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Listric growth faults in the Kenya Rift Valley

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Abstract—Many of the major faults in the Kenya Rift Valley are curved in section, were active over considerable periods and form sets which are related in space and time. They can, therefore, be regarded as systems of listric growth faults. The Elgeyo Fault marks the western limit of rift structures at this latitude and displaces the basement surface by up to about 6 km. The Kamasia Hills are a block rotated above this fault plane. Movement on the Elgeyo Fault has been grossly continuous since at least 16 Ma ago but deposition of volcanics and sediments has generally kept pace with the growth of the escarpment. The Kaparaina Arch is a rollover anticline on the downthrown side of the Saimo Fault on the eastern side of the Kamasia Hills. On the eastern side of the rift, the block between the Bogoria and Wasages-Marmanet Faults has shown continued rotation since about 15 Ma. The Pleistocene lavas on the rift floor here show rollover into the Bogoria Fault and have formed a facing near the top of the escarpment. Area balancing calculations suggest depths to décollement of 25 km for the Elgeyo Fault, 6 km for the Saimo Fault and 12 km for the Bogoria Fault. The most direct evidence for the listric nature of the faults is provided by microearthquakes near Lake Manyara which appear to lie on fault planes connected to surface escarpments.

INTRODUCTION

THE purpose of this paper is to show that at least some of the major faults in the Kenya Rift Valley belong to systems of related listric growth faults. This necessitates demonstration that the fault planes are concave upwards, that movement on them has been continuous over long periods and that they are related in time and space.

The first suggestion that faults in the Kenya Rift might be listric was the proposal by Shackleton (1951) of a curved Elgeyo Fault to account for the greater dip of the lavas in the Kamasia Hills on the downthrown side than of their equivalents on the upthrown Uasin Gishu Plateau. More recently the idea has gained currency that the dips of fault planes are perpendicular to the dips of the truncated beds and hence that faulting is associated with rotation of the fault blocks (Chapman *et al.* 1978). King (1978) has proposed that the tilted fault blocks of the Kamasia area may rest on a major fault plane which flattens at depth.

Temperley (1966) drew attention to the common occurrence of faced scarp structures on the eastern side of the Kenya Rift Valley, explaining them as due to rejuvenation of movement on old fault planes. Since then multiple periods of movement on the rift bounding faults have been reported from several areas of the rift, e.g. west of Lake Magadi (Crossley 1979), on the Bogoria Fault (McCall 1967, Griffiths 1980), in the Kamasia area (Chapman *et al.* 1978) and in the Kedong–Olorgesaile area (Baker & Mitchell 1976).

McCall (1967) classified the fault scarps of the eastern central part of the rift into three major groups each characterized by a similarity of spacing and degree of erosion, and hence all considered to be of the same age. Two later minor phases of faulting were confined to the rift floor. Subsequent authors have subscribed to the concept of distinct phases of faulting affecting several faults simultaneously and migrating progressively towards the axis of the rift (e.g. Baker & Mitchell 1976, Chapman *et al.* 1978, Crossley 1979). The idea of movement occurring on several related faults at the same time is thus now well established.

This paper is based on a review of the literature on the Kenya Rift Valley. It attempts to synthesize previously reported ideas and observations in the light of the literature on listric growth faulting. The most direct way of observing the listric nature of faults is with reflection seismology but unfortunately there are no publicly available seismic reflection data nor relevant deep well data from the Kenya Rift. The evidence used in this paper is largely taken from geological mapping combined with geochronological and geophysical data, principally earthquake seismic and gravity data.

Two areas are described in detail, one comprising the Elgeyo Escarpment and Kamasia Hills, the other centred on Lake Bogoria (Fig. 1). Brief consideration is given to four other areas which provide further useful evidence. These are the Lake Manyara area, the Kirikiti-Lengitoto Fault, the region west of Lake Turkana and the Mzima Fault.

THE ELGEYO–KAMASIA AREA

The western margin of the Kenya Rift Valley north of the Equator is formed by the Elgeyo Escarpment, a fault scarp 1000–1500 m high extending for about 100 km (Fig. 2). It is oriented generally N–S except for a NW-trending 15 km stretch at 0° 30′ N latitude. The Kerio Valley at its foot separates the fault scarp from the Kamasia Hills 25–30 km to the east. The Kamasia Hills may be regarded as a strongly faulted anticline, the principal faults being the east side downthrown en éche-

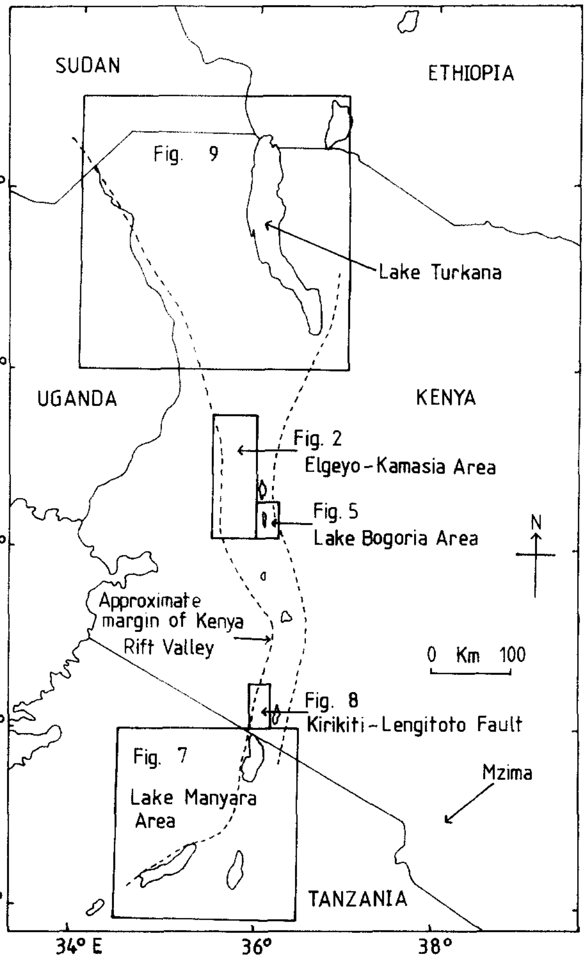


Fig. 1. Index map of Kenya Rift Valley showing the locations of areas discussed.

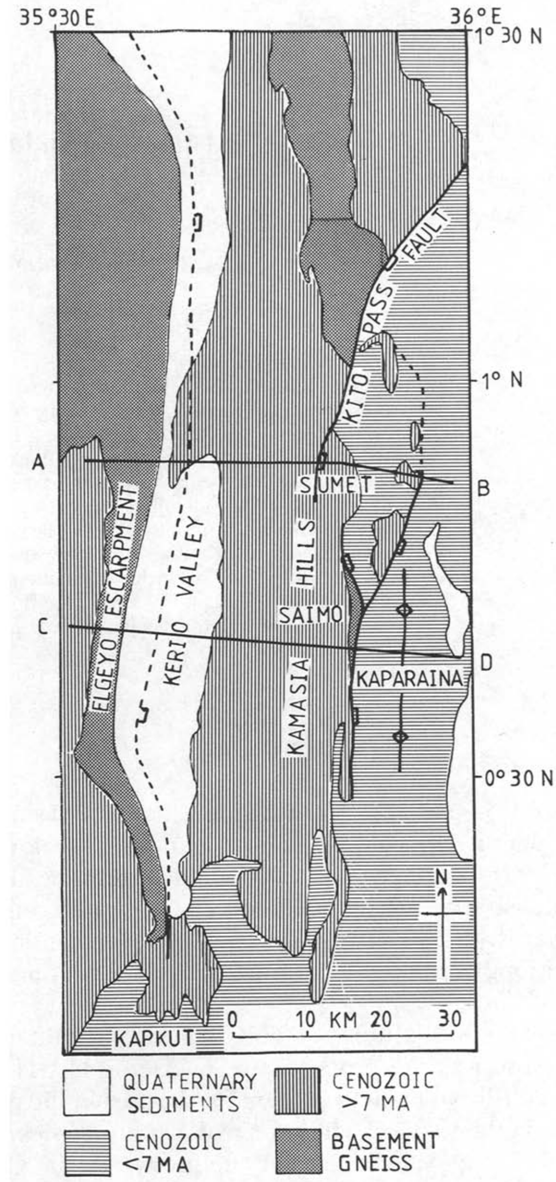


Fig. 2. Geological map of the Elgeyo–Kamasia area, based on Chapman *et al.* (1978). The Elgeyo Escarpment defines the western margin of the Kenya Rift Valley. The Kamasia Hills are a westward tilted block between the Elgeyo Fault under the Kerio Valley and the Saimo and Kito Pass Faults to the east. The Kaparaina Arch is a rollover on the hangingwall of the Saimo Fault. A–B and C–D: lines of sections in Fig. 4.

lon Kito Pass and Saimo Faults. The Sumet area between them shows a complex pattern of more minor faults, some of them strongly curved. North and south of the major faults the structure passes into E-dipping monoclines. The Kamasia Hills are connected to the Elgeyo Escarpment by the Kapkut Highlands at the southern end of the Kerio Valley.

The Elgeyo Fault

Most of the height of the Elgeyo Escarpment is formed of basement gneisses of the Mozambique 'System'. The steeply dipping foliation of the gneisses closely parallels the trend of the escarpment, including a change of direction at 0° 30′ N, strongly suggesting control of the Cenozoic fault by basement structures (Fig. 2).

West of the escarpment the surface dips gently westwards at only about 1° and earthquake studies have shown that the crust has a normal continental structure (Maguire & Long 1976). The anomalous deep structure under the Rift Valley begins at the Elgeyo Escarpment, so this may be taken as marking the precise boundary of the rift structure.

The basement rocks are overlain in the escarpment by sediments and volcanics dating back to about 16 Ma, capped by the Uasin Gishu Phonolite flows of 14.5–12 Ma. The Uasin Gishu Phonolite outcrops extensively to the west but the underlying Cenozoic rocks wedge out rapidly. The sub-Cenozoic surface shows considerable topographic irregularity but with a general rise westward to a culmination somewhere beneath the phonolite flows. This has generally been interpreted as an eroded monocline which was progressively buried by the Cenozoic rocks and finally overstepped by the Uasin Gishu Phonolite (Chapman *et al.* 1978). However it might equally well have been an eroded fault scarp in which case the earliest movement on the Elgeyo Fault would be dated at >16 Ma.

The northern end of the fault is marked by the abrupt decrease in height of the escarpment at 1° 20′ N and the appearance of basement rocks on both sides of the fault at 1° 45′ N. At its southern end the fault dies out rapidly south of 0° 20′ N and passes into a set of splay faults and then a monocline (Fig. 2).

The throw of the fault must be at least 3000 m at 0°30′N extrapolating from the dip and thickness of the Cenozoic succession in the Kamasia Hills (Chapman *et al.* 1978). However, Swain *et al.* (1981) showed, on the basis of gravity traverses at 0°30′N and a seismic refraction line at 0°30′N, that the basement was about 3000 m below sea level on the downthrown side of the fault at these latitudes, giving a throw of about 5000 m. Their gravity map shows a negative anomaly along the Kerio Valley, with its lowest values of −60 mgal at 0°40′N, between the two traverses along which the anomalies are both about −40 mgal. This indicates that the greatest throw on the fault is much more than 5000 m and coincides with the concave bend in the fault trace.

The model section for 0°30′N presented by Swain *et al.* (1981) shows that about 10 km in front of the Elgeyo Fault the basement surface reaches a culmination from which it slopes gently eastwards but increasingly steeply westwards reaching 20° at the fault plane. In their model at 1°N the highest point on the basement surface lies 20 km in front of the Elgeyo Fault and is downfaulted on its eastern side but slopes with an increasing westward dip towards the Elgeyo Fault.

The Ngorora Beds are sediments of 12.5–10 Ma which overlie phonolite lavas in the Kamasia Hills equivalent in age to the Uasin Gishu Phonolite. Detailed sedimentological study of these beds has shown that the Elgeyo Escarpment was present and intermittently growing throughout their period of deposition (Pickford 1978a). The overlying Ewalel Phonolite of 9–7 Ma was erupted from a source near the southern end of the Elgeyo Fault and flowed against the fault scarp.

Chapman *et al.* (1978) have noted that the western side of the Kamasia Hills is largely the dip slope of the 7 Ma old Kabarnet Trachyte. West of Saimo the surface is convex, being horizontal for 10 km from the Saimo Fault and then dipping towards the Elgeyo Escarpment. This form resembles the underlying basement surface although the hinge line is further west at the lower horizon. North and south from Saimo the Kabarnet Trachyte has a gently concave surface. At 1°N the trachyte has a gentle easterly dip close to the foot of the escarpment, the relative youthfulness of this probable fault drag being attested by the Kerio River which runs in a sharply incised bed between the trachyte outcrop and the escarpment rather than along the present axis of the depression which lies further east.

The model sections of Swain *et al.* (1981) show that although the basement surface falls towards the Elgeyo Fault the thickness of lavas on the downthrown side under the Kerio Valley is much the same as exposed in the Kamasia Hills. The extra thickness is taken up with sediments. This suggests that sedimentation has consistently infilled the wedge shaped depression formed by movement on the Elgeyo Fault. Whenever lavas were erupted the surface of the basin was approximately horizontal so that the volcanics have a uniform thickness. The volcanic formations would represent much shorter time spans than the sediments. The Uasin Gishu Phonolite is the only formation to have successfully overstepped the fault scarp. This is the local representative of the Plateau Phonolites which overstepped the walls of the Rift Valley all over the Kenya Dome and it is significant that Fitch *et al.* (1978) derived a K–Ar isochron age of 12.50 ± 0.15 Ma using data from throughout the Plateau Phonolite outcrop. It seems that this huge volume of rock (about 25,000 km^3, Williams 1972) was erupted in a very short time.

The Kamasia Hills block has continually rotated while the Elgeyo block has remained horizontal. It follows that the Elgeyo Fault must be concave upwards at depth (Wernicke & Burchfiel 1982). While it is not possible to derive a detailed model of the form of the fault at depth, various lines of evidence provide clues. Techniques have been developed for calculating profiles of listric normal faults identified on seismic sections (Davidson 1986). The simplest is to take a profile perpendicular to the fault plane and calculate the area between a horizon on the downthrow side and the hypothetical unfaulted continuation of that horizon from the upthrown side. Division of this wedge shaped area by the heave of the fault gives the depth to the décollement.

The basement on Saimo reaches almost the same elevation as in the Elgeyo Escarpment and Swain *et al.* (1981) have modelled the basement surface in this area. The subsided cross-sectional area for the basement measured on Fig. 4, section C–D is 83 km^2. Using Swain *et al.*'s 63° dip for the fault plane and a throw of 6 km gives a heave of 3 km and hence a depth to décollement of 28 km.

On their two gravity model profiles Swain *et al.* (1981) show the basement surface dipping into the Elgeyo fault plane at 16° and 19° at a depth of about 4 km below the Kerio Valley or 5.5 km below the crest of the escarpment. From the well established principle that fault planes in the Kenya Rift Valley are perpendicular to the truncated bedding it follows that the Elgeyo fault plane is dipping at 90° − 17.5° at a depth of 5.5 km below the crest of the escarpment. Taking the crest of the escarpment as the original unfaulted surface and assuming that the fault has the form of a circular arc, the fault plane becomes horizontal at a depth of 18 km.

Maguire & Long (1976) reported one earthquake which may be associated with the Elgeyo Fault. This arrived at Kaptagat from the east with an apparent velocity of 9.6 km s^{-1} and a P–S separation of 2.7 s. This event would have travelled entirely within the upper crustal zone for which Maguire & Long (1976) give a P wave velocity of 5.8 km s^{-1}. From this information and the P to S wave velocity ratio of 1.74 which they used it follows that the focus was at a depth of 15 km and the epicentre 12 km from Kaptagat. The seismic station lies about 15 km west of the crest of the Elgeyo Escarpment but given the poor accuracy of epicentral locations possible with the data (Maguire & Long 1976) it is reasonable to infer that this earthquake originated on the Elgeyo Fault at a depth of the order of 15 km.

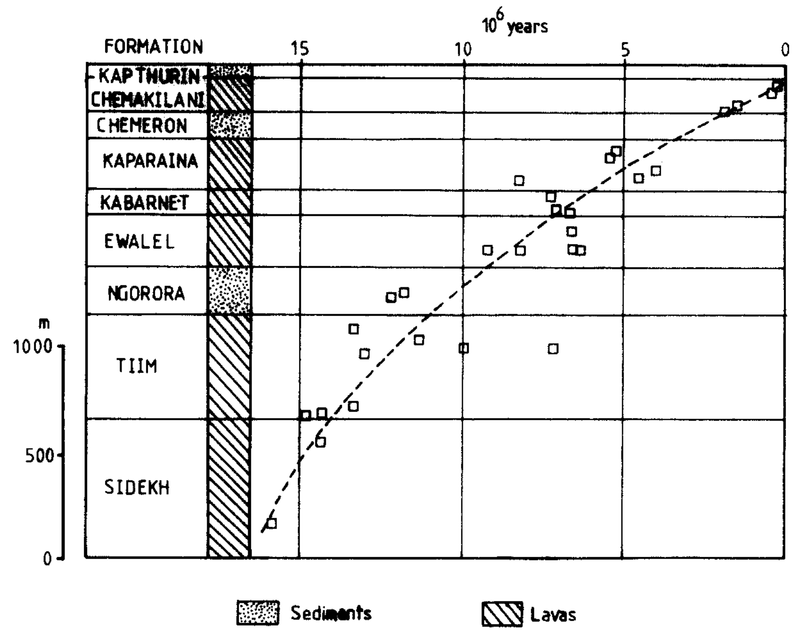


Fig. 3. Plot of accumulated thickness against K–Ar age for the Kamasia area, based on Chapman & Brook (1978, fig. 14:4) and Bishop *et al.* (1971, fig. 2). The rate of accumulation is fairly constant.

Of these three lines of argument the cross-sectional area balancing is most likely to approximate the true depth of décollement and so should be most strongly weighted. The depth at which the Elgeyo Fault flattens is thus of the order of 25 km below the crest of the escarpment. This depth is shown in Fig. 4.

Kamasia Hills

The Saimo Fault scarp exposes basement, reaching almost the same elevation as in the Elgeyo Escarpment, overlain by Cenozoic volcanic rocks beginning at 12 Ma (Fig. 2). Chapman *et al.* (1978) estimate from the thickness of displaced formations that the maximum throw on the fault is 4000 m but this might be an underestimate since the depth of the basement on the downthrown side is unknown. The Kito Pass Fault, for which Chapman *et al.* give a maximum throw of at least 2000 m, also exposes basement forming an arch trending NNW as far as the Elgeyo Escarpment.

Chapman *et al.* (1978, fig. 5) presented a structural block diagram of the Kamasia Hills which clearly illustrates the rapid growth and decline of the Saimo Fault from south to north. The two gravity model profiles of Swain *et al.* (1981) pass the two ends of the Saimo Fault, the southern one showing no displacement of the basement surface while the northern profile has the basement downfaulted to the east by 2.5 km in two steps in the Sumet area.

During the deposition of the Ngorora Beds the Saimo and Kito Pass fault blocks were positive areas but these sediments were deposited in the Sumet area in between. Exposures in this complexly faulted area show consistent thickening of subunits on the downthrown sides of faults and backtilting of the upthrown blocks with dips of up to 17° (Pickford 1978a, fig. 16:6). These are clearly rotating growth faults.

Chapman *et al.* (1978) believed that movement on the Saimo Fault was mostly confined to two events at 7 and 2–0.5 Ma. However, Pickford (1978a) stated that the Saimo and Sidekh areas stood above the Ngorora depositional basin at 12.5–10 Ma while faulting was active in the Sumet area. It is notable that no rocks of Ngorora or younger age are exposed on the high ground of Saimo or Sidekh. In fact all rocks younger than the Kabarnet Trachyte of 7 Ma are confined to the eastern foothills of the Kamasia (Fig. 2). Furthermore the succession exposed in the Saimo fault scarp consists almost entirely of lavas with very few sediments. It seems that the Saimo and Sidekh areas have been positive regions since at least 16 Ma, generally standing above adjacent sedimentary basins and overstepped only by widespread lava formations, even this ending after the Plateau Phonolite at about 12 Ma. The major Kamasia Hills faults thus have a very similar history to the Elgeyo Fault, the major difference being the backtilting of the Kamasia blocks compared with the near horizontality of the Elgeyo block.

Tauxe *et al.* (1985) demonstrated, on the basis of detailed K–Ar and palaeomagnetic dating of the Ngorora Beds in part of the Sumet area, that sediment accumulated at a uniform rate of 88 m Ma^{-1} in the lower beds and 250 m Ma^{-1} in the upper beds even though the volcanic rocks from which they were derived were deposited sporadically in the source area. They attributed this to a uniform rate of subsidence of the depositional basin.

Bishop *et al.* (1971) presented a plot of approximate accumulated thickness against K–Ar dates for the Cenozoic succession in the Kamasia area. This shows considerable scatter but they drew a curve indicating an accumulation rate falling from 300 m Ma^{-1} at the bottom

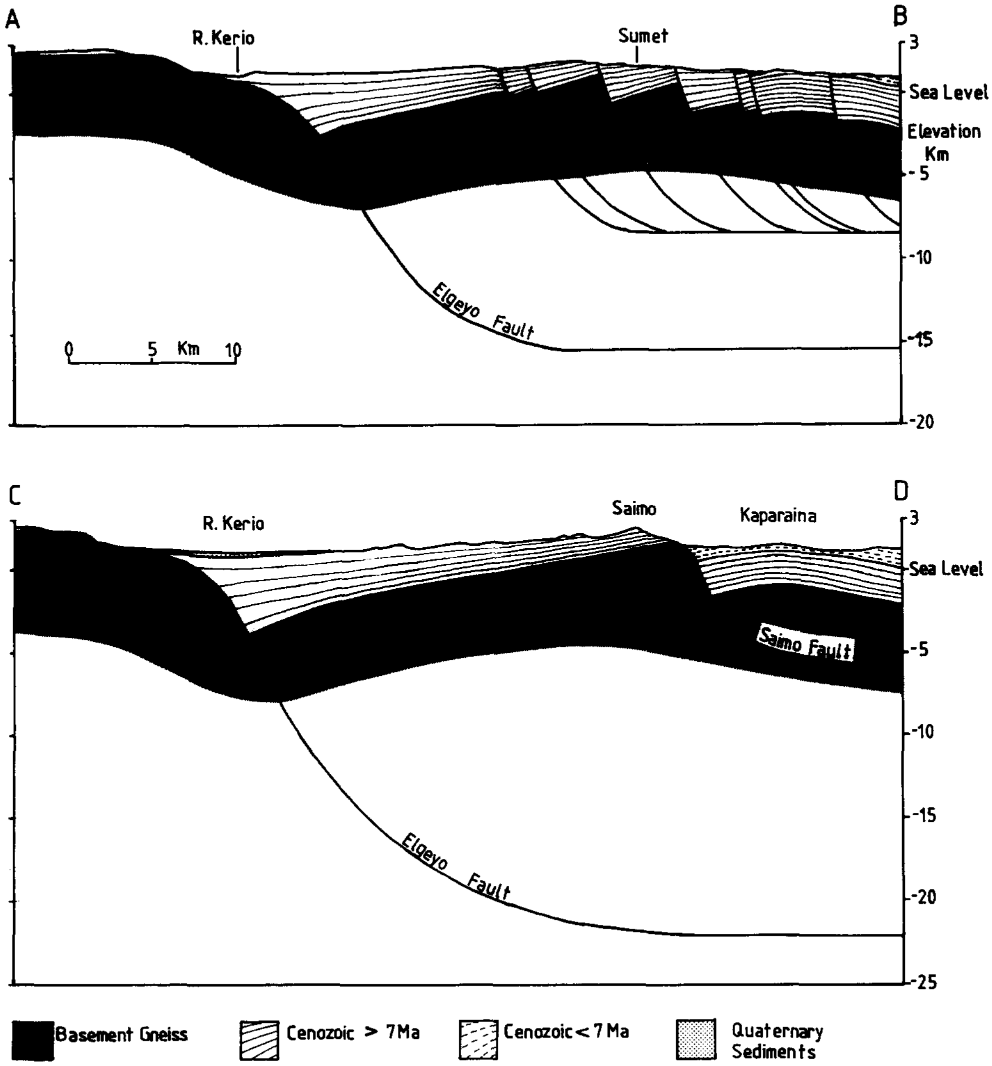


Fig. 4. East–west sections across the Elgeyo–Kamasia area. Locations shown in Fig. 2. Vertical = horizontal scale. Depth of décollement of Elgeyo and Saimo Faults in C–D as calculated in text. Depth of Elgeyo Fault in A–B shown shallower in conformity with shallower depth to basement surface. Sumet faults assumed to coalesce at an intermediate depth.

of the succession to 100 m Ma^{-1} at the top. Chapman & Brook (1978) gave a later version of this plot (Fig. 3) from which a crudely linear accumulation rate of about 150 m Ma^{-1} can be deduced. These rates for the entire Cenozoic succession are similar to the rates found by Tauxe *et al.* (1985) for the Ngorora Formation. The detailed rate of accumulation has varied in a range of up to a few hundred m per Ma and the average rate may have slowed down.

Eastern Kamasia Foothills

The principal structure in the eastern Kamasia Foothills is the Kaparaina Arch, an anticline 8–10 km wide which can be followed for 55 km between 0°30′N and 1°N (Chapman *et al.* 1978) (Fig. 2). The exposed core of the anticline consists of the 5.4–3.9 Ma old Kaparaina Basalt. The anticline reaches a topographical culmination in the Kaparaina Range, 10 km east of Saimo, where it is intruded by a NNE-trending basalt dyke swarm, implying local extension (Fig. 4, section C–D). The Kaparaina Basalt is overlain by sediments of the Chemeron Formation (4–2 Ma) which formed in unconnected basins east of the Kaparaina Range and between the range and the Saimo Escarpment. The western outcrop was tilted down to the west before being unconformably overlain by younger sediments. The Kaparaina Arch therefore shows phases of growth before and after the deposition of the Chemeron Beds.

The Arch represents the onset of rollover into the Saimo Fault and lies 6 km east of its surface trace. Since rollover into a fault begins above the point where the fault plane becomes horizontal (e.g. model cross-sections in Gibbs 1983) it follows that the Saimo Fault becomes horizontal 6 km east of its surface trace. For two simple models of the form of the fault, a plane dipping at 45°, such as characterize major faults in Greece (Jackson & McKenzie 1983), and a circular arc, the fault becomes horizontal at a depth of 6 km. This depth is shown in Fig. 4, section C–D.

Farther north, east of the Sumet area, there is a series of structures described by Chapman *et al.* (1978) as horsts anticlinally folded perpendicular to the faulting. Pickford (1978b) interprets these as radial anticlines on the surface of a block bounded by an eastward downthrowing arcuate fault that starts just north of Saimo and passes northeast to about 15 km east of Sumet and then north to meet the Kito Pass Fault (Fig. 2). These structures are probably an outlying feature of the complex faulting between the Saimo and Kito Pass Faults, the compression being due to the area lying in the re-entrant between the two major normal faults.

Synthesis

The structure of the Elgeyo–Kamasia area can be interpreted as a related set of listric growth faults. Faulting begins abruptly at the Elgeyo Fault which has been in existence since 12 Ma and probably since before 16 Ma, the only formation ever to have overstepped it being the upper part of the Uasin Gishu Phonolite of 14.5–12 Ma.

The throw on the Elgeyo Fault is at its greatest, more than 5 km (perhaps as much as 6 or 7 km), at about 0°40′N. Here the downthrown block shows rollover into the fault at both the top of the basement and the present land surface. The rollover increases with depth, accommodating a wedge of sediment while the thickness of lavas remains the same. The adjacent part of the Kamasia Hills shows the basement on Saimo rising to almost the same elevation as in the Elgeyo Escarpment. The greatest throw on the Saimo Fault thus corresponds to the greatest throw on the Elgeyo Fault. In front of Saimo, the Kaparaina Arch attests to rollover of the next downthrown block with dykes showing extension across its crest (Fig. 4, section C–D).

A different pattern is seen at about 1°N. Here there is some fault drag against the Elgeyo Fault. On the Kamasia Hills a complex pattern of relatively minor faults lies between the major Saimo and Kito Pass Faults. East of the Kamasia Hills a compressionally folded area is bounded by an arcuate fault which seems to be an outlier of this complex faulting (Fig. 4, section A–B).

The 25–30 km separation of the Elgeyo Escarpment and Kamasia Hills is similar to the 30–40 km spacing of the crests of major rotated fault blocks in Greece and the Basin and Range Province (Jackson & McKenzie 1983). The similarity in the elevation of the top of the basement on Saimo as in the Elgeyo Escarpment is also a characteristic of other major tilted fault block areas.

To summarize, the Elgeyo Fault is a major listric growth fault which has grown continuously for at least 12 and probably 16 Ma and is still active now. The rotated block of the Kamasia Hills is dominated by two major faults with an area of complex faulting between and passing into monoclines to the north and south (Chapman *et al.* 1978, fig. 5). The culmination of the Saimo Fault corresponds with the greatest throw on the Elgeyo Fault and at this latitude rollover into both faults is most marked. East–west sections at 0°40′N and 1°N, illustrating the structure proposed here are shown in Fig. 4.

THE LAKE BOGORIA AREA

The eastern side of the rift valley at 0°15′N is formed by three fault escarpments, the Ngelesha-Aruru Fault in the east, the Wasages-Marmanet Fault and the Lake Bogoria Fault nearest the centre of the rift (Griffiths 1980). The Lake Bogoria Fault has its greatest topographical expression of 800 m east of Lake Bogoria where it forms an arc through nearly 90° with the lake on the inside. The ends of the arc bend back to form straight escarpments which diminish to the north and south (Fig. 5). The total length of the fault is about 60 km. The Wasages-Marmanet Fault scarp is up to 850 m high and is sinuous so that it is only about 6 km east of the Lake Bogoria Fault scarp at Kaon but moves off to 25 km away farther south.

The Lake Bogoria and Wasages-Marmanet fault scarps both expose the Samburu Basalt (21–14 Ma, Chapman & Brook 1978) overlain by the Rumuruti Phonolite (12–10 Ma). The Rumuruti Phonolite shows an eastward dip of generally less than 5°. The Samburu Basalt underlies it with only a slight unconformity at Kaon but farther west the unconformity is quite marked, the basalt dipping at up to 30° (McCall 1967). This may be explained by rotation of the block west of the

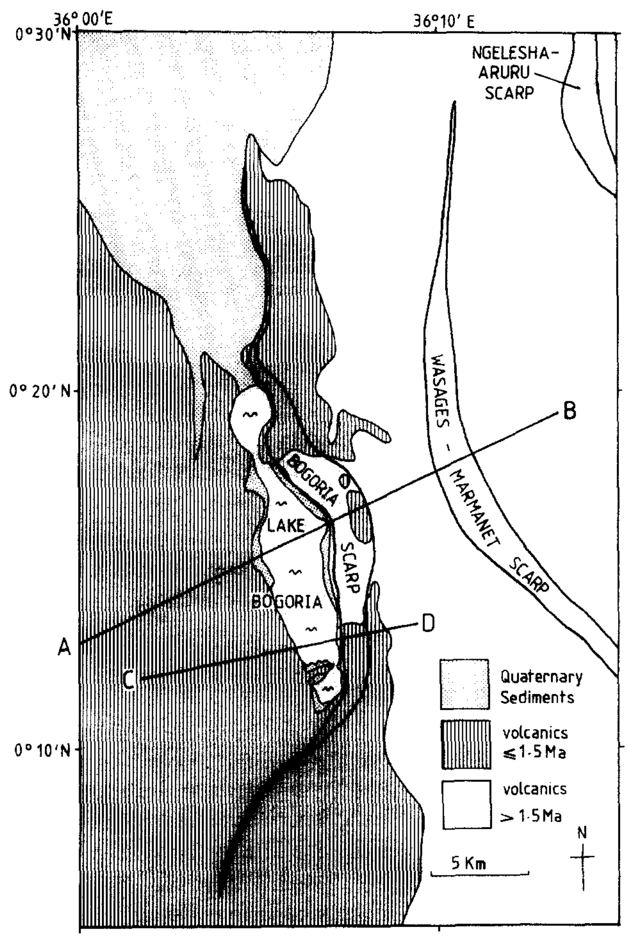


Fig. 5. Geological sketch map of Lake Bogoria showing major fault escarpments. A–B: line of section in Fig. 6. C–D: line of section in King (1978, fig. 3:9).

Wasages-Marmanet Fault after the emplacement of the Samburu Basalt and before the eruption of the Rumuruti Phonolite, the area east of the fault remaining relatively undisturbed.

The junction between the Samburu Basalt and Rumuruti Phonolite is very irregular and there are isolated patches of phonolite on fault escarpments exposing basalt (see map in McCall 1967). This can be accounted for by the phonolite burying a landscape of fault blocks of basalt. Rejuvenation of the faulting then produced mantled fault escarpments.

An angular unconformity separates the Rumuruti Phonolite from the overlying Tasokwan Trachyte (7.6–6.6 Ma). This younger formation occurs on the dip slopes behind the Lake Bogoria and Wasages-Marmanet fault scarps and field relationships show that the fault scarps existed in their present positions during its emplacement (Griffiths 1980). There was therefore further rotation of the fault blocks between the eruptions of the Rumuruti Phonolite and Tasokwan Trachyte. This continual rotation of a block on one side of a fault without rotation of the other side is only possible if the fault plane is curved (Wernicke & Burchfiel 1982).

The central part of the Bogoria Escarpment east of the lake has a ledge of Hannington Trachyphonolite which also forms the floor of the rift valley west of the lake. The elevation of the top of the trachyphonolite outcrop above the water surface varies being at its greatest, about 800 m, by the centre of the lake. This is also where the lake is widest and hence presumably where the easterly dipping trachyphonolite surface reaches the fault plane at its deepest point. It is therefore the point where the throw of the fault is greatest. At the northern and southern ends of the lake the trachyphonolite is exposed over the whole height of the the escarpment, probably mantling the Samburu Basalt and Rumuruti Phonolite behind.

There is thus evidence that the Lake Bogoria and Wasages-Marmanet Faults have been in existence since at least 12 Ma and have been active several times since then. Maguire and Long (1976) recorded a set of earthquake epicentres about 20 km wide and 100 km long stretching from the Equator to just north of Lake Baringo and including Lakes Baringo and Bogoria. The width of the belt is similar to the estimated accuracy of epicentre location, so the belt may really be narrower than 20 km. These earthquakes probably record modern movement on the Bogoria Fault, as is also suggested by the concentration of fumaroles in Lake Bogoria. Continued activity on the Wasages-Marmanet Fault is shown by a magnitude 7 earthquake in 1928 which produced a fissure 30 km long and up to 2.5 m wide along the foot of the escarpment (McCall 1967).

The Rift Valley floor west of Lake Bogoria is composed of Pleistocene lavas, chiefly the Hannington Trachyphonolite. The surface of this lava pile is approximately horizontal over most of the rift floor but the easternmost 10 km dips increasingly steeply towards the Bogoria Escarpment. The rift floor is broken up into a horst and graben structure by minor faults spaced at intervals of a few hundred metres. This small-scale faulting fades away within a few km of the main Bogoria fault. On the upthrown side similar small-scale faulting shows a quite different orientation, NNE–SSW rather than N–S as well illustrated by King (1978, fig. 3:9). These observations suggest that the Lake Bogoria Fault is concave upward at depth. Normal faulting on this fault would require rollover of the downthrown surface towards the fault plane and hence tensional faulting. Shallowing of the principal fault plane close to its surface expression would restrict the throw of the minor faults in this area. The stress fields above and below the major fault plane would be disconnected from each other, allowing the development of differently oriented fault patterns on either side of the principal fault trace.

An estimate of the profile of the Bogoria Fault may be made by using the surface of the Hannington Trachyphonolite as a reference plane for area balancing. The cross-sectional area for this surface on the profile A–B in Figs 5 and 6 is 5.8 km^2. The equivalent area on the profile in King (1978, fig. 3:9) (shown as C–D in Fig. 5) is 2.4 km^2. However there are problems in measuring the heave of the fault.

The edge of the Hannington Trachyphonolite on the upthrown side has been eroded back while on the downthrown side it is hidden beneath the scree and the lake. McCall (1967) and Griffiths (1980) both show the fault trace where the Samburu Basalt appears at the top of the scree slope. This is about 1 km from the edge of the Hannington Trachyphonolite facing and so suggests a heave of about 1 km. The erosional recession of the edge of the facing will make this an overestimate but on the other hand the downthrown edge of the Trachyphonolite must be to the west of the surface trace of the fault plane unless the fault plane is vertical. These two correcting factors counteract each other but the recession of the fault scarp is probably more important. The heave is therefore less than 1 km. The downthrow of the Hannington Trachyphonolite surface on the profile A–B and on King's (1978) profile is about 1000 and 700 m, respectively. Using the same dip on the fault surface as for the Elgeyo Fault (Swain *et al.*, 1981) the heaves would be 500 and 350 m. With these figures the depth to décollement on profiles A–B and C–D (Fig. 6) would be 12 and 7 km, respectively.

The different depths along the two profiles are quite plausible and imply that the fault plane is spoon shaped, concave upwards both perpendicular to the general trend of the surface trace and parallel with it. This is compatible with the 90° arc formed by the central highest part of the fault scarp. The profile A–B cuts the escarpment at the point where the throw is greatest while along the profile C–D the fault plane is shallowing towards the southern arm of the fault escarpment.

The dip of the trachyphonolite surface just before passing under the lake is 5° on profile A–B and 7° on C–D. Extrapolating these dips to the fault surface and using throws of 1000 m and 700 m, the depths at which a circular arcuate fault would become horizontal are 11 and 6 km, respectively. The similarity between these

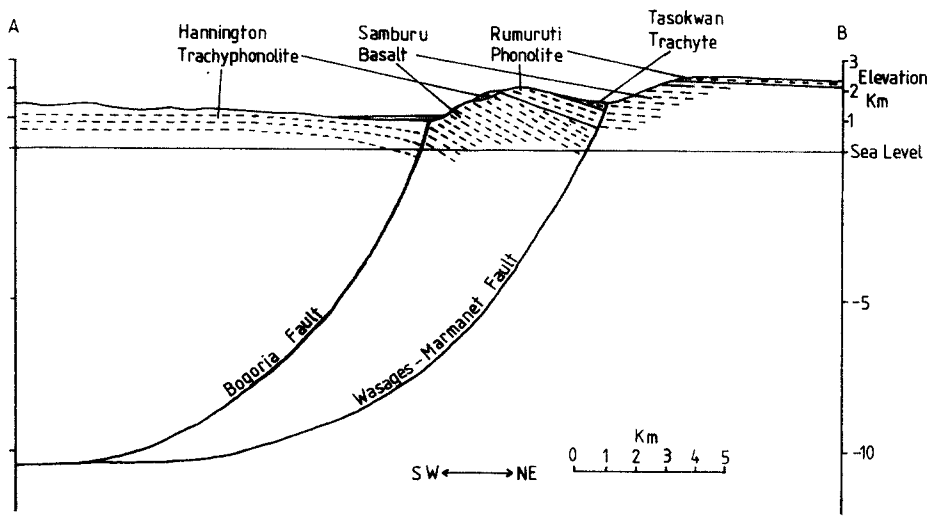


Fig. 6. Section across the Bogoria and Wasages-Marmanet Faults. Location of section shown in Fig. 5. Vertical = horizontal scale. The increasing dip with age of the formation between the two faults shows a progressive rotation of this block. The Hannington Trachyphonolite dips increasingly towards the Bogoria Fault and forms a prominent facing on the escarpment. Depth to décollement of the Bogoria Fault is explained in text. Wasages-Marmanet Fault is assumed to join Bogoria Fault at depth.

figures and those derived by the area balancing method suggests that they may approximate the true depth to décollement.

A possible cause of rollover into a major fault is antithetic faulting. In the case of the Bogoria Fault this is clearly not the case. The small scale faulting on the downthrown side throws both ways in the centre of the rift but becomes uniformly synthetic close to the major fault scarp. It therefore tends to reduce the rollover.

The history of the Wasages-Marmanet Fault overlaps with that of the Bogoria Fault. This means that their fault planes cannot intersect (Jackson & McKenzie 1983). The Wasages-Marmanet Fault must either join the Bogoria Fault at depth or pass beneath it. Since the surface traces are so close the former possibility is adopted in Fig. 6.

Griffiths (1980) has described box faulting between the Lake Bogoria and Wasages-Marmanet Faults and between the latter and the Ngelesha-Aruru Fault (Fig. 5). These are both complex areas about 4 km square in which short faults intersect each other at nearly 90° angles. North from each box fault system a set of ramps, downfaulted on their western side and dipping gently northwards, descend to the rift floor. The overall effect is to relay the topographic margin of the rift valley in a more northeasterly direction than the trend of the faults themselves. Griffiths noted the similarity of this box faulting to the fault complex at Sumet in the Kamasia Hills. It seems that small areas of complicated faulting are characteristics of the offsets between major faults. They may result from the overlap of the differently oriented stress fields associated with the growth of adjacent major faults.

The evidence presented here strongly suggests that the major faults on the eastern side of the Kenya Rift Valley at the latitude of Lake Bogoria are listric growth faults. The important points are several episodes of movement on the Lake Bogoria and Wasages-Marmanet Faults, continued rotation of the block between the two faults, small-scale faulting on and rollover of the surface of the downthrown block towards the Bogoria Escarpment, and signs of present day activity on the faults. An E–W cross-section illustrating the structure favoured here is given in Fig. 6.

OTHER AREAS

The Lake Manyara area

Rykounov *et al.* (1972) recorded microearthquakes in the Lake Manyara area of northern Tanzania (Fig. 7). Their data are recorded in the form of a map of the epicentres and an E–W section showing the location of the foci, projected onto the plane of the section, reproduced here in a modified form as Fig. 7(b). Unfortunately the data points are not correlated between the two diagrams and are not clearly resolved within them. Rykounov *et al.* noted that there was little correlation of the foci with surface geological features. However, their data do show a deepening of the foci eastwards. Lake Manyara is bordered on the western side by a fault scarp which reaches its greatest height of 800 m just south of the lake. The data show a concentration of foci east of the escarpment at this point. On the vertical section the foci seem to cluster about two lines dipping eastwards and flattening at about 15 and 40 km, respectively (Fig. 7b). The western end of the shallower line lies beneath the surface trace of the Manyara Fault while the western end of the deeper line can be projected to reach the surface at the 600 m high fault escarpment on the northwest side of Lake Eyasi.

Macintyre *et al.* (1974) dated lavas between Lakes Manyara and Natron and hence identified a major phase of faulting on the Manyara-Natron Fault at 1.15–1.2 Ma. They also noted that there was geomorphological evi-

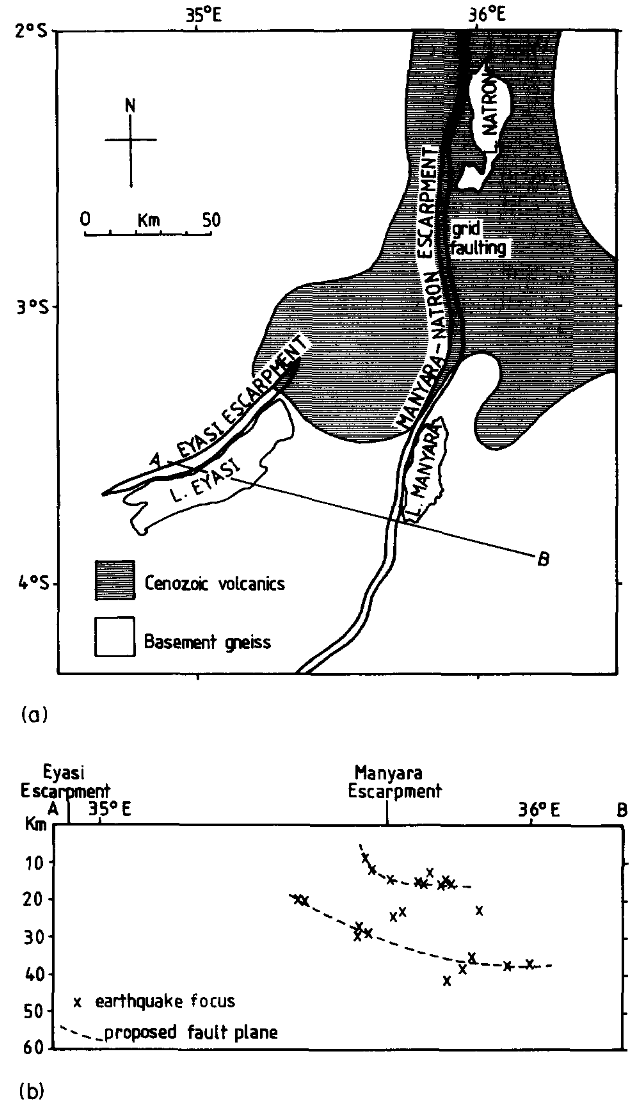


Fig. 7 (a) Sketch map of the Manyara area showing major E- and SE-facing fault escarpments. A–B: line of section in (b). (b) Section across Manyara area showing projected locations of microearthquake foci, after Rykounov *et al.* (1972). Location of section shown in (a). Vertical = horizontal scale. The foci may indicate two fault planes which intersect the surface at the Manyara and Eyasi Escarpments.

dence for an older phase of faulting and that there was a pattern of small-scale grid faulting on the downthrown side.

This area of northern Tanzania thus shows good evidence for listric growth faulting. The Manyara-Natron Fault resembles the Lake Bogoria Fault in showing at least two phases of movement and grid faulting on the downthrown side. The presence of lakes and hence closed depressions at Natron, Manyara and Eyasi on the downthrown sides of faults suggests rollover into the fault planes. Of most significance is the direct evidence from microearthquakes for listric fault planes which can be correlated with fault escarpments at the surface.

The Kirikiti-Lengitoto Fault

The Kirikiti-Lengitoto Fault runs approximately N–S along the western side of the rift between 1°30′S and 2°S. It consists of a succession of segments 10–20 km long trending alternately NNW and NNE. The Endosapia Gorge cuts through the facing of lavas into the fault escarpment just north of one of the bends and exposes the fault plane in the basement rocks (Fig. 8) (Crossley 1979). The fault scarp is exposed over a height of 150 m although its throw is 600 m. The fault plane strikes northeast and dips to the southeast at 65°. There is a 5 m thick shear zone with slickenside lineation plunging at 45° to the northeast (Fig. 8). These slickensides show that movement on this part of the fault plane was not completely in a dip-slip direction but had a strike-slip component. The direction of movement of the downthrown block would have been along an azimuth of about N 70°E, more nearly eastwards than the northeasterly strike of the fault plane at this locality. This is to be expected of a fault with a gross N–S trend but composed of segments of varying orientations. For such a fault a horizontal component of movement can only be in an E–W direction if the fault plane is not to open along one

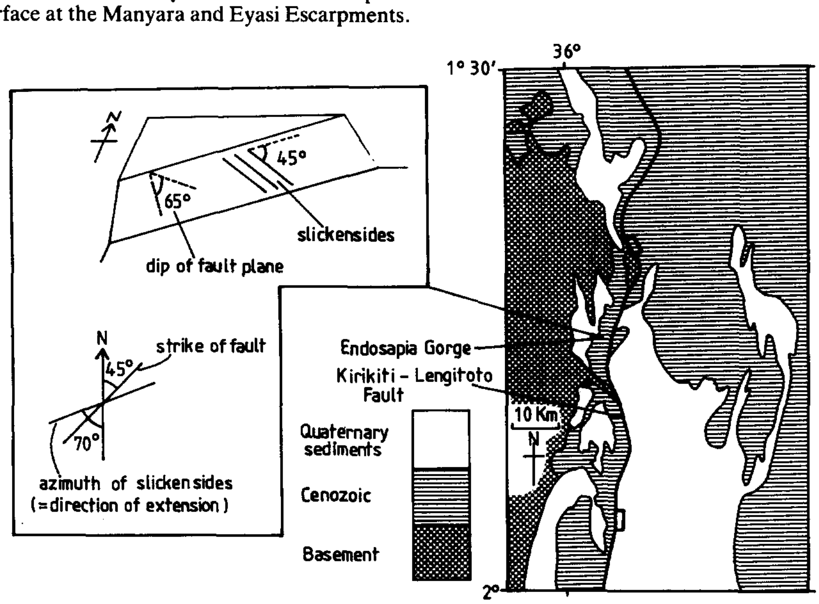
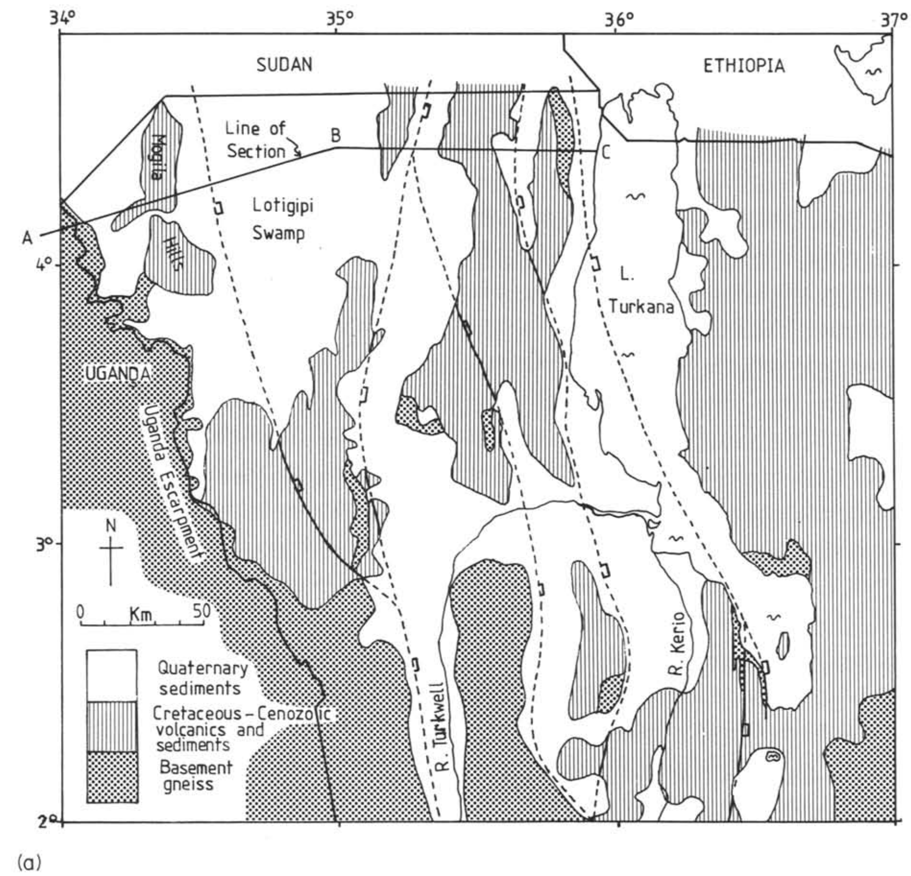


Fig. 8. Map of Kirikiti-Lengitoto Fault area after Crossley (1979). Diagram shows features of a fault plane exposed in the Endosapia Gorge, as described by Crossley. Slickensides imply movement in a nearly eastward direction.

set of segments. A slight opening along the fault plane is in fact shown by Crossley's observation of a dyke intruded along it.

West of Lake Turkana

The area between Lake Turkana and the Kenya/Uganda border consists of a series of N–S trending ridges separated by low lying areas floored by Pleistocene alluvial sediments (Fig. 9). These ridges expose basement rocks overlain by Cenozoic lavas and sediments (Walsh & Dodson 1969). North of 3°N the first three ridges west of the lake show faulted eastern scarp slopes and gentle westerly-dip slopes in which the Cenozoic rocks show dips of 12–17°. These ridges are then separated by the Lotigipi Swamp from the Mogila Hills in which Cenozoic rocks are exposed dipping at similar angles to the east. West of the Mogila Hills lies the Uganda Escarpment which has been regarded both as an eroded fault or fault line scarp and as an eroded monocline.



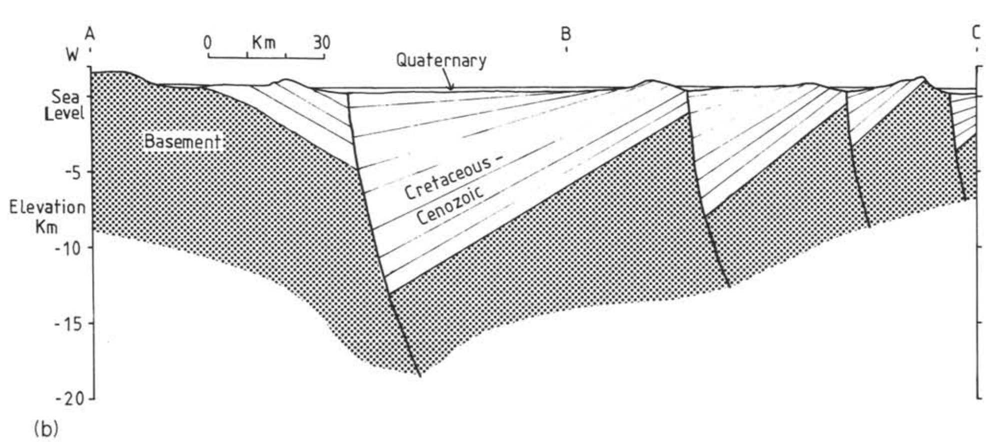


Fig. 9. (a) Geological sketch map of the area west of Lake Turkana, based on Walsh & Dodson (1969) and Rhemtulla (1970). The major faults are shown, dashed where buried under Quaternary cover. A–B–C: line of section in (b). (b) East–west section across area west of Lake Turkana. Location shown on (a). Vertical exaggeration = 4. This area consists of a series of westward tilted half graben infilled with Cretaceous and Cenozoic sediments and volcanics. The crest of the westernmost fault escarpment has retreated by about 50 km and is now the Uganda Escarpment. Eastward-dipping beds in the Mogila Hills overlie eroded basement west of the trace of this fault at depth.

The cover succession begins with a series of generally coarse-grained, often conglomeratic, sediments, the Turkana Grits, which contain both early Miocene and late Cretaceous fauna (Arambourg & Wolff 1969). These sediments must have been derived from some nearby readily eroded source. This may have been a major fault scarp. It is proposed here that the Uganda fault scarp was initiated during the late Cretaceous. Since then it has been eroded back, sourcing the Turkana Grit. The original location of the fault trace was under the Lotigipi Swamp 30–50 km east of its present location. The E-dipping Cenozoic rocks of the Mogila Hills lie on the eroded surface of the basement rocks west of the original fault scarp. The range of hills east of the Lotigipi Swamp represent tilted fault blocks in front of the main rift bounding fault, an analogous situation to the Kamasia Hills (Fig. 9b). Cerling & Powers (1977) noted that faulting and the principal depositional centre in this area have migrated eastward with time while Rhemtulla (1970) recorded that faults near the southern end of Lake Turkana showed multiple periods of movement.

The Mzima Fault

The Mzima Fault, between Nairobi and Mombasa, was the first faced scarp structure to be recognized in Kenya (Temperley 1966). Here a basalt lava flowed against a partly eroded fault scarp in Pleistocene lake beds. Rejuvenation of the fault produced a facing of basalt about 2 km long and 7 m high. Temperly noted that the lava flowed over part of the outcrop of the lake beds after they had been 'tilted and faulted' and presented a series of sketch sections, reproduced here as Fig. 10, to show the evolution of the structure. These clearly show the dip of the lake beds increasing towards the fault. This fault, the earliest described faced scarp structure in Kenya, seems also to be a listric growth fault showing distinct phases of movement and rollover of the downthrown side into the fault plane.

CONCLUSIONS

This paper has presented evidence that many of the major and even some of the minor faults in the Kenya Rift Valley are listric growth faults. It may be true in general for all the major faults in the Kenya Rift.

In the past, discussion of the history of the major faults has been in terms of isolated periods of movement separated by periods of volcanic eruption or sedimentary deposition. However detailed mapping, particularly in faulted sedimentary sequences, has tended to show that the faulting actually takes place in a succession of minor pulses. Accumulation of volcanic and sedimentary rocks frequently keeps pace with the faulting so as to minimize the elevation of the fault scarp. It seems that the major faults have lifespans of the order 10^6–10^7 years with growth taking place at a grossly uniform rate which may gradually slow down.

The elevation of the basement surface along the Uganda and Elgeyo Escarpments resemble fault line scarps but this hypothesis suffers from the apparent absence of fault traces at their feet. Some authors have proposed that the escarpments are eroded monoclines instead. The resolution of the problem suggested in this paper is that the fault trace lies far in front of the crest of the escarpment. For the Uganda Escarpment it lies under the Lotigipi Swamp and for the Elgeyo Escarpment it is under the Kerio Valley. The Uganda Escarpment has receded about 50 km since the late Cretaceous (Fig. 9), the Elgeyo Escarpment about 10 km since the eruption of the Uasin Gishu Phonolite of 12 Ma (Fig. 4) and the Bogoria Escarpment about 1.5 km since at least part of the Hannington Trachyphonolite of 1.5–0.3 Ma (Fig. 6). It seems that fault escarpments are eroded backwards at a rate of the order of 1 km Ma^{-1},

A common feature of faulting in the Kenya Rift is the progressive transfer of the zone of active faulting across a belt of faults from the oldest, marked by a mature erosional escarpment, to the youngest which defines the present location of the rift valley. This is demonstrated in this paper for the Elgeyo–Kamasia area and is described by Baker & Mitchell (1976) for the Kedong–Olorgersaile area and by Cerling & Powers (1977) for the Lake Turkana area. The depth to décollement estimates in this paper show that a significant feature of these fault systems is that the older, outer faults sole out at greater depth than the younger faults closer to the axis of presently active rifting.

Calculation of the depth to décollement for a listric fault depends crucially on knowing the heave of the fault (Davison 1986). In the Kenya Rift this is extremely difficult because the foot of the fault escarpment is commonly buried by younger rocks and the crest is eroded back from its original position. This makes estimates of the heave of the fault very conjectural and hence reconstruction of the form of the fault only very approximate.

The interpretation of the subsurface structure of the rift would be greatly facilitated if there were seismic reflection data available in this area but unfortunately this is not the case. However, Rosendahl *et al.* (1986) have published the results of a seismic reflection survey in Lake Tanganyika which shows the Tanganyika Rift to consist of a succession of crescent shaped half grabens,

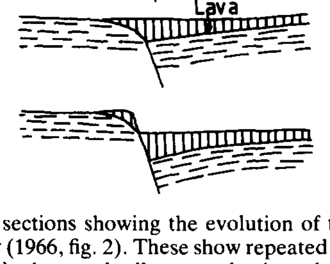


Fig. 10. Sketch sections showing the evolution of the Mzima Fault, from Temperley (1966, fig. 2). These show repeated movement on the same fault plane and rollover on the downthrown side.

formed by listric faults of the order of 100 km long. Major listric faulting is thus well established in another part of the East African Rift System.

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