30 million years

The major fault zones bordering the main grabens, linked by the transverse faults inside these grabens, along with the Tanganyika-Rukwa-Malawi lineament and, to a lesser extent, with the Aswa and Zambezi lineaments, form the principal directions of fracturing in East Africa.

3. PROPERTIES OF THE LITHOSPHERE BENEATH EARS

Based on the seismic wave tomography and gravity anomaly studies, the structure and evolution of the lithosphere and upper mantle provide the continental wide structural background for the seismotectonic map (Meghraoui, 2015). The lithospheric and crustal structure is addressed through the P and S waves anisotropy tomography and the results of receiver functions.

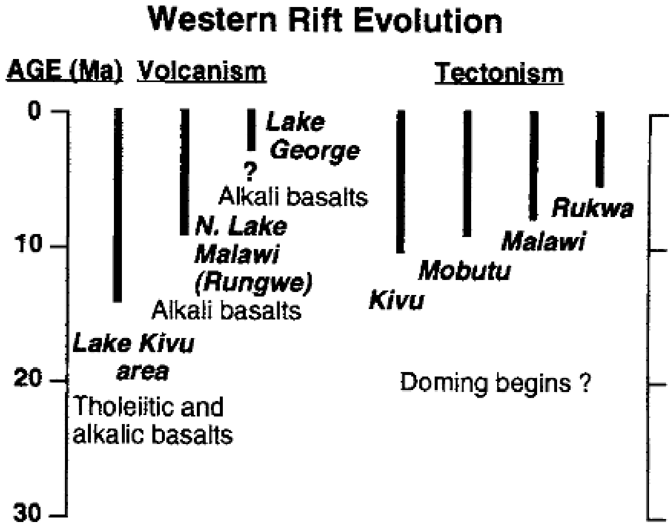


FIGURE 4: Schematic diagram illustrating the volcanic and tectonic evolution of the Westem Rift during the last 30 million years

The East African Rift and related plume extending from Malawi to the Red Sea illustrate the geodynamics of the mantle below Africa and the underlying mantle convection (Figure 5, a and b). In comparison with the cratons, the plume corresponds to the signature of hot materials and testifies for the volcanic activity with continental deformation in agreement with the seismicity distribution visible in the map.

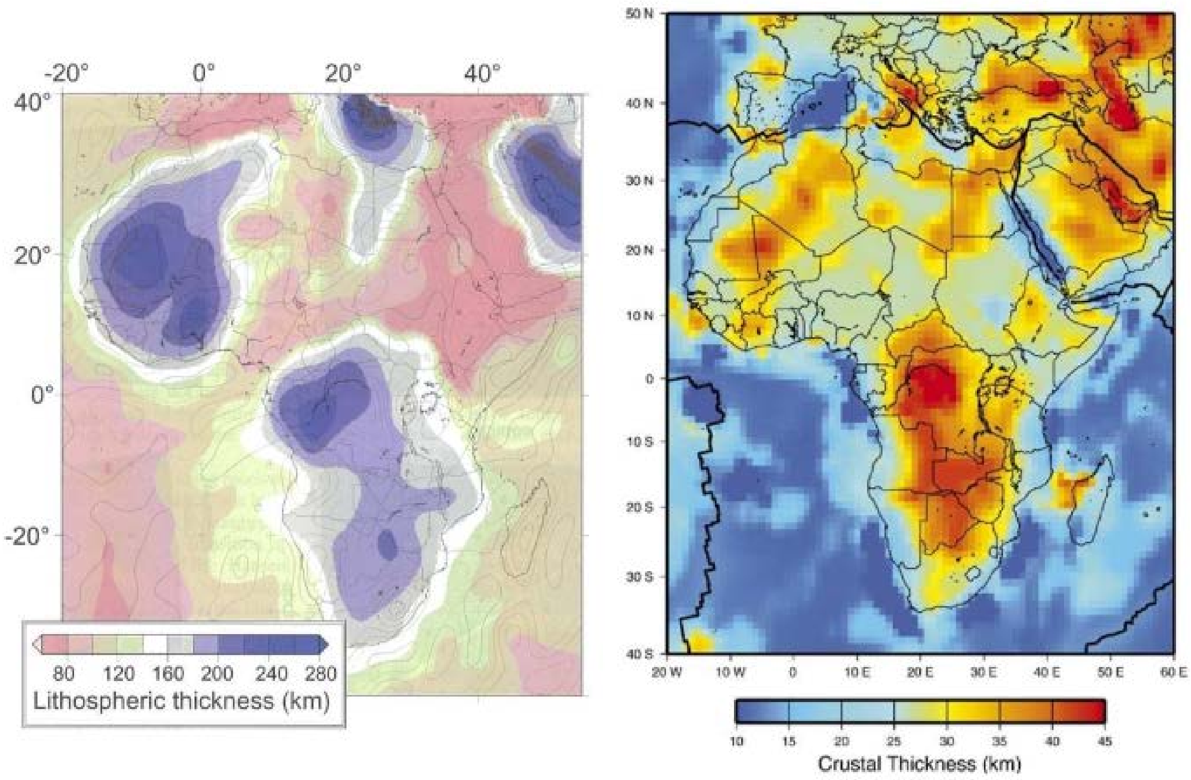
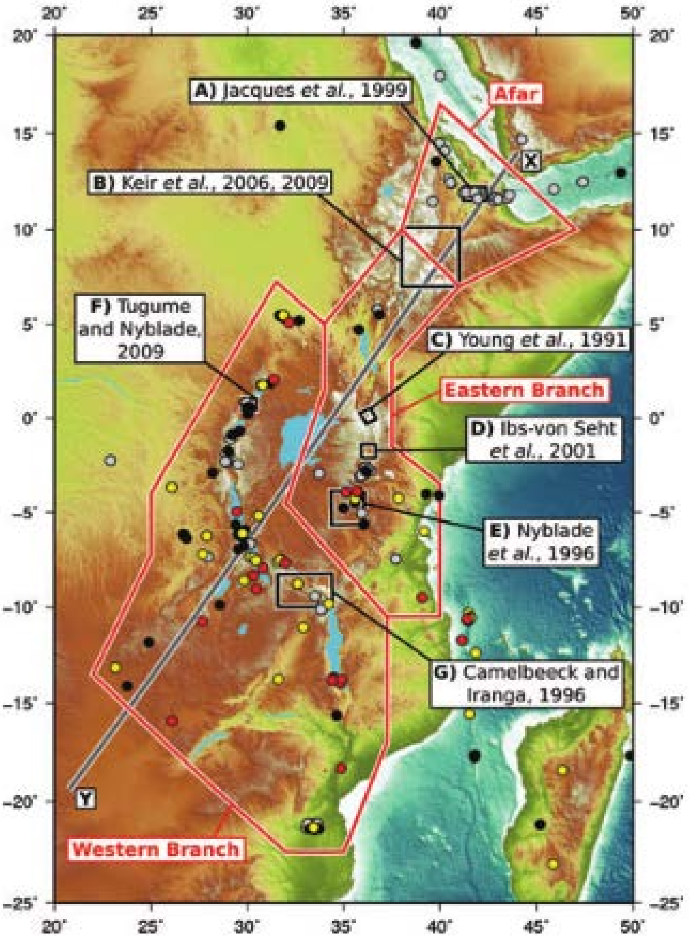


FIGURE 5: a) Estimated lithospheric thickness from tomography and gravity anomalies (left) and b) Crustal thickness from tomographic data (Magraui et al., 2016) (right)

An attempt has been made to show earthquake depths with the lithospheric thickness of East African Rift (Craig et al., 2011, and the reference therein). Along the northern sections of the East Africa, the lithosphere is too thin (<110 km). We expect the off-rift lithosphere to be of a thickness consistent with the maximum stable thickness of the lithosphere in the oldest oceans, limited by the development of convective instabilities to ~100 km (Craig et al., 2011, and the reference therein).

Studies of lithospheric thickness along the Northern Rift, Main Ethiopian Rift indicate that, under the very northern most sections of the rift near Afar, where the extension factors are highest, and rifting has progressed furthest, the lithosphere has been thinned appreciably from its pre-rift, steady-state thickness. All regions where such thinning has taken place demonstrate seismogenic thicknesses of 15–20 km or less.

The contrast between upper-crustal seismicity in regions of thinner lithosphere and whole-crustal seismicity in regions with thicker lithosphere is seen particularly clearly in northern Tanzania (Craig et al., 2011). In northern Tanzania, the eastern branch of the EARS splits into three separate segments (Figure 1). The Pangani branch goes southeast and eventually links up with the marginal



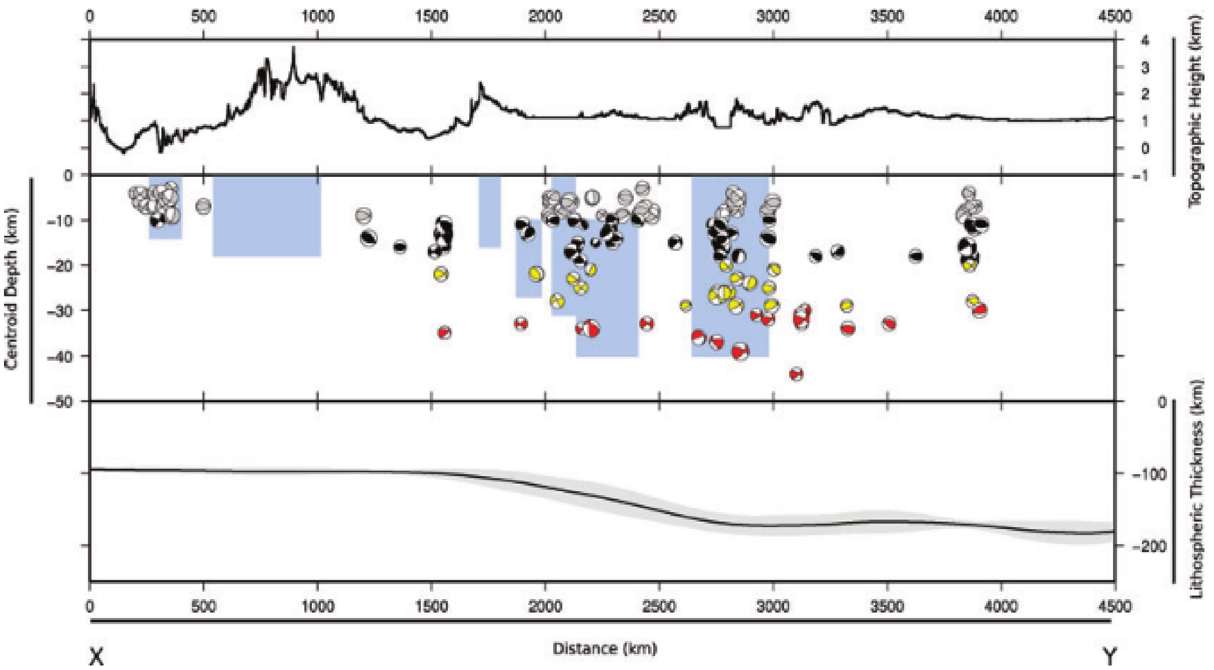


FIGURE 6: a) (Top) The areas of local seismic networks, discussed in the text, are shown by labelled black boxes. Red boxes show the areas of depth/frequency. Grey line shows the line of projection for Figure 6b; b) (Bottom) Earthquake modelling results are shown by coloured circles. Shading is indicative of centroid depth.

extensional basins along the Tanzanian coast. The westernmost segment enters the Tanzanian craton along Lake Eyasi, and rapidly dies out. The central Manyara–Balangida branch passes around the edge of the Tanzanian craton, at least on the surface, passing along a system of aseismic extensional depressions, before linking up with the western branch at the Rungwe triple junction, between Rukwa and Lake Malawi.

Typical values place the Moho at 40–44 km beneath the western branch and 37–42 km beneath the eastern branch (Craig et al., 2011). Crustal thicknesses in Afar of 13–28 km (Figure 6) are consistent with higher extension factors and thinner crust at the northern end of the Main Ethiopian Rift (Craig et al., 2011, and the reference therein). Crustal thickness along the rift is presumed to increase with decreasing extension factor southwards along the Main Ethiopian Rift through Turkana up to the 40 km values seen in places around southern Kenya and Tanzania. Both receiver function analysis and *PmP*-phase analysis find crustal thicknesses of 40–44 km around Rukwa and Lake Tanganyika.

4. PRE-RIFT CRUSTAL DEVELOPMENT OF THE EARS

Pre-existing structures have controlled the location and the initial stages of rifting in the north of the East African Rift System (Figure 7) specifically the Northern (Main Ethiopian) Rift (Keranen and Klemperer, 2008). Two distinct Proterozoic basement terranes, one strongly underplated beneath its northern part, created rheological boundaries that localized extension and permitted the Main Ethiopian Rift (MER) to propagate. Southward propagation in the northern MER along the eastern limit of mafic underplating is mainly controlled by the Proterozoic terrane boundary stalled for 5–6 M.y. at what would become the NMER–CMER boundary, south of which the strong rheological heterogeneity provided by the underplated block was no longer available.

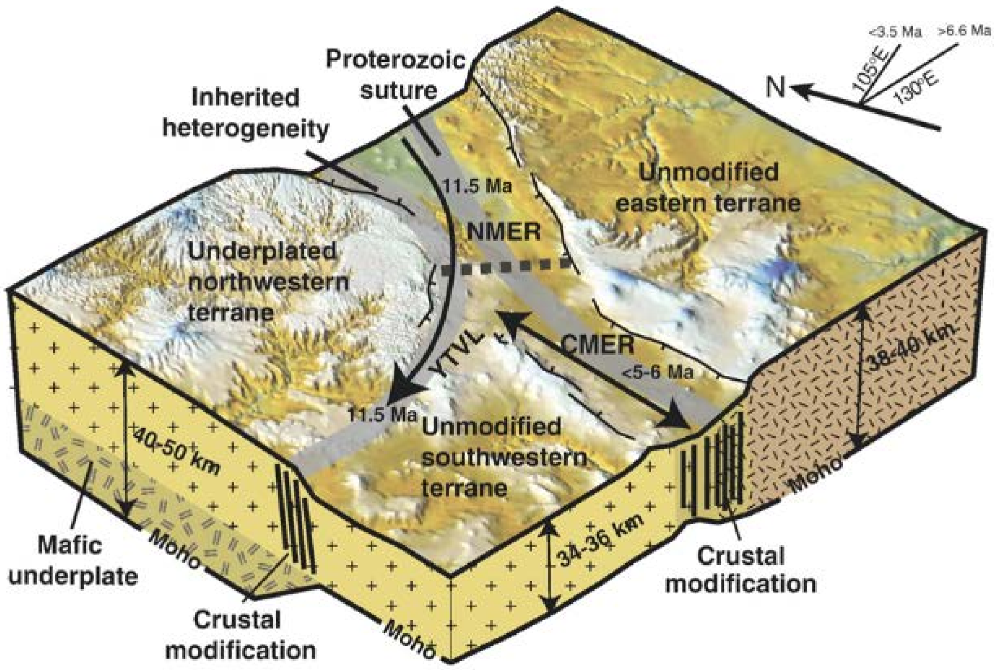


FIGURE 7: Lithospheric terranes, crustal thickness, and locations of Cenozoic extension and volcanism. Cenozoic extension follows the rheological boundaries along the edges of the underplated northwestern block and the boundary between the eastern and western crustal terranes (Keranen and Klemperer, 2008).

Experimental modeling and field work have shown that segments of the East African Rift System is controlled by pre-existing fabrics and it inherited its trend from the older zones of weakness between the cratons, and belts (Aanyu and Koehn, 2011). Therefore understanding the crust development before attaining the present configeration of Africa as a continent is necessary to know the evolution of EARS.

4.1 Orogenies of the East African Region

The Congo and Tanzanian Cratons are the relatively stable early rocks of Archean basement composed mainly of granitised formations suffered a protracted period of tectonic activity superposed on old cratonic blocks and spanning the period c. 3000–2500 Ma (Figure 8). The rocks of the Archean gneissic granulitic complex constitute the oldest rock formations that stratigraphically underlie younger cover formations.

The Paleoproterozoic Ubendian orogeny occurred between 2100 and 1800 Ma, producing high-grade metamorphic lithologies through two deformational phases. The early (more regional) deformational phase (2100–2025 Ma) marked by an E–W to WNW–ESE foliation and granulite-facies metamorphism, is interpreted as a product of collisional orogeny along the southwestern margin of the Tanzania, and possibly Congo cratons (Figure 8). The second phase (1950–1850 Ma; Aanyu and Koehn, 2011 and the references therein) is characterised by large NW–SE trending dextral shear zones and was restricted to the Ubendian belt. This resulted from a N-ward compressional stress regime that also caused N-ward directed thrusting of rocks of the Usagaran belt immediately south of Tanzania Craton and intrusion of late- to postkinematic calc-alkaline granitic batholiths dated at c. 1860 Ma. Hence, the upper time limit for the second event is placed at c. 1860 Ma, although a phase of tectonic re-activation occurred locally at c. 1725 Ma. In Uganda, rocks of Ubendian age form the most extensive of the cover formations and are predominantly argillaceous with some gneisses but with quartzites at or near the base of the system (basal quartzites), which are thickest around the Rwenzori and thin towards the east.

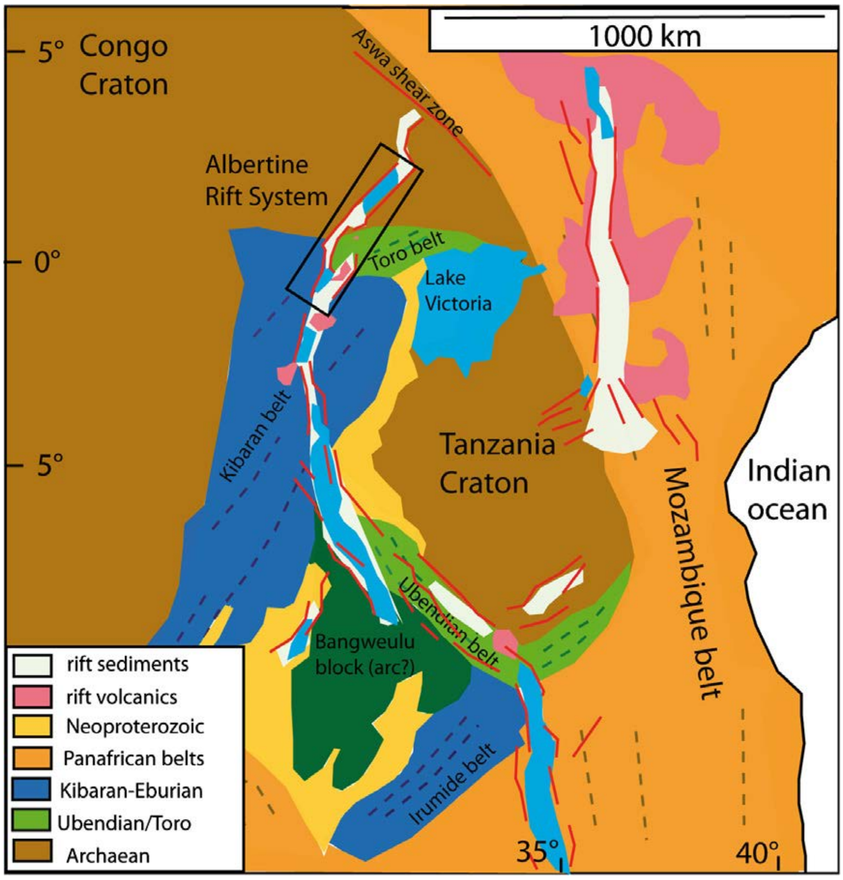


FIGURE 8: Simplified relative location of cratons and mobile belts constituting the basement structure of East Africa prior to rift development

The Mesoproterozoic Kibaran orogeny, is a short-lived but prominent c. 1375 Ma tectono-magmatic event (Aanyu and Koehn, 2011 and the references therein). The rocks of this orogeny form the Karagwe–Ankole belt (KAB) northeast of the Ubende belt-Rusizian basement extension and east of the western rift; and the Kibara belt (KIB) which is the domain occurring southwest of the Ubende belt-Rusizian basement extension . It is indicative of intra-cratonic regional-scale emplacement under extensional stress regime and the mantle-derived magmas penetrated a zone of weakness at the rheological boundary between the Archean Tanzania Craton and the adjacent Paleoproterozoic basement (2100 Ma mobile belt) to the left, both of which were overlain by Mesoproterozoic (meta)sedimentary rocks.

Since rocks of both the Paleoproterozoic Ubendian and Mesoproterozoic Kibaran belts in this region are strongly deformed with polyphase faults and ductile shear zones, they must have a profound effect on the younger deformational episodes including rift formation and evolution, during which the Rwenzori massif develops.

The East African Orogen (EAO; Stern, 1994) is a Neoproterozoic– early Cambrian mobile belt that today extends south along eastern Africa and western Arabia from southern Israel, Sinai and Jordan in the north to Mozambique and Madagascar in the south. The East African–Antarctic orogen defines a broad belt of 650–500 Ma orogenesis that can be traced within Gondwana from northeast Africa through South Africa to East Antarctica (Figure 9). It likely incorporates multiple phases of collision and accretion and collectively defines one of the largest and most continuous orogenic belts within Gondwana. It formed during the waning stages of the Proterozoic when Gondwana was nearing the final stages of its formation.

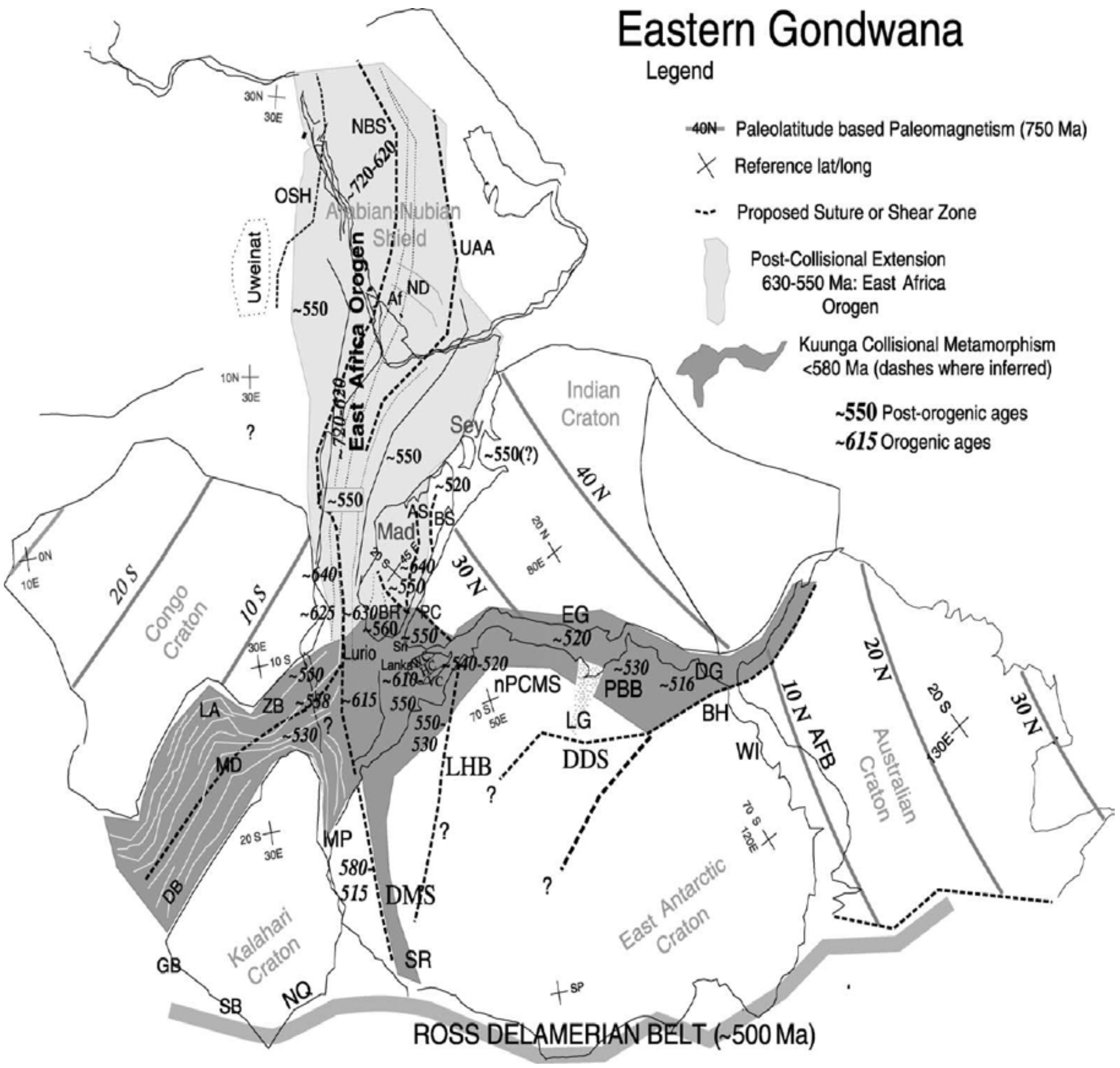


FIGURE 9: Geographic distribution of age provinces within eastern Gondwana showing the regions of 650– 500 Ma East African Orogen (light shading), 570– 530 Kuunga collisional metamorphism (darker shading). Reconstruction of Gondwana illustrating the extent of the East African–Antarctic Orogen.

Amongst all, the EAO is given here a particular interest as the EARS rift most of this orogeny (Figure, 10). Traditionally, the EAO is subdivided into the Arabian–Nubian Shield (ANS) in the north, composed largely of juvenile Neoproterozoic crust (e.g. Stern, 1994) and the Mozambique Belt (MB) in the south comprising mostly pre-Neoproterozoic crust with a Neoproterozoic–early Cambrian tectonothermal overprint (Fritz et al., 2013; Figure 10). The later is now changed to Eastern Granulite – Capo Dolgado Nappe Complex (CDNC, for e.g. Fritz et al., 2013).

The EARS preferentially splits the EAO and specifically the ANS. This notion can be attributed to the juvenile crust of ANS and rheologically acted as softer compared to the thick and stiff crust of predecessor orogenies. The EARS was unable to pass as a narrow belt to the south rather diverged in Tanzania in Tanzania Divergence Zone. The stiffness of the Eastern Granulite – Cabo Delgado Nappe Complex due to the presence of reworking crust and the superimposing of Kuunga Orogeny.

5. THE WESTERN RIFT – IMPLICATION FOR GEO-THERMAL EXPLORATION STRATEGY

The western branch of the EARS is an arcuate narrow rift characterized by a series of elongated and deep half-graben basins, developed in a Precambrian basement, and separated from each other by transfer zones. The Western rift system generally follows older orogenic belts and avoids the Archaean cratons.

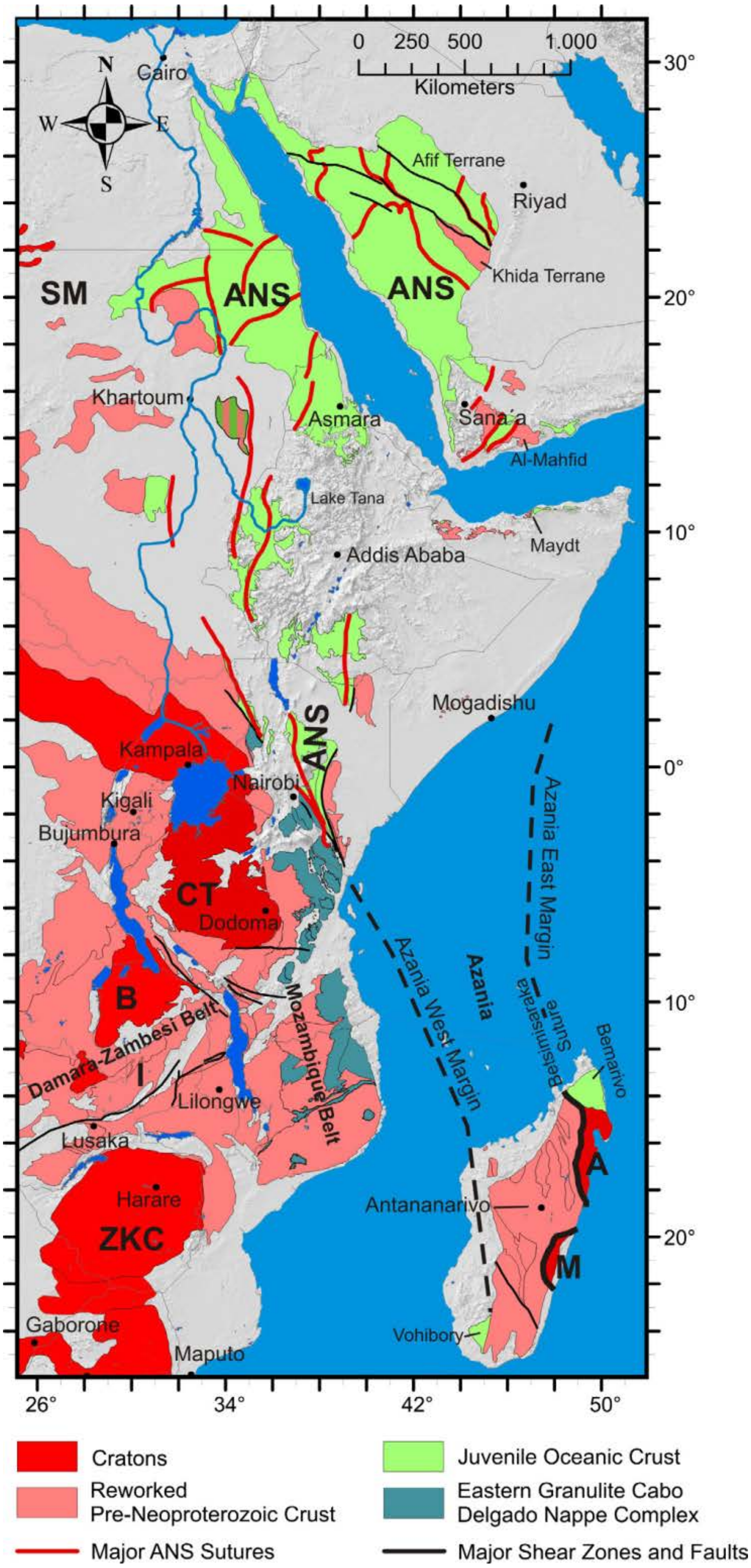


FIGURE 10: Distribution of crustal domains in the East African Orogen. SM, Sahara Metacraton; CTB, Congo–Tanzania–Bangweulu Cratons; ZKC, Zimbabwe–Kalahari Cratons; I, Irumide Belt; A, Antogil Craton; M, Masora Craton; ANS, Arabian Nubian Shield.

The volcanic activity in the western branch of the EARS is less well-developed compared with the eastern branch and the rift development is also at its initial stage. Therefore the geothermal exploration mechanism is rather different from the well developed Eastern branch of the EARS. Besides, the rift basins are filled with thick detrital sediment deposits compared to the thin cover of that of the eastern arm. Since the rift stage is at its inception stage, the geothermal exploration strategy should not be the same as that of the Eastern branch of the EARS. Therefore each region of the Western Branch of the EARS, from north to south is discussed mainly based on its tectonic development.

5.1 Ruwenzori region

The 5000 m high Rwenzori Mountains are situated within the Albertine rift (Figure 11), which is part of the western branch of the EARS (Figure 10). They represent a non-volcanic basement block composed of rocks of Proterozoic and Achaean age (Sachau et al., 2011, and the references therein).

Since most of the geothermal surface manifestation of Uganda and possibly some of Congo occurs surrounding the Rwensori mountains, thorough investigation and characterization of tectonic evolution is necessary for exploration of geothermal. Recent studies on Ruwenzori now incorporated in this report to better understand the tectonics and so its impact to geothermal.

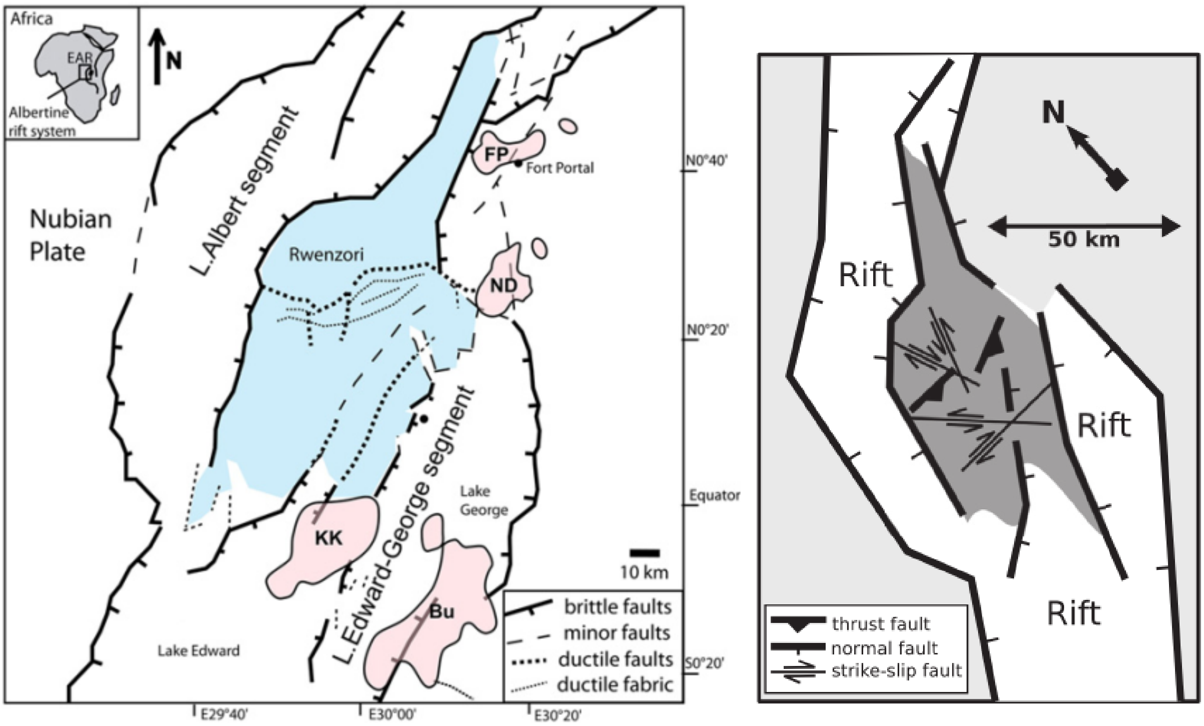


FIGURE 11: a) Tectonic map of the Rwenzori area (left). The Rwenzori mountains (light blue area) are surrounded by two rift segments, with the exception of a region in the NE, where they are connected to the Victoria plate. Volcanic fields are marked by red areas (FP: Fort Portal; ND: Ndale; KK: Katwe-Kikorongo; Bu: Bunyaruguru); b) The strike-slip systems indicate roughly N–S and E–W compression, respectively (Sachau et al., 2011) (right). Thrust faults indicate a vertical r-vector, also associated with N–S compression. Normal faults illustrate the NW–SE oriented rift opening.

Most of the studies conducted so far indicate that the tectonic regime of Rwenzori to uplift within extension regime. But recently (Sachau et al., 2011) deduced a strike-slip faulting with rotational component. The stress field resulting from the computer simulations explains previously unexplained key structural features of the Rwenzori mountains. These structures are N-ward oriented thrust faults and two different sets of strike-slip structures indicating a rotation of the stress-field with N–S and E–W compression, respectively, in the central block.

The observed earthquake swarms in the Rwenzori area has been interpreted to be originate from crustal fluid migrations and/or CO_2 emanations rising from a magmatic body in the upper mantle (Figure 12). From the above observation and the compositions and isotope ratios of the mineral waters and their incorporated gases also point to a mixing with a magmatic or mantle related gas flow and water source (Bahati et al., 2005; Lindenfeld et al., 2012, and the reference therein). However the postulated magmatic body as supported by the S-wave receiver function provides evidence of two consecutive reduction in shear wave speed at depths between 55-80 km and 140-160 km, interpreted as the shallower discontinuity likely marks the upper bound of an altered mantle lithosphere that might be composed by melt infiltrations and the later deeper discontinuity is interpreted as the lithosphere–asthenosphere boundary, respectively (Figure 13).

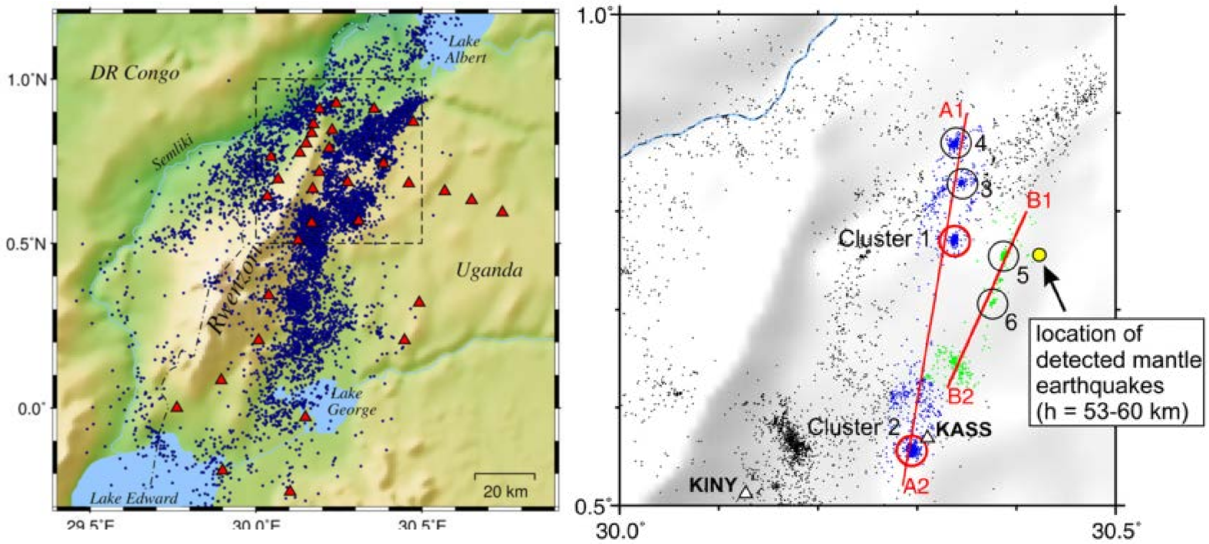


FIGURE 12: Seismic station network (triangles) and recorded local seismicity from February 2006 to September 2007 (from Lindenfeld et al., 2012) (left); b) The central region for which earthquakes are relocated by the HypoDD method (right).

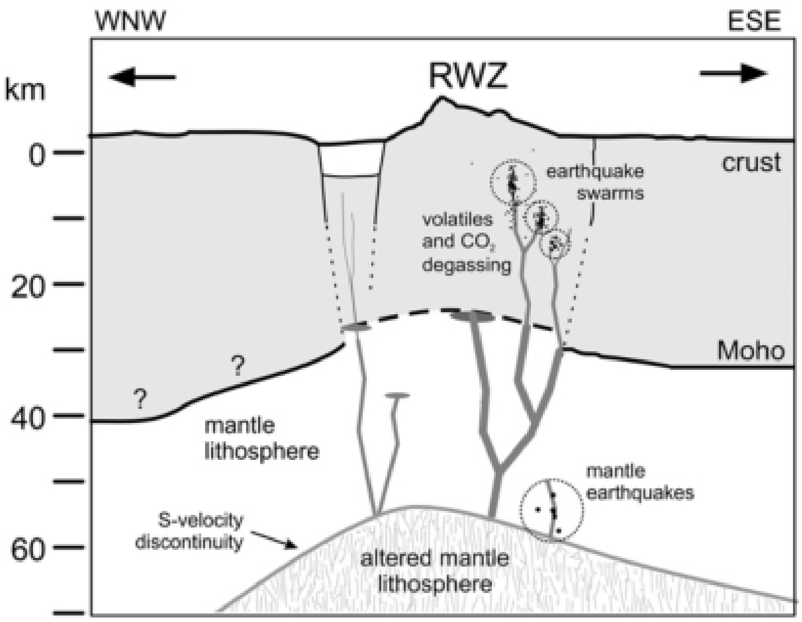


FIGURE 13: Cartoon illustrating the system of magma and fluid transport in the lithosphere beneath the Rwenzori region. The locations of mantle earthquakes and earthquake swarms are indicated. The depths of the Moho and of an upper mantle discontinuity are documented by P- and S-wave receiver functions (Lindenfeld et al., 2012, and the reference therein).

5.2 Regions of transfer zone and volcanism

The volcanic activity in the western branch as mentioned above is at its incipient stage and is restricted to four transfer zones (Ebinger, 1989), these are from north Toro-Ankole, Virunga volcanic province, south Kivu and Rungwe (Figure 14).

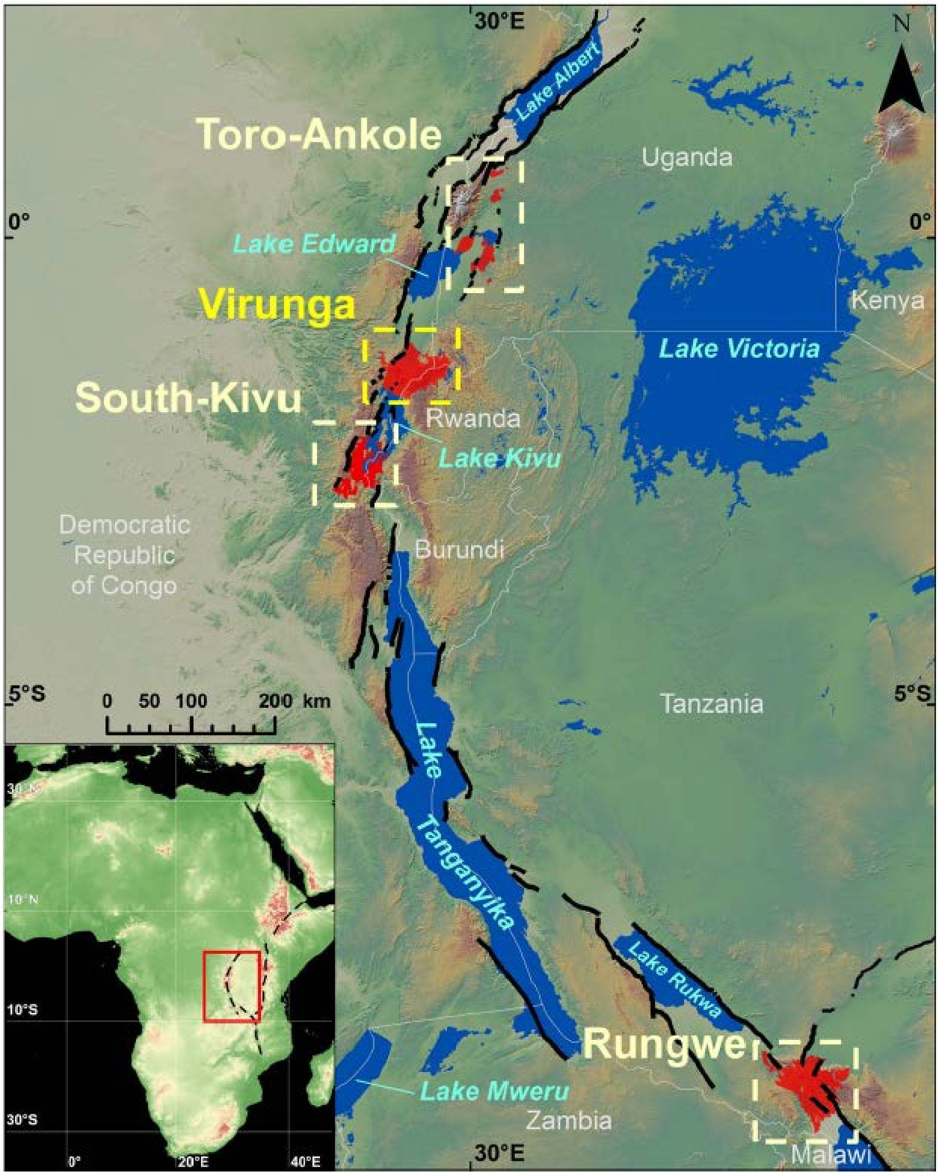


FIGURE 14: The western branch of the East African Rift System and its four volcanic provinces. The inset shows the location of the EARS in Africa.

Within narrow rift-related transfer zones, the presence of the underplated magma localizes again strain in the overlying crust, thus influencing the surface fault pattern; on the other hand, localised deformation causes magma to accumulate in correspondence to the transfer zone, with a main flow pattern that is perpendicular to the extension direction. The resulting patterns of magma emplacement and deformation may offer insights for explaining the correspondence to the volcanic provinces with major transfer zones in the Western Branch of the East African Rift System (Corti et al., 2003). Thus these transfer zones may contribute to the rifting process by sourcing upper crustal dikes that propagate laterally into the tips of rift basins.

A detail study on the relationship of transfer zone and magmatism is presented by Muirhead and others. 2015. They developed a conceptual model on magmatic architecture of the transfer zone with implication of the magma-driven rifts, which is quite important in geothermal exploration (Figure 15). Stress fields induced by magma reservoirs produce radial dikes, whereas extension-oblique dike intrusions form from the mechanical interaction between segmented rift basins. Extension-normal dikes occur at the boundary between transfer zones and the ''distal ends'' of rift basins, where their orientations are controlled by the regional extensional stress field. This model can be used in geothermal exploration across the western rift.

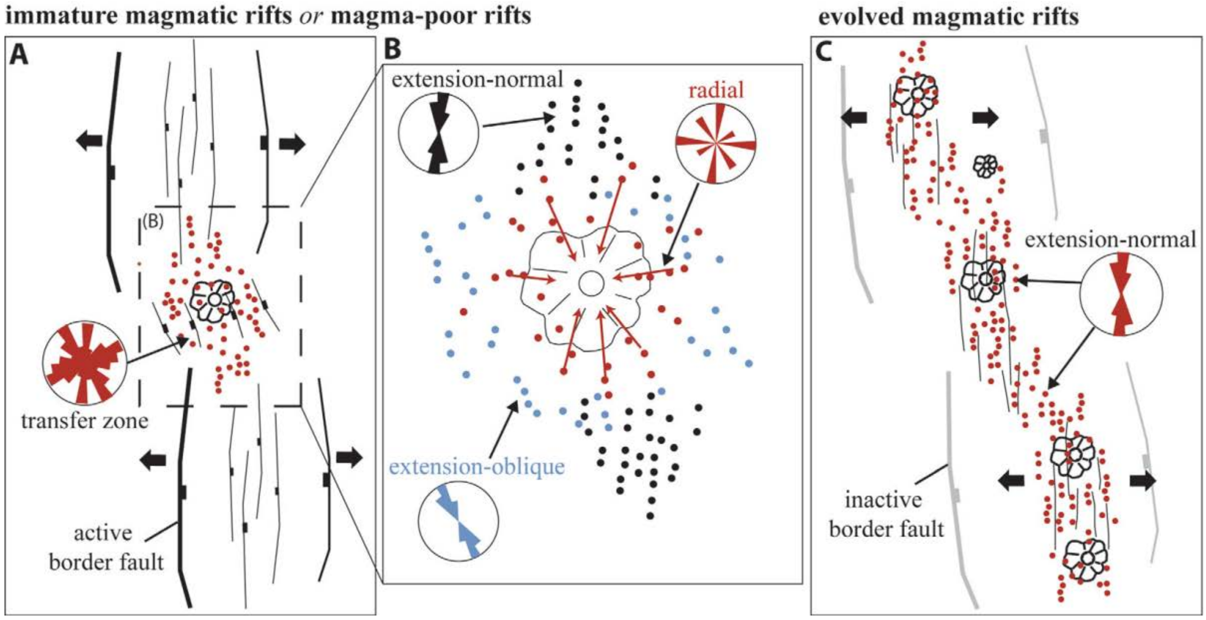


FIGURE 15: Schematic representation of the distribution of cones and their trends (Muirhead, 2015). Rose diagrams provide a conceptual example of the expected cone lineament trends in the different structural settings. Black arrows represent the regional extension direction (E-W in this example). Figure 13a shows immature magmatic rifts and magma-poor rifts. Cones (red dots) focus in transfer zones and the distal tips of basins. B shows the range of cone trends distributed within transfer zones in immature magmatic rifts and magma-poor rifts.

As explained above, almost all the volcanism in the western branch of the EARS are localized within the transfer zone. The Virunga-Volcanic Province, South Kivu and Tore Ankole volcanic areas are places are amongst this category. Thus detail study of the tectonism of the transfer zone is necessary to outline and explore targets of geothermal resources present if any. In this context the relationship between the alignment of the rift and the transfer zone is necessary to characterize the geothermal manifestations.

5.2.1 Regions of Virunga Volcanic Province (VVP) and Kivu Region

The VVP is one of the four volcanism corresponds to the transition between the basins of Lake Edward and Lake Kivu, which are two half-graben basins with the main rift faults forming the western border. Rifting in this part of Kivu basin have reactivated NE-SW oriented weakness structures of the Mesoproterozoic Karagwe-Ankole Belt, which represent the Precambrian basement in which the rift propagated.

This observation suggests that this cone alignment might represent, for this section of the rift, the trace at ground surface of the main active faults located at depth. Except for the 1938–1940 eruption, atypical eruptions consistently occur N15 km away from the caldera centre, at the northern or southern ends of the N50°E cone alignments (Figure 16). This is in line with the direction of the transfer zone, which is also follows, the Precambrian basement structure.

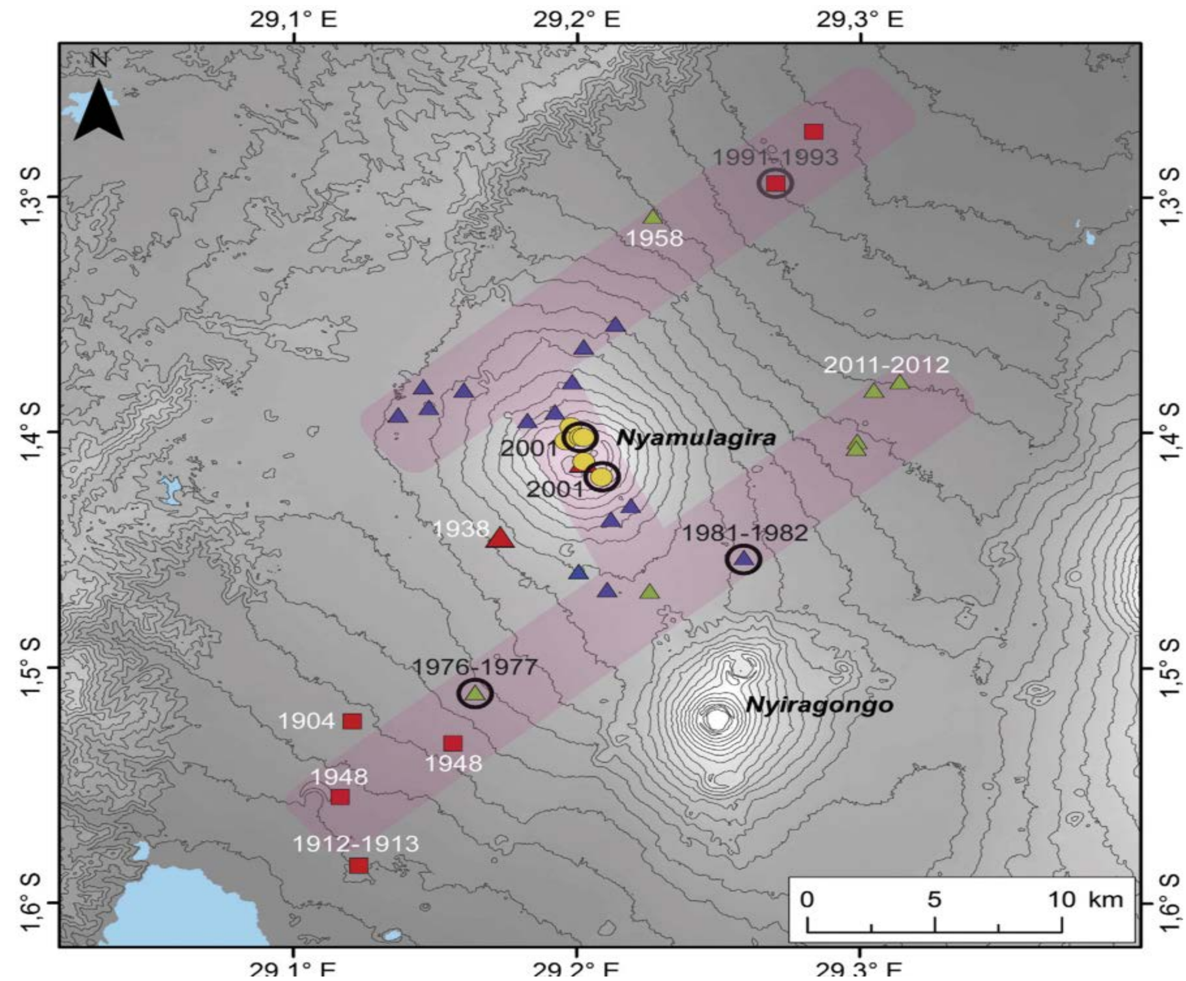


FIGURE 16: Precambrian structures also promote the preferential location of some eruptive vents along N50°E-orientated axes located north and south of the main edifice of VVP.

The continuity between the VVP and SKVP, characterized by volcanic cones continuing into Lake Kivu, strongly suggests that volcanism in this region has a common origin (Smets et al., 2016). This observation is consistent with the hypothesis of a magma underplating below the Kivu region (Corti et al., 2003).

Extension-normal cones (black circles, black rose diagram) occur near the boundary between the transfer zone and the distal tip of the basin (Figure 17).

A potential implication is that the injection of extension-normal, upper crustal dikes laterally from transfer zones into the rift axis may facilitate magmatic rifting in both magma-poor and immature magmatic rifts of the EAR, albeit localized to the distal ends of these basins. This indication of the presence dyke is in agreement with the concentration of hydrothermal springs on roughly the same spot suggests that the area is well positioned at the intersection of the N-S alignment of the cones and the transfer zone of the VVP. Thus enhances the possibility of resource target of geothermal exploration.

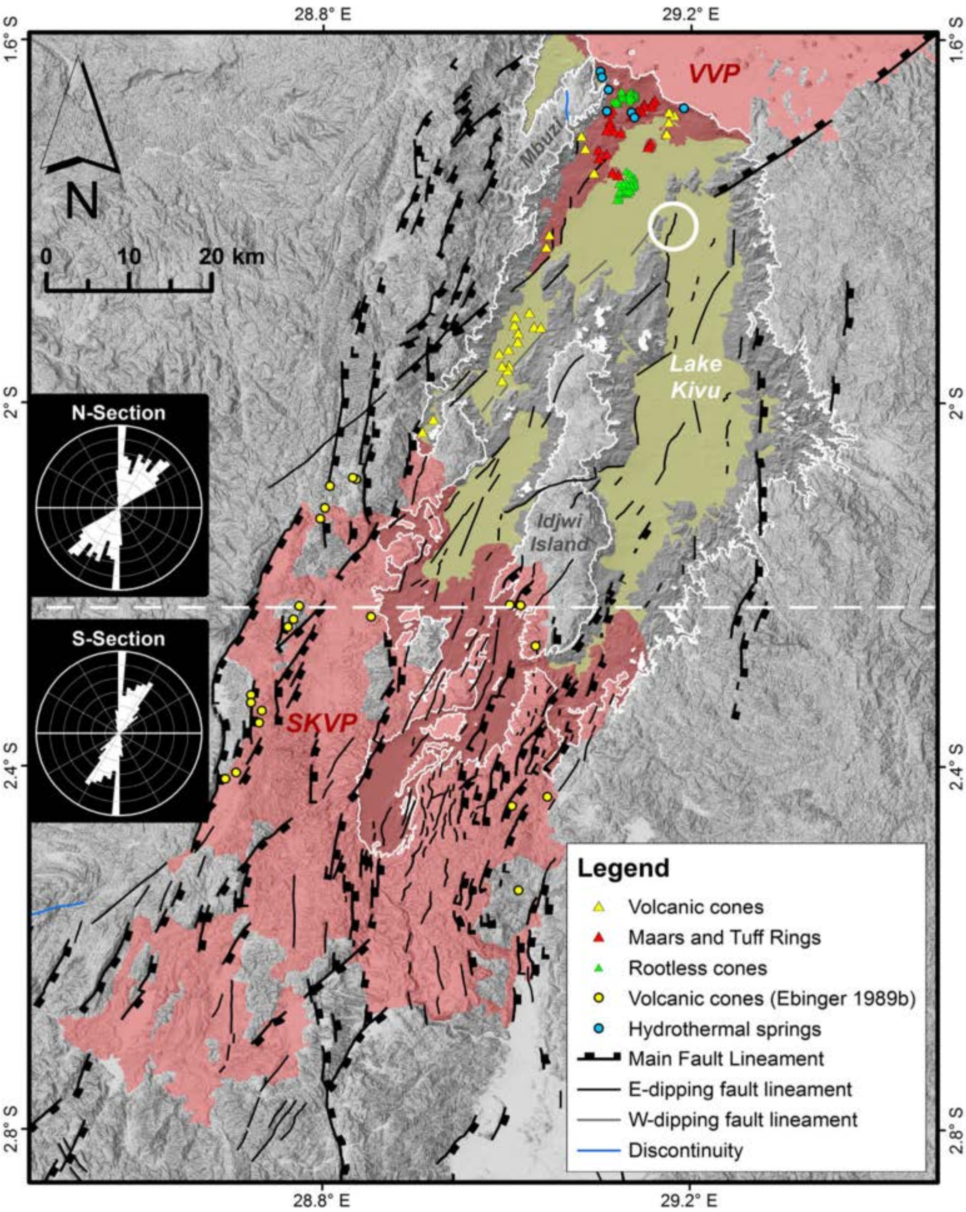


FIGURE 17: Simplified geological map of the Lake Kivu basin realized in this work. Areas coloured in grey, red and yellow are Precambrian rocks, lava fields and lacustrine sediments, respectively.

5.2.2 Buragama Rusizi Basin

Upper Miocene basalt flows cover the accommodation zone between the Kivu and Rusizi extensional basins, a region where transfer faults link basins bounded by border faults on opposite sides of the rift valley (Ebinger, 1989). Within this accommodation zone, the narrow Bugarama graben is bounded to the east by seismically active normal fault (Figure 18).

- High-angle (45^o-75 ^{ø}) normal faults separated by 1-5 km occur in a 10- to 15-km•wide zone along one side of asymmetric basins. These planar border fault systems with large throws (1-6 km) may penetrate the crust.
- Three structural relations consistently observed within the Kivu and Rusizi regions indicate that zones of crustal thinning are limited to rift basins bounded by approximately 100-km-long border fault segments.
- The Western rift system generally follows older orogenic belts and avoids the Archaean craton.

5.2.3 North Tanzanian Divergent Zone

Cone lineament data presented in the regional study by Muirhead et al. (2015) reveal spatial variations in the distribution and geometries of upper crustal dike intrusions across rift basins in the East African Rift. Upper crustal diking in the North Tanzania Divergence is focused in a transfer zone. Dike intrusions here rarely occur along the central axis of rift basins (e.g., Natron-Magadi basin), suggesting that dikes play a limited role in accommodating upper crustal extension in this part of the rift.

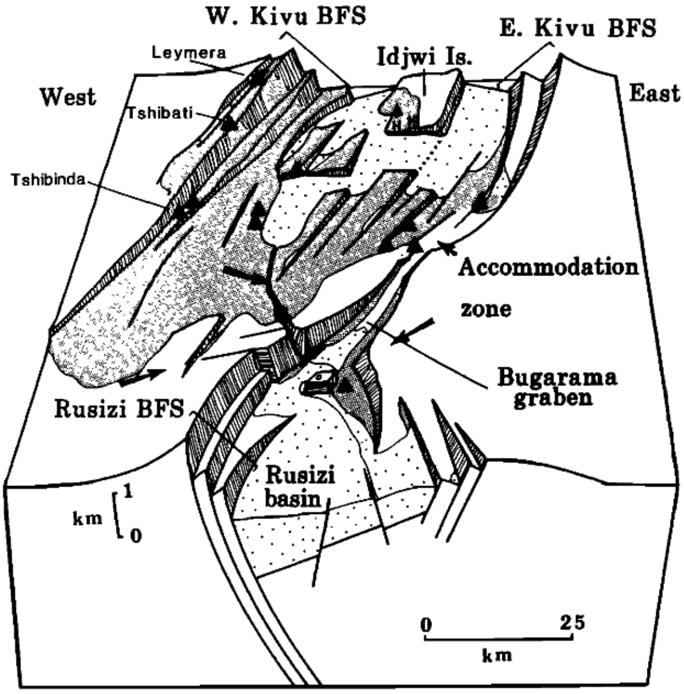


FIGURE 18: Three-dimensional model of spoon-shaped rift basins and accommodation zone between basins (Ebinger, 1989).

The presence of NW-SE trending, extension-oblique cone lineaments in the North Tanzanian Divergence can be explained as the result of either dikes intruding preexisting, transverse structures and/or local rotations of tension from rift segment interactions (Muirhead et al., 2015, and the references therein).

Transfer zones in the EAR in general and this area exhibit complex dike geometries that can be attributed to local stress fields related to pressurized magma chambers and mechanical interactions between rift segments, in addition to the regional stress field. These transfer zones may contribute to the rifting process by sourcing upper crustal dikes that propagate laterally into the tips of rift basins.

Proterozoic basement in Kenyan and Tanzanian regions of the EAR contains NW-SE striking shear zones , act as transfer zones, which are aligned locally with recently active, transverse faults (Smith and Mosley, 1993).

REFERENCES

Aanyu, K., Koehn, D., 2011: Influence of pre-existing fabrics on fault kinematics and rift geometry of interacting segments: Analogue models based on the Albertine Rift (Uganda), Western Branch - East African Rift System. *Journal of African Earth Sciences, 59*, 168-184.

Bahati, G., Pang, Z.H., Ármannsson, H., Isabirye, E.M., Kato, V., 2005: Hydrology and reservoir characteristics of three geothermal systems in Western Uganda. *Geothermics, 34*, 568–591.

Braile, L., Keller, G., Wendlandt, R., Morgan, P., and Khan, M., 2006: The East African Rift System. In: Olsen, K.H. (editor), *Continental rifts: Evolution, structure, tectonics*. Elsevier Science B.V., Amsterdam, Netherlands, 466 pp.

Chorowicz, J., 2005: The East African Rift System. *Journal of African Earth Sciences, 43*, 379–410.

Chorowicz, J., Le Fournier, J., Vidal, G., 1987: A model for rift development in Eastern Africa. *Geological Journal, 22*, 495–513.

Corti, G., Bonini, M., Conticelli, S., Innocenti, F., Manetti, P., Sokoutis, D., 2003: Analogue modelling of continental extension: A review focused on the relations between the patterns of deformation and the presence of magma. *Earth Sci. Rev., 63*, 169–247.

Craig, T.A., Jackson, J., Priestley, K., and Kenzie, D., 2011: Earthquake distribution patterns in Africa: Their relationship to variations in lithospheric and geological structure, and their rheological implications. *Geophys. J. Int., 185*, 403–434.

Ebinger, C.J., 1989: Geometric and kinematic development of border faults and accommodation zones, Kivu-Rusizi Rift, Africa. *Tectonics, 8*, 117–133.

Fritz, H., Abdelsalam, M., Ali, K., Bingen, B., Collins, A., Fowler, A., Hauzenberger, C., Johnson, P., Kusky, T., Macey , P., Muhongo, S., Ghebreab, W., Stern, R., Viola, G., 2013: Orogen styles in the East African Orogen: A review of the Neoproterozoic to Cambrian tectonic evolution. *Journal of African Earth Sciences, 86*, 65-106.

Keranen, K., and Klemperer, S.L., 2008: Discontinuous and diachronous evolution of the Main Ethiopian Rift: Implications for development of continental rifts. *Earth and Planetary Science Letters, 265*, 96– 111.

Lindenfeld M., Rümpker, G., Link, K., Koehn, D., Batte, A., 2012: Fluid-triggered earthquake swarms in the Rwenzori region, East African Rift – Evidence for rift initiation. *Tectonophysics, 566-567*, 95-104.

Macgregor, M., 2015: History of the development of the East African Rift System: A series of interpreted maps through time. *Journal of African Earth Sciences, 101*, 232–252.

Meghraoui, M., 2016: The seismological map of Africa. *Episodes, 39*, 9-18.

Muirhead, J.D., Kattenhorn, S.A., and Le Corvec, N., 2015: Varying styles of magmatic strain accommodation across the East African Rift. *Geochem. Geophys. Geosyst., 16*, 2775–2795.

Sachau, T., Koehn, D., and Passchier, C.W., 2011: Lattice-particle simulation of stress patterns in a Rwenzori-type rift transfer zone. *Journal of African Earth Sciences, 61*, 286-295.

Smets, B., Delvaux, D., AnnRoss, K., Poppe, S., Kervyn, M., d'Oreye, N., and Kervyn, F., 2016: The role of inherited crustal structures and magmatism in the development of rift segments: Insights from the Kivu basin, western branch of the East African Rift. *Tectonophysics, 683,* 62-76.

Smith, M., 1994: Stratigraphic and structural constraints on mechanisms of active rifting in the Gregory rift, Kenya. *Tectonophysics, 236*, 3-22.

Smith, M., and Mosley, P., 1993: Crustal heterogeneity and basement influence on the development of the Kenya Rift, East Africa. *Tectonics, 12*, 591–606.

Stern, R.J., 1994: Arc assembly and continental collision in the Neoproterozoic EastAfrican Orogen — implications for the consolidation of Gondwanaland. *Annual Reviews of Earth and Planetary Sciences 22*, 319–351.

WoldeGabriel, G., Aronson, J.L., Walter, R.C., 1990: Geology, geochronology and rift basin development in the central sector of the main Ethiopia Rift. *Geological Society of American Bulletin 102*, 439–458.