Is structure the main control of river drainage and sedimentation in rifts?

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Abstract - Changes in the topography of rifts as they evolve toward aulacogens and passive margins control river drainage and, therefore, sedimentation. The African continent offers an opportunity to assess the effect of rifting on river drainage and sedimentation at several stages of development. The East African system is young and sediment-starved as a result of both the regional doming that accompanies early phases of rifting and the back-tilting of footwall fault blocks. Each of these factors diverts rivers away from subsiding basins. The longitudinaliy segmented nature of an early rift discourages axial drainage and major river inputs originate on the hanging-wall platform or roll-over of half graben sub-basins. During later stages of development, regional subsidence may enlarge the drainage basin, while sedimentary infilling of the trough may facilitate axial drainage. A good example of this occurs in the West African system where the Benue River flows for 1200km along a rift trough and the drainage basin incorporates the catchment of the River Niger. Sedimentary sequences accumulating in rift basins are characterised by strong cyclicity of depositional style. This can be taken to reflect the sporadic nature of the pulses associated with faulting. However, it can also arise from the changes in base-level that are brought about by fluctuations in climate in closed basins and by eustatic adjustments in sea-level where marine inundation has occurred. Taking examples from the East African system and from the Gulf of Suez, changes in sedimentation that are consequent upon marine or lacustrine transgression and regression are shown to be almost indistinguishable from those triggered by boundary fault activity. Cyclical patterns of sedimentation can only be attributed to tectonic activity if corroborative evidence is available.

INTRODUCTION

Over the past 5 years, ideas concerning the pattern of sedimentation that occurs in rifts have changed considerably. Following the work of Bridge and Leeder (1979), models of basin fill have reflected recent advances in the general understanding of rift structure (e.g. Frostick and Reid 1987a and 1987b; Leeder and Gawthorpe 1987). These advances have occurred because of the significant expansion of geophysical exploration that followed the oil crises of the late 1960's and 1970's.

Of particular importance has been the work carried out in the Tertiary rifts of East Africa (Rosendahl et al. 1986; Khan et al. 1986; Rosendahl 1987). However, African rifts are not confined to the east of the continent, nor to the Tertiary period. There have been sporadic bouts of crustal extension since Pre-Cambrian times (Grove 1986). Moreover, while younger rifts have occasionally re-used old lines of weakness (e.g. the Rukwa rift in Tanzania), they have often established new fault trends more in keeping with prevailing stress fields, as in much of the Gregory Rift in East Africa. As a result, Africa offers for study a unique range of rift development, from the early stages of continental rifting to the evolution of passive margins.

The pattern of river drainage associated with rifts can be expected to change as structural evolution takes place and as erosion alters the landform. Because rivers are major suppliers of sediment, it might also be expected that sympathetic changes would occur in sedimentation. Structural history should exert a strong control on the character of the basin fill. This paper examines the sedimentary consequences of rift development by drawing out the similarities and contrasts in drainage patterns and sediments of three major systems. The Tertiary-Quaternary system of East Africa is young and still active. In contrast, the older Cretaceous rift of West Africa is tectonically quiescent, though still having topographical expression. Both are presently continental in character. On the other hand, the Red Sea has a long history of marine influence, and is evolving into a passive margin.

GEOLOGICAL BACKGROUND

The geological history of the African Rift System is well documented (e.g. Baker 1970; Baker et al. 1972; Fairhead 1986) and only a brief review is included here.

The development of the East and West African and the Red Sea Rift systems spans two major periods of continental fragmentation. The first period was associated with the opening of the
South Atlantic during the late Cretaceous. Two arms of a trite rift system evolved to form the Bight of Benin. The remaining arm - the Benue Trough - failed to develop in the same way. As a result, tectonic activity died out and the flanking fault scarps have been softened by erosion.

The East African Rift System contrasts with that of West Africa in having a long history of volcanism (Baker 1986; Williams and Chapman 1986). This gives, among other things, a good chronostratigraphical framework for the interpretation of basin fills (Baker et al. 1971; Fitch et al. 1978; Chapman and Brook 1978). Fairhead (1986) suggests that the contrast reflects differences in the degree and rate of extension rather than any fundamental differences in mantle structure. Estimates of the age of inception of the East African Rift System range from late Cenomanian to early Miocene (Arambourg 1933; Baker et al. 1972; King 1978; Williamson and Savage 1986). It is associated with a phase of continent fragmentation which was responsible for development of the Red Sea/Gulf of Aden proto-ocean. The Red Sea/Gulf of Aden and East African Rifts meet in the Afar "triangle" of Ethiopia. Like its West African counterpart, the East African Rift represents the failed arm of a trite system, although at a much earlier stage of development.

The Red Sea/Gulf of Aden proto-ocean is opening at an estimated rate of 10 mm a$^{-1}$ (Courtillot et al. 1987) as a consequence of the movement of the Arabian plate northeastwards away from Africa. The Gulf of Suez is a northern extension of this system, with a history of faulting dating back to the Oligocene (Sellwood and Netherwood 1984). The earliest phases of movement were probably associated more with crustal relaxation following the Alpine orogeny than with incipient continental break-up. However, later developments of Plio-Pleistocene age reflect the relative motions of the Arabian and African plates. Recent small scale activity is confined to offshore locations close to the axis of the Gulf (Garfunkel and Bartov 1977). Most of the differential plate movement is currently being accommodated by strike-slip movement on the Gulf of Aqaba/Dead Sea fault (Courtillot et al. 1987). In consequence, the Gulf of Suez is also a rift that has failed to develop fully. However, it has a history of marine inundation which dates back almost to its inception.

THE INFLUENCE OF RIFTING ON REGIONAL PATTERNS OF DRAINAGE

A major geomorphological and sedimentological consequence of continental rifting is the diversion and realignment of river drainage, whatever the pre-existing pattern. The character of the re-ordered drainage will control, in turn, the pattern and nature of clastic sediment supply to the developing rift basins. It therefore becomes a chief determinant of the architecture of the resulting basin fill. Where major river systems are captured by the rift, thick accumulations of alluvial sediments may develop. In contrast, areas of the rift receiving no fluvial inputs may become zones in which organic-rich, carbonate and diatomaceous deposits occur (Cohen and Thouin 1987; Cohen in press).

One widely held misconception is that rifts are sumps that always attract drainage from a wide area. The development of facies models in which extensive alluvial fans emanate from the rift footwalls reflects this idea (e.g. Hubert et al. 1976). However, the topographical expression of rift structure, especially during the early phases of development, may not encourage diversion of rivers into the rift. This is the case for all rifts located over mantle plumes, where regional doming is one of the first expressions of incipient continental break-up. As a result, sediment supply to young rift basins can be highly localised and small-scale (Frostick and Reid 1987 a,b; Leeder and Gawthorpe 1987). In this type of rift it is only as the system ages that there is a tendency for the integration of drainage over a wide area. At this stage clastic sediment inputs can become more evenly distributed about the basin, but by this time it may be topographically indistinct, as in the case of the Abu Gabra system of southern Sudan.

The trends in drainage development that occur as rifts evolve are evident in the river systems of Africa. For this reason, the regional patterns of drainage contributing to the West and East African systems and to the Red Sea have been extracted from published maps. The areas involved are located in Fig. 1; the drainage nets and major faults are shown in Figs. 2, 3 and 4. In addition, a larger scale analysis of the rivers and structure of the 145 000 km² Suez drainage is shown in Fig. 5.

Immediately obvious are the very limited catchment areas of the East African, the Red Sea/Gulf of Aden and the Suez rifts (Figs. 2, 4 and 5). All can be considered as young systems. Rivers draining to each of these rifts have headwaters less than 350 km from the basin axes along their entire length - over 9000km in total. Much of the local drainage flows away from, rather than towards, the basins. This is especially evident in the Gulf of Suez (Fig. 5). Here, much of the drainage close to its western side flows towards the Nile, while very little of the Sinai contributes on its eastern side, most rivers flowing northeastswards towards the Wadi el Arish - an ephemeral river that drains to the Mediterranean.

So great is the trend away from rift basins during early rifting that Lake Victoria, caught between 2
branches of the East African system (Fig. 2) receives the waters of 8 major rivers, all backshed from the rift shoulders. Indeed, because the topographic gradient is often away from the rift (Frostick and Reid 1987a), several of the present rift lakes, notably Tanganyika and Malawi, are starved of clastic sediment. As a result, sedimentation rates are low (0.5 mm a⁻¹ compared with almost 7 mm a⁻¹ for L. Turkana) and both biogenic and carbonate facies are particularly well developed (Cohen and Thouin 1987).

The situation in West Africa is very different (Fig. 3). Here the rift is a major drainage corridor reaching towards the Atlantic Ocean. The Benue River runs for most of its course along the basin axis and is joined by the immense Niger system which runs its last 480 km within the Benue trough. A thickness of > 5 km of predominantly clastic sediment has accumulated since the Cretaceous (Ojo and Ajakaiye 1976). Pleistocene rates of sedimentation have been particularly high, producing substantial growth of the Niger Delta (Allen 1965, 1970). This is thought to have resulted from the southward impingement of Saharan dunes, and the transport of this easily eroded material from the Inland Delta region of the Niger and down the Benue which was, at that time, swollen by overspill from Megachad (Grove 1986).

**STRUCTURAL CONTROL OF DRAINAGE DIVERSION**

**Early Rifting Phase**

Diversion of river drainage is probably achieved very early in the history of rift basin development. In fact, the early phase of rifting in East Africa and other, similar, areas situated over mantle plumes, involves regional doming (McKenzie 1978; Rosendahl 1987) so that the river courses on peneplaned cratons such as that of Africa are likely to be threatened before substantial rift faulting occurs. As doming increases, even entrenched rivers
Fig. 2. River drainage and structure of the East African Rift. M - Malagarasi River; 2 and 3 - the locations of the Koobi Fora Formation and Kaphurin Formation sequences, respectively, given in Fig. 7.

are going to be affected. Once the rift basins develop, the topographic gradient will exclude rivers from entering the rift over a considerable portion of its length (LeFournier et al. 1985; Frostick and Reid 1987a,b).

It is increasingly acknowledged that the fundamental structural unit of rifts is a half graben controlled by movement on a single, major, listric boundary fault (Gibbs 1984 Rosendahl et al. 1986). Footwall collapse and the backtilting of fault blocks, reinforces the tendency already established by doming to shed rivers away from the rift basin
(Frostick and Reid 1987a). Few rivers and little sediment enter the rift on its faulted margins (see Table 1). However, on the hanging-wall a gradient is established towards the basin, and rivers which might have previously drained across an uninterrupted surface are captured by the developing rift basin. For example, the invertebrate fauna of the Malagarasi River (Fig. 2, letter M), which now drains to the eastern shore of Lake Tanganyika, indicates that it was at one time a tributary of the Congo River (A. Cohen, personal communication). The hanging-wall platform or roll-over therefore becomes a major route for rift rivers, sometimes responsible for over 70% of the total basin drainage area (Table 1).

Of great consequence for the pattern of sediment supply to rift basins is the discontinuous nature of the boundary faults. Half graben segments are rarely more than 150 km long and the major boundary fault shifts from one side of the rift axis...
Fig. 4. River drainage and structure of the Red Sea/Gulf of Aden proto-ocean. 1 - location of the Hadar Formation sequence shown in Fig. 7.
Fig. 5. River drainage and structure of the Gulf of Suez Rift. 1, 2, 3 and 4 - locations of the Abu Zenima, Wadi Gharandal, Wadi Dara and Wadi Kharim sedimentary sequences depicted in Figs. 8 and 9:

- Qa Plain: G - Wadi Ghuweibba; A - Wadi Araba.
TABLE 1. Rift basin catchments and the source of rivers.

<table>
<thead>
<tr>
<th>Lake Basin</th>
<th>Total Catchment Area, km²</th>
<th>% Axial Drainage</th>
<th>% Drainage Across Boundary Faults</th>
<th>% Drainage Across Hanging Wall/Roll-over</th>
</tr>
</thead>
<tbody>
<tr>
<td>Malawi</td>
<td>102,599</td>
<td>5.6</td>
<td>25.6</td>
<td>68.8</td>
</tr>
<tr>
<td>Tanganyika</td>
<td>151,685</td>
<td>9.1</td>
<td>19.9</td>
<td>1.0</td>
</tr>
<tr>
<td>Turkana (Present-day)</td>
<td>148,403</td>
<td>57.5*</td>
<td>9.9</td>
<td>32.6</td>
</tr>
<tr>
<td>Turkana (9000 a BP)</td>
<td>201,995</td>
<td>42.2*</td>
<td>7.3</td>
<td>50.5</td>
</tr>
</tbody>
</table>

* the major part of this is the Omo river, a stream that flows for more than 90% of its course outside of the rift.

to the other in an apparently orderly manner (Rosendahl et al. 1986). Between the segments there are topographical highs (the transfer zones of Gibbs 1984; accommodation zones of Rosendahl et al. 1986) which ensure separation of the rift into a series of sub-basins. Each sub-basin acts as a depocentre, even if the rift is flooded by marine incursion or by a lake (Cohen in press; Specht this volume).

At these early stages of rifting, integration of drainage along the length of the rift is inhibited by these structural and topographic highs. For example, Lake Bogoria (990 m asl) is separated from the Lake Baringo segment of the Gregory Rift by the Loboi Plain which lies only 9 m higher (Vincens et al. 1986). Because of this, true axial drainage is rare, and it often contributes little to the total catchment (see e.g. L. Tanganyika and L. Malawi, Table 1). Rivers captured by the developing basins and flowing along the axis of the system often encounter a lake after comparatively short distance. This is the situation at the northern end of Lake Turkana, where the Omo River apparently acts as an axial system. But it only spends its last 30 km - < 4% of its total length - within the rift (Butzer 1971). For the remaining 800 km of its length it drains the Ethiopian highlands to the west of the rift, and is effectively backshed away from it. In fact, the Omo is a good example of one rift segment capturing a river diverted away from another, adjacent, one. This happens frequently and results in the development of large fans or fan deltas at transfer zones.

**Antithetic and Synthetic Faulting and Drainage**

Small scale antithetic and synthetic faulting on the hanging wall block can produce significant local diversions of drainage. This is evident, for example, in the Gulf of Suez (Fig. 5, letter Q). Here, the river draining the Qa Plain is already within 12 km of the Sinai coast when it is caught behind a small antithetic fault and, as a result, runs parallel to the coast for over 40 km before entering the Gulf. Similar diversions of drainage have been inferred for ancient river systems draining across the hinged margin of a Triassic half graben (Frostick et al. 1985). An appreciation of the possible influence of these smaller structural elements on sedimentation has helped to provide a plausible explanation for an apparent conflict between different indicators of palaeocurrent direction in a situation where available outcrops are restricted in area.

**Effect of Pre-existing Structures on Drainage**

Cross-cutting pre-rift structures can exert a strong influence on the development of streams entering a rift basin, and therefore on the localities of clastic sediment accumulation. The topographical lows sometimes exploited by rivers that are attracted to a rift often correspond to faults or folds which pre-date the current phase of tectonism. This is the case on the northwestern shore of the Gulf of Suez, where a set of east-west trending faults probably generated during the emplacement of the Syrian Arcs (Eocene; Chenet et al. 1987) are exploited by the Wadi Araba and Wadi Ghuweibba...
the rift axis. This provides the opportunity for development of long axial rivers such as the Benue (Fig. 3). The sedimentary models of Bridge and Leeder (1979) and Leeder and Alexander (1987) are most appropriate for this stage of rift development. These axial rivers may be responsible for large deltaic wedges of material where a failed rift meets the ocean, as in the Niger Delta. However, if subsidence continues there may be marine inundation. In this case, the site of deltaic accumulation will be controlled by the position of the shoreline.

Young passive margins often remain sediment starved long after their initiation because the redistribution of sediment by axial rivers is precluded by marine inundation. This is the case with the Red Sea system (Fig. 4) which receives no major river inputs along the whole of its 2000 km length.

**Patterns of Sedimentation in Rifts**

The analysis of gross relationships between evolving structure and drainage together with recently developed facies models (Frostick and Reid 1987a; Leeder and Gawthorpe 1987) makes a good starting point for an examination of sedimentary sequences in rifts. It would be advantageous to identify features of sedimentation that are common despite the differences in structural and environmental history that inevitably distinguish one system from another.

**Sequences in African Lake-filled Rifts**

Sediments in the East African rift have long aroused interest (e.g. Reck 1951; Isaac 1965; Pickford 1978; Frostick et al. 1986). This is largely because they contain an almost unique record of Cenozoic vertebrates including our own early ancestors (e.g. Leakey and Leakey 1978). As a result, good stratigraphical and sedimentological information is readily available (e.g. Bishop 1978; Frostick et al. 1986). This provides a good basis for inter-basinal comparison of sedimentary sequences.

Examples of Plio-Pleistocene sequences from 3 different lake basins are shown in Fig. 7. The sites are located on Figs. 2 and 4. It is interesting to note that these, and most of the other sequences reported in the literature, are for sites on the unfauluted margin of half graben i.e. on the roll-over or hanging-wall platform. This margin is, topographically, the most subdued and the sequences are therefore likely to be complicated by repeated transgressions and regressions as lake level rises and falls (Frostick and Reid 1986). It is not, therefore, surprising that the 3 sequences depicted in Fig. 7 consist of complex interfingings of lake, lake-shore and fluviatile sediments.
Pedogenic horizons are common (e.g. Fig. 7, Kapthurin Formation of Lake Baringo at approximately 20 m). They represent periods of relative quiescence in the landscape and may correspond with phases of boundary fault inactivity. Eroded surfaces also occur sporadically throughout the sequence (e.g. Fig. 7 at 27 m in both the Hadar and Koobi Fora Formations). These are related to periods of river incision that might be caused by either tectonism or climatic change.

The most striking feature of Fig. 7 is the cyclicity in all 3 sequences, each from widely separated basins. Deposits of silt/clay alternate with sand/gravel. Each cycle varies in thickness between 0.5 and 20 m. They reflect both the instability of base-level and spasmodic changes in sediment supply.
Fig. 7 Sedimentary sequences from 3 East African rift lake basins. Hadar Formation after Taleb and Tercelin 1979; Koobi Fora Formation after Bowen and Vondra 1973; Kapthurin Formation after Tallon 1978. See Figs. 2 and 4 for location.
Sequences in African Marine Rifts

Examples of Miocene to Pleistocene sequences from the Gulf of Suez are given in Figs. 8 and 9. Their location is given in Fig. 5. The Abu Zenima section (Fig. 8) is on the eastern shore close to the end of a major boundary fault. By contrast Wadi Gharandal (Fig. 8) is further north and lies close in front of a boundary fault thought to have been active during Miocene times. Wadi Dara and Wadi Kharim (Fig. 9) are on the western margin of the Gulf in an area where the present-day topography is subdued. At this point along the Gulf, the boundary fault is on the opposite shore and the sites are therefore on the hanging-wall platform. This is a structural setting very similar to that of the sites in the continental basins already described above.

Once more, all of the sequences exhibit marked cyclicity of sediment type and depositional environment. In these cases this largely reflects chan-

Fig. 8. Sedimentary sequences from the eastern coast of the Gulf of Suez. See Fig. 5 for location.
Fig. 9. Sedimentary sequences from the western coast of the Gulf of Suez. See Fig. 5 for location.
changes in the relative levels of land and sea. The alternation of marine, marine shelf, beach and pro-delta sediments with more proximal delta and alluvial deposits is a feature of the Miocene to Recent sedimentation throughout the Gulf of Suez.

Frequently, the sequences contain thick evaporite deposits (Sellwood and Netherwood 1984). This is thought to reflect a structural instability that led repeatedly to the formation of a barrier at the mouth of the Gulf, cutting it off from the Mediterranean (Hassan and El-Dashlouty 1970). However, evaporites are a widely acknowledged feature of marine-inundated continental rift sedimentation (see e.g. Lowell and Genik 1972; Ponte et al. 1980).

CAUSES OF CYCLICAL DEPOSITION IN RIFTS

Within an active rift setting there is a temptation to explain most fluctuations in sedimentation in terms of the effects of faulting. There is no doubt that movements on major boundary faults will have considerable long term effect on both the character and quantity of detritus supplied to the basin and on the relative levels of land and sea or lake (Blair 1987). By diverting rivers, faulting may also cause abrupt switches in the location of clastic sediment inputs, causing a local reduction in sedimentation rates and a change to non-clastic deposition. But it is frequently forgotten that other changes in the character of the basin may also influence the style of sedimentation. Of special importance in lake basins are climatic fluctuations, while in marine-inundated rifts, a significant factor is eustatic changes in sea-level.

Climatic Change and Tectonism in Lake-filled Rifts.

Changes in the levels of the African rift lakes on both short and long time-scales are well documented (Grove et al. 1975; Gasse et al. 1980; Reid and Frostick 1986). These changes are a response to a variety of factors that interplay: seasonal rainfall - e.g. the level of Lake Turkana changes by over a metre each year (Reid and Frostick 1985); climatic change - e.g. Chew Bahir in southern Ethiopia has all but dried out over the past 9000 years in response to increased aridity (Grove et al. 1975); and volcano-tectonic alterations in basin shape - e.g. Lake Turkana is thought to have been affected by the tumescence associated with Miocene lava (Watkins 1986).

Two examples of African lake-level fluctuations over the past 40000 a. are shown in Fig. 10. These are for Lake Abhe in Ethiopia (Gasse 1977) and Lake Mobutu (formerly Albert) in Uganda (Adamson et al. 1980). It is difficult to be absolutely certain about the causes of these variations in lake level. Synchronous and sympathetic changes in widely separated basins tend to suggest that climatic change is responsible. For example, between

Fig. 10. Eustatic sea-level curve (after Vali et al. 1977) and lake level curves of 2 African rift lakes (L. Abhe after Gasse 1977; L. Mobutu after Adamson et al. 1980).
10,000 and 9000 a BP all of the African lakes stood at a high level (Butzer et al. 1972; Adamson et al. 1980; Owen et al. 1982). At this time it is thought that rainfall was considerably greater than at present. As a result of this particular change in water-balance, as well as at other times (Frostick & Reid 1980), lake basins filled up and overflowed from one to the other.

Figure 6 shows the present Lake Turkana catchment alongside an estimate of its extent during the 9000 a BP high lake stand. The catchment of the lake was enlarged considerably in the early Holocene by the overspill of Chew Bahir and the Galla lakes through the Bakate Corridor. There was another significant, though smaller, extension of the contributing drainage area in the south. Here, the Suguta (proto-Lake Logipi) overspilled westward through the Kamuge River to join the Kerio, so circumventing The Barrier that had so recently erupted and divided it from Lake Turkana (Truckle 1976). However, Lake Turkana itself was now no longer the low point of a closed basin. It overflowed northwestward through the Logipi depression to the Nile (Butzer et al 1972; Harvey and Grove 1982). This high lake stand is documented by widespread diatomaceous silty clay and by reefs of *Etheria* sp. almost 100 m above the present lake level (Frostick and Reid 1980; Owen and Renaut 1986).

A lack of coincidence between the lake level curves for different basins does not necessarily rule out climatic change as a cause. Rainfall does not always increase or decrease by the same amount even over comparatively small distances. However, tectonism can only be invoked as a significant cause of lake level fluctuations if there is other corroborating evidence; e.g. the fine lake sediments which result from sudden basin subsidence being followed by an influx of very coarse detritus from the hinterland (Blair 1987), or an identification of growth faults in the sedimentary record. Even without tectonism, cyclical patterns of sedimentation such as those depicted in Fig. 7 can be produced. However, without the basin subsidence that accompanies rifting, the great thicknesses of rhythmic sedimentary sequences that are seen in rifts could not be preserved.

Eustatic Changes of Sea-level and Tectonism in Marine-Influenced Rifts.

Global curves of eustatic rise and fall in sea-level are widely used (Vail et al 1977; Fig. 10) despite reservations about the methods employed in their construction. Although sea level is not as unstable as lake level in internal drainage basins, fluctuations are both large and widespread. The successive eustatic fluctuations of sea level will be recorded as interbedded sequences of marine and aluvial deposits.

Many small-scale cycles of deposition could be linked to periodicity in sediment supply, which itself could have either a climatic or a tectonic cause. The Wadi Gharandal sequence of the Gulf of Suez (Fig. 8) is the one presented here that is most likely to reflect the influence of fault activity, since it is situated at the foot of what was, at the time of deposition, a steep, active fault scarp. In this context it is interesting to note the relative simplicity of the sequence, with its repeated superposition of coarse alluvium and fine marine muds and little evidence of reworking. This contrasts with the complexity of the Abu Zenima section, which, although only 30 km along strike, is on the hanging-wall platform of the neighbouring half graben; the major boundary fault here is on the western side of the Gulf. Changes in river gradient brought about by fault activity have caused much reworking of sediment, while the lower topographic gradient has meant significant shifts of the shoreline during the transgressions and regressions associated with eustatic changes in sea level. This difference in the complexity of sedimentary sequences from one side of a half graben to the other is a feature of one of the latest facies models of rift basin sedimentation (Frostick and Reid 1987a) in which the most complex sequences occur on the unfaulted margin of the rift.

**CONCLUSION**

The topographical consequences of continental rifting are reflected in the river system which, in turn, controls the location, quantity and nature of clastic sediment supplied to rift basins. Drowning in those rifts associated with mantle plumes combines with backtilting of footwall fault blocks to shed rivers away from the developing topographical lows. The faulted side of the basin becomes a bypass margin and is therefore, frequently sediment starved. For example, in Lakes Malawi, Tanganyika and Turkana the drainage area of rivers discharging across boundary faults is only 25, 20 and 10% respectively of the total catchment of each basin (Table 1).

In East Africa and the Red Sea/Gulf of Aden, the catchments of the rift systems are surprisingly small. Only a few, very small streams reach the Gulf of Aden, and only 3 medium-sized rivers - all ephemeral at present - discharge into the Red Sea along its length of more than 4500 km.

The same is true of the East African rift. In some of the lake basins, e.g. Tanganyika, the total catchment is only 4 times the surface area of the lake itself. Major rivers entering the rift are few, and tend to utilise either the accommodation zones between sub-basins where the obstructive boundary...
faults die out, or the hanging-wall platform cum roll-over. The unfaulted margin of the rift becomes the major source of sediment (Table 1). However, accommodation zones are frequently the sites of large-scale accumulations of clastic sediments e.g. the Kerio River delta on the western shore of Lake Turkana. In fact, the faulted margin may be an area of the basin that is relatively starved of river sediment and may be a site of organic-rich or carbonate deposits.

The fact that the rift is a series of linked subbasins during the early phases of development discourages the integration of drainage along the rift axis. However, in later phases, when the intensity of faulting diminishes and regional subsidence plays an increasingly important role, large axial rivers may run the length of the rift, as in the Benue trough. Whereas young rifts are comparatively starved of siliciclastic inputs and may accumulate organic-rich deposits which act as petroleum source-rocks, later phases of rifting are characterised by widespread alluvium which may act as the potential reservoir rock.

The sedimentary fill of rift basins reflects not only the influence of tectonism, but also the effects of environmental changes that are independent of structural control, notably climatic fluctuations and eustatic changes in sea-level. In closed lake basins, shifts in climate produce changes in water-balance which are reflected in lake level. During wetter phases, lakes extend and may overflow one to the other. In marine basins, eustatic changes in sea-level will affect sedimentation in a way that is not dissimilar to the influence of climatic change on lake basins. The rise and fall of sea-level will produce cyclical sedimentary sequences almost identical to those produced as a consequence of boundary fault activity.

The lateral extent of the inundation resulting from transgression is of economic significance. This is because fine lake or marine sediments act as a seal to the more permeable alluvial deposits with which they interdigitate. The degree of inundation depends not only on the magnitude of the rise in water level, but also on topographic factors. In narrow, steep-sided rifts such as Tanganyika, a 5 m rise in water level would shift the shoreline by 30 m (Evert 1980). In contrast, the same rise in water level in Lake Turkana would cause an average shift of 800 m (Butzer 1971).

The complex interplay of tectonism, climatic change and eustatic adjustments of sea-level in rift basins makes it difficult to isolate specific causes of changes in depositional style. There is a temptation to cite tectonism as the major factor. However, this understates the role of the other factors and can lead to serious errors in palaeoenvironmental reconstruction when considering ancient rift sediments.

Acknowledgements: The authors' familiarity with the East African rifts arises from research financed by the National Environment Research Council of the UK. We would like to thank Richard Leakey for his help and encouragement over the years. The work in Egypt was made possible through a grant from Norsk Hydro plc and by the generous assistance of the Egyptian Geological Survey. We are indebted to Ali Mazzer for his help in the field, and to Bruce Sellwood for his advice before our fieldwork.

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