Evolution of large normal faults: Evidence from seismic reflection data

Chris K. Morley

ABSTRACT

Recent advances have been made in understanding how extensional faults and basins develop as faults propagate and link. Evidence for these linkage patterns in seismic reflection data can be seen in data from East Africa. Early fault linkage patterns for boundary faults can follow three possible paths. Fault linkage and propagation occur either (1) prior to significant basin formation, (2) after minor faulting has created an extensive area of subsidence, or (3) during basin development. The data from East Africa show examples mainly of paths 1 and 2. Transverse anticlines (anticlines developed parallel to and in the hanging wall of the strike of faults) associated with boundary faults are common features. They represent either the sites of old synthetic transfer zones or a region of low fault displacement along the strike of a fault where two or more depocenters of different ages overlap. As fault activity decreases over time, displacement tends to be concentrated on progressively narrower parts of the fault. This pattern is developed particularly well in continental rifts and may help discriminate late synrift sedimentation from postrift sedimentation where strike lines across the hanging wall of the fault are observable.

INTRODUCTION

The evolution of normal faults has received considerable attention over the last few years as research from several areas has converged. Extensive studies of fault populations have tried to establish rules governing basic fault dimensions, such as fault length vs. displacement and fault displacement vs. number of faults of a particular size in a population (e.g., Muraoka and Kamata, 1983; Barnett et al., 1987; Walsh and Watterson, 1988; Marrett and Allmendinger, 1991). Numerical modeling, experimental data, and field observations of fault propagation and linkage have led to the development of models that predict the way displacement dies out along a fault toward the fault tip. Such models can then be tested against natural

Manuscript received May 26, 2001; revised manuscript received June 4, 2001; final acceptance January 2, 2002.

AUTHORS

Chris K. Morley ~ *Department of Petroleum Geoscience, Universiti Brunei Darussalam, Bandar Seri Begawan, Brunei Darussalam; cmorley@mail.ubd.edu.bn*

Chris Morley is currently an associate professor in the Department of Petroleum Geoscience at the University of Brunei Darussalam. Previously, he worked for Amoco, Elf Aquitaine, and Aberdeen University. He has worked extensively on the structural geology of sedimentary basins and the impact of structure on sedimentation. His current areas of interest include deformation associated with mobile shales, the structural geometry and evolution of rift basins, the impact of fault segmentation and linkage on sedimentation, the integration of outcrop and subsurface structural geology, and the structural and tectonic evolution of southeast Asia. His key geographic areas of research include the East African rift system, the Baram Delta province, Brunei, and the Tertiary rift basins of Thailand.

ACKNOWLEDGEMENTS

I would like to thank Amoco Petroleum Company and the National Oil Company of Kenya (as it then was) for allowing me access to publish the East African rift data and the many colleagues in Amoco who helped to work up the data set. The Cairo Hedberg conference and field trip to the Gulf of Suez led to many stimulating discussions, some of which have found expression in this article. Thanks are also due to John Dolson and GUPCO for funding that indirectly led to this article.

fault populations (e.g., Cowie and Scholz, 1992; Burgmann et al., 1994; Cowie and Shipton 1998). As well as considering individual faults, the simultaneous evolution of numerous faults during linkage also has been the subject of both numerical modeling (e.g., Cowie, 1998) and outcrop studies (Peacock and Sanderson, 1991; Cartwright et al., 1995; Dawers and Anders, 1995).

One of the challenges of testing the models is to find natural examples that permit the details of fault evolution to be determined. Various categories of natural examples have been used:

1. Mesoscopic, well-exposed faults in outcrop that show different stages of fault development frozen in time (e.g., Cartwright et al., 1995). With the absence or poor exposure of syntectonic sedimentary rocks, only the relative timing of structural development can be deduced. The three-dimensional (3-D) geometry of any one fault zone cannot be determined.
2. Large, well-exposed regions where the syntectonic sedimentary rocks can be investigated in detail. Inferences drawn from the sedimentation patterns can be used to reconstruct the development of associated faults (e.g., Newark basin, USA, Schlische [1992]; Gulf of Suez, Gawthorpe et al. [1997]). No subsurface perspective exists, so each example represents a part of only the fault evolution.
3. Two-dimensional (2-D) seismic reflection data across large faults. Such data sometimes permit determination of the evolution of the fault through time where the synextension sedimentary sequence is sufficiently thick and internally reflective. The 2-D data, however, do not permit detailed mapping of the fault geometry (Morley, 1999a, b; Contreras et al., 2000).
4. Three-dimensional seismic reflection data. These data have the advantages of category 3, plus greater details of the map-view geometry of the fault can be determined (e.g., Mansfield and Cartwright, 1996; Morley and Burhannudinnur, 1997; Dawers and Underhill, 2000).
5. Detailed investigations of individual fault zones (Dawers et al., 1993; Peacock and Sanderson, 1996; Burhannudinnur and Morley, 1997; Cowie and Shipton, 1998).

Very few studies have used seismic reflection data to determine how faults have evolved through time. One reason is that the resolution of the seismic data affects how much the early linkage pattern can be seen. Hence, to examine the details of fault evolution with confidence, it is best to study major boundary faults, which may have strike lengths of between 50 and 200 km and have some age control from well data. This raises the problem of finding data sets that are sufficiently extensive and complete. Four such studies are Morley (1999a), Contreras et al. (2000), Dawers and Underhill (2000) and McLeod et al. (2000). This article presents the results of seismic reflection surveys across large extensional faults that illustrate some of the ways in which large faults evolve.

EARLY GROWTH OF FAULTS

To successfully test numerical models of fault growth and linkage (e.g., Cowie and Scholz, 1992; Burgmann et al., 1994; Cowie and Shipton, 1998), it is necessary to compare them with natural examples. Key questions include the following: How much time does it take for faults to propagate together and form major boundary fault systems? What are the typical patterns of fault linkage? How can the early linkage pattern be detected from seismic reflection and outcrop data?

It has been intuitively obvious to generations of geologists that large faults have grown by linkage of smaller faults. This is implicit in models of crack growth for the earliest stages of faulting (e.g., Griffith, 1924; Brace and Bombolakis, 1963). The time taken to achieve evolution from small cracks to large faults, however, has not been the subject of many detailed studies. Individual crack propagation can be very fast (close to the speed of sound [Griffiths, 1924]), yet larger scale fault propagation and linkage and the buildup of significant displacement appears to be much slower and is even detectable in the sedimentary record (e.g., Anders and Schlische, 1994; Gawthorpe et al., 1997). Fault development from short, isolated faults to major boundary faults may be so slow as to be seen in the tectonic subsidence record. For example, the first 6 m.y. of synrift subsidence in the Gulf of Suez is thought to have occurred predominantly on relatively short, numerous faults. Only toward the end of this somewhat prolonged stage did large boundary faults develop (Patton et al., 1994; Gupta et al., 1998). This change in fault geometry is accompanied by a marked acceleration of tectonic subsidence (Gupta et al., 1998).

Schlische (1991) proposed that the growth of boundary faults and accompanying growth of sedimentary basins produced a predictable pattern of axial and flexural margin onlap that should be seen in the sedimentary record. As the boundary fault gets longer, the extent of the basin gets larger and causes progressive onlap of the axial and flexural margins (Figure 1a). Schlische and Anders (1996) proposed that the linkage of faults could be detected in the filling pattern of basins, and they presented several different potential filling patterns. The basic model assumption was that boundary faults grew by along-strike propagation and by linkage with other faults during basin development (Figure 2a). The resulting gradual variations in displacement along boundary faults were predicted to affect the location and thickness of synextension sedimentary rock units deposited at various stages of boundary fault development (Figure 1).

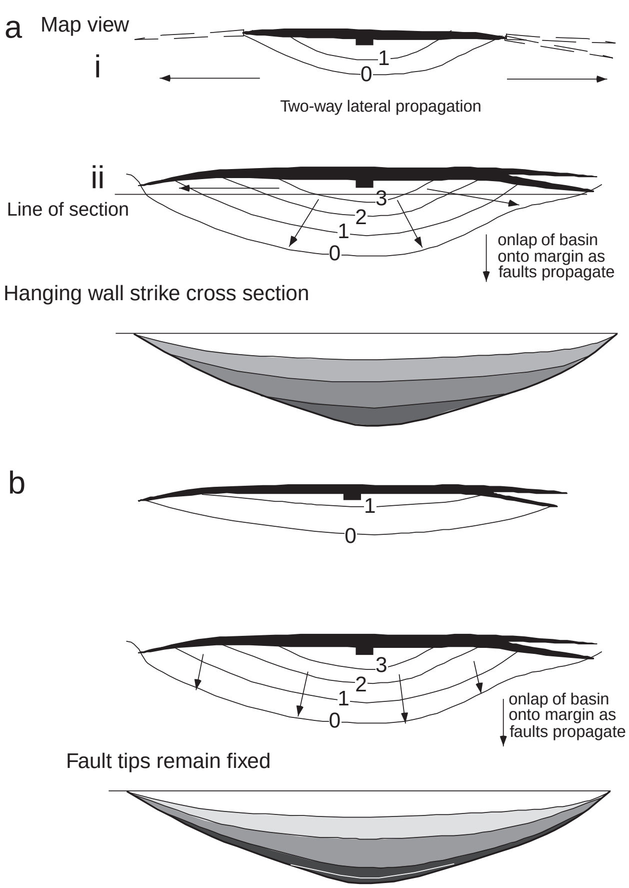


Figure 1. Two basic models for boundary-fault propagation and depocenter development. (a) Development of a sedimentary basin during boundary-fault propagation (modified from Schlische and Anders, 1996). (b) Development of a sedimentary basin that occurs largely after the boundary fault has propagated (Morley, 1999a). i = early stage; ii = later stage; 1–3 = structure contours (depth in km).

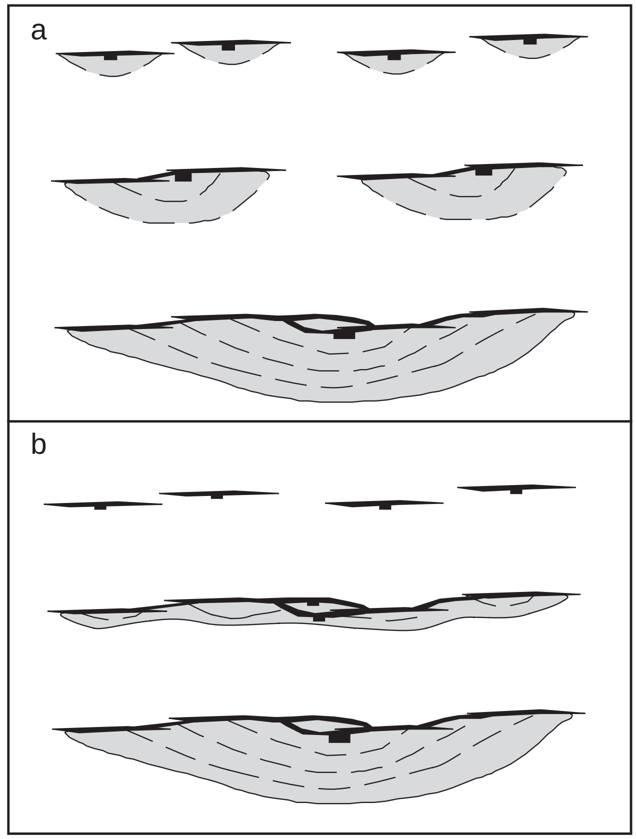


Figure 2. Schematic illustration of how the basic models a and b in Figure 1 might develop. (a) Linkage of progressively larger faults over a relatively long period of the basin history (based on models by Schlische and Anders [1996]). (b) Early linkage of minor faults to form a long boundary fault. The basins lie in a narrow belt along the hanging wall of the fault. As the fault builds displacement, there is little axial propagation, and the basin progressively onlaps the flexural margin.

Morley (1999a) examined some boundary faults in East Africa to test the Schlische and Anders (1996) model. He found little evidence of progressive propagation of boundary faults during the main stage of basin development (Figures 1b, 2b). The lowest synrift reflections displayed little evidence for early fault segmentation, although two examples (Lupa fault, Lake Rukwa, and Lokichar fault, north Kenya) showed evidence of development of some early isolated depocenters along later boundary faults.

The results of Morley (1999a) showed that in East Africa fault propagation and linkage occurred prior to significant basin development (Figure 2b). In this model, axial propagation of the rift basin occurs earlier than the timing of most pronounced basin onlap onto the flexural margin. This represents an end member to the model developed by Schlische and Anders (1996), which is based upon significant basin development occurring during fault linkage.

In nature, examples probably exist that cover the range of geometries between the two models. Morley (1999a) assumed that linkage of minor faults to form a major boundary fault occurred rapidly (possibly on the order of hundreds of thousands of years) because of the absence of early, isolated basins at the base of the synrift sequence. The presence of such basins would indicate earlier, unlinked faults that later amalgamated to form full-length boundary faults. Gupta et al. (1998), however, have shown that for the Gulf of Suez, the rift initiation stage can be associated with thin (less than 200 m) basin fill that has a duration of several million years. Such thin sequences might be difficult to detect from seismic data. Hence, the time required for linkage remains somewhat ambiguous. Sediment supply might also be a factor, in that a rift system that links and propagates relatively slowly might be expected to be a filled or overfilled system. If sediment supply to the same fault system is low, however, and the basins remain underfilled, then the system might incorrectly be inferred to have propagated and linked relatively rapidly.

A third kind of early rift basin fill is exhibited by the Usangu Flats in Tanzania (Figure 3). This region is one of the youngest areas of rifting in the East African rift system and was initiated probably about 2 Ma (Ebinger et al., 1989). A boundary fault has been inferred along the northwestern margin of the basin as a result of surface morphology (Figure 4). Sparse seismic reflection data suggest that this boundary fault does not have much displacement and does not control the basin geometry. A map made at the base of the synrift section shows the early rift basin fill to have a broadly synclinal geometry (Figure 4). The thickening toward the basin center is achieved by expansion of section across numerous rotational and nonrotational normal faults (Figure 5). These faults are of similar displacement, and no dominant fault trend is apparent. The total thickness of the basin fill is about 1 km (Harper et al., 1999). This region illustrates the first stage of development for a half graben where the boundary fault develops relatively late in the basin evolution (Figure 6). The large basin was established well before the boundary fault became dominant and represents the kind of rift initiation stage envisioned for the Nukhul Formation in the Gulf of Suez (Patton et al., 1994) and from numerical models (Cowie, 1998). Figure 5 shows a seismic example of the minor faults from the Usangu Flats. It illustrates a region where the faults

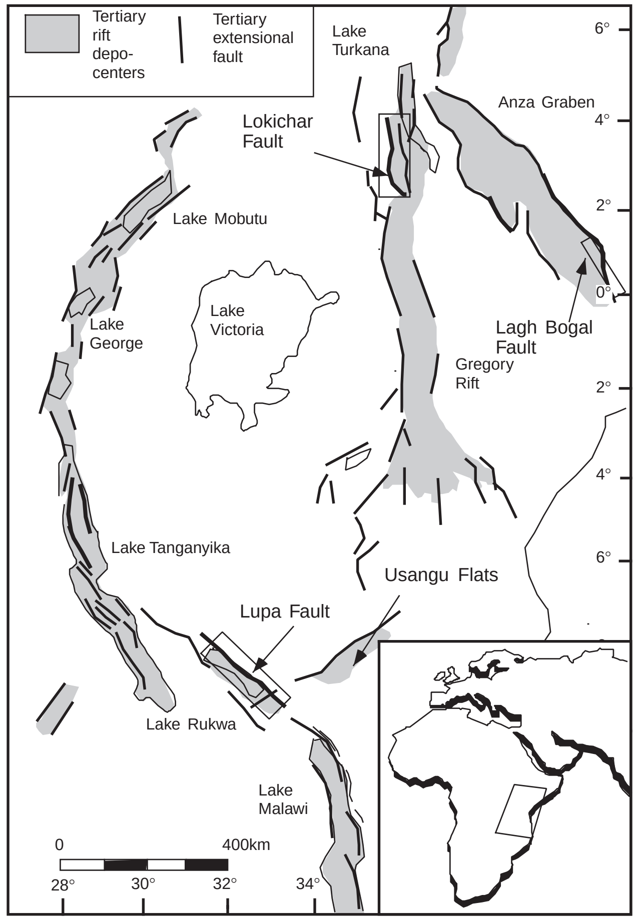


Figure 3. Location map of faults from the East African rift system discussed in this article (modified from Morley, 1999a).

are mildly rotational, although no dramatic expansion of section into the faults can be seen. The largest fault is associated with a fault-propagation fold in its hanging wall, reminiscent of those described in the Gulf of Suez (Patton, 1984; Patton et al., 1994; Gawthorpe et al., 1997; Sharp et al., 2000). The fold may even be cored by high-angle to overturned curved faults, similar to those predicted by finite-element modeling by Patton (1984). These folds seem to develop commonly at the tips of blind propagating faults (possibly at relay ramps), particularly where there is a pronounced lithology change, such as the passage from crystalline basement to sedimentary rocks.

The Usangu Flats basin and faults trend northeast-southwest, an uncommon direction for structures in the East African rift system. The basin is located on Archean cratonic rocks, which is unusual because the rift system tends to be located within the Proterozoic mobile belt that surrounds the Tanzanian Archean craton (McConnell, 1972) (Figure 3). The basin might

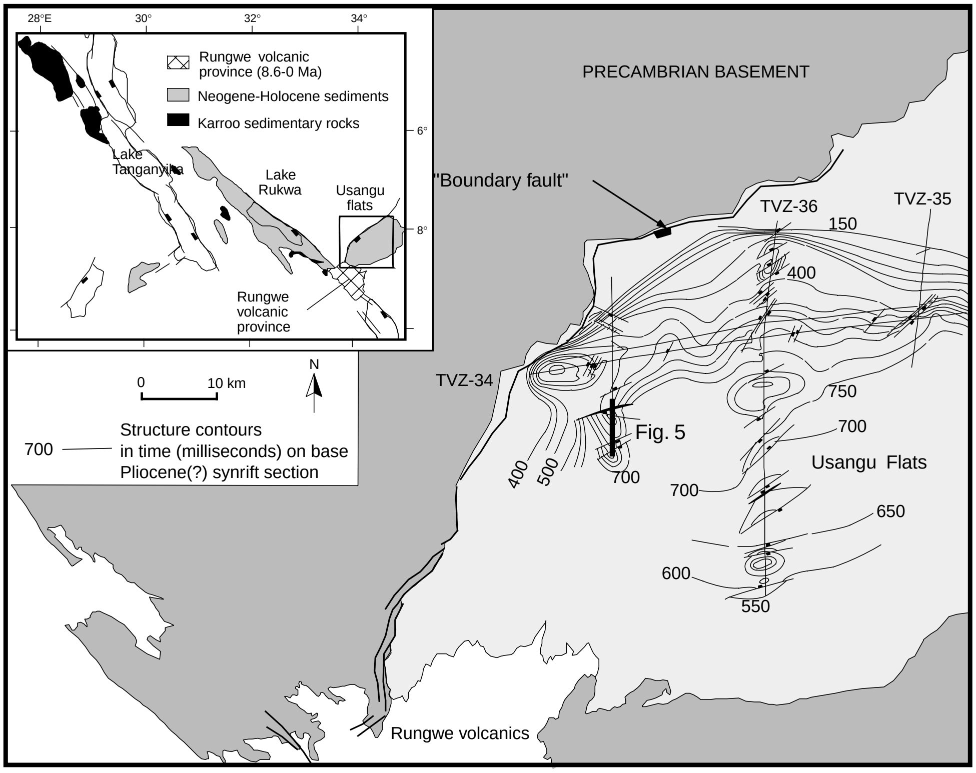


Figure 4. Geological map of the Usangu Flats area, Tanzania (modified from Ebinger et al., 1989) and structure-contour map of the base, late Tertiary, synrift section (based on seismic reflection data from Harper et al. [1999]; 50 ms two-way traveltime [TWTT] contour interval).

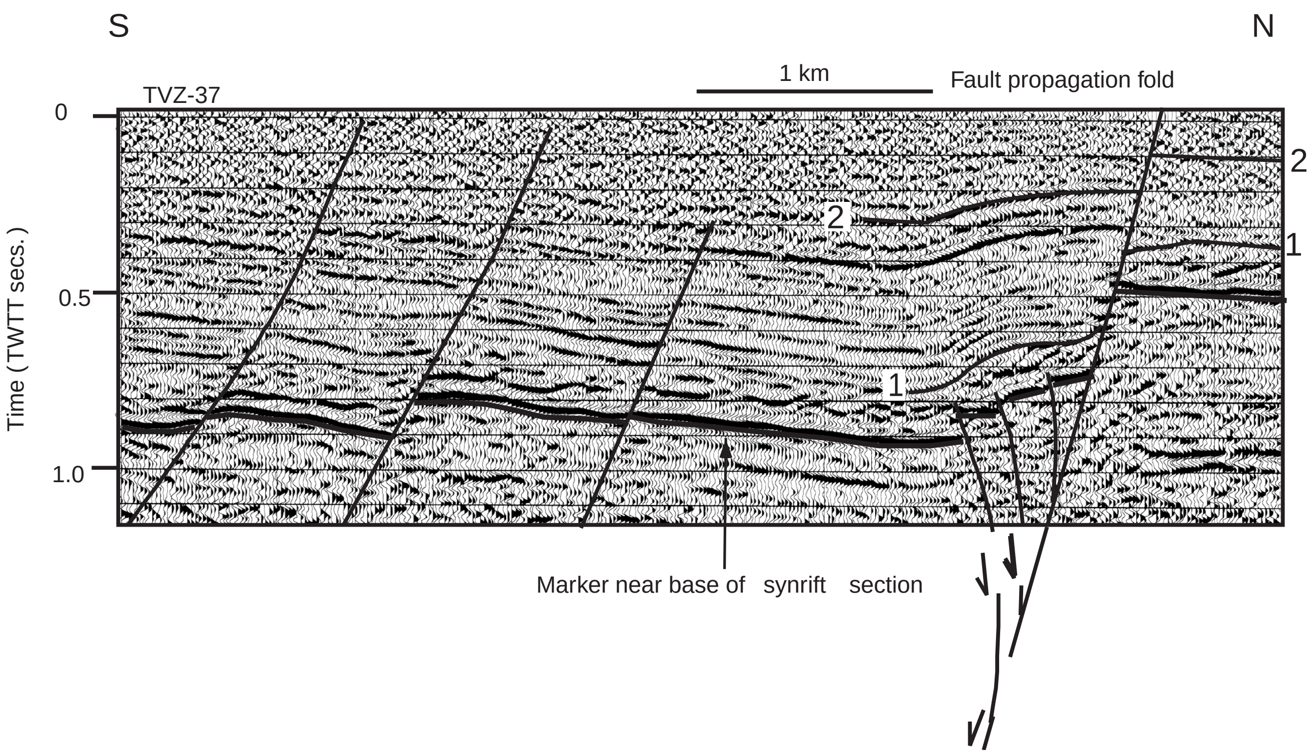


Figure 5. Portion of seismic line TVZ 37 from the Usangu Flats showing minor faults characteristic of the early stages of rift development. Note that the largest fault is associated with a fault-propagation fold (modified from Harper et al., 1999). See Figure 4 for location.

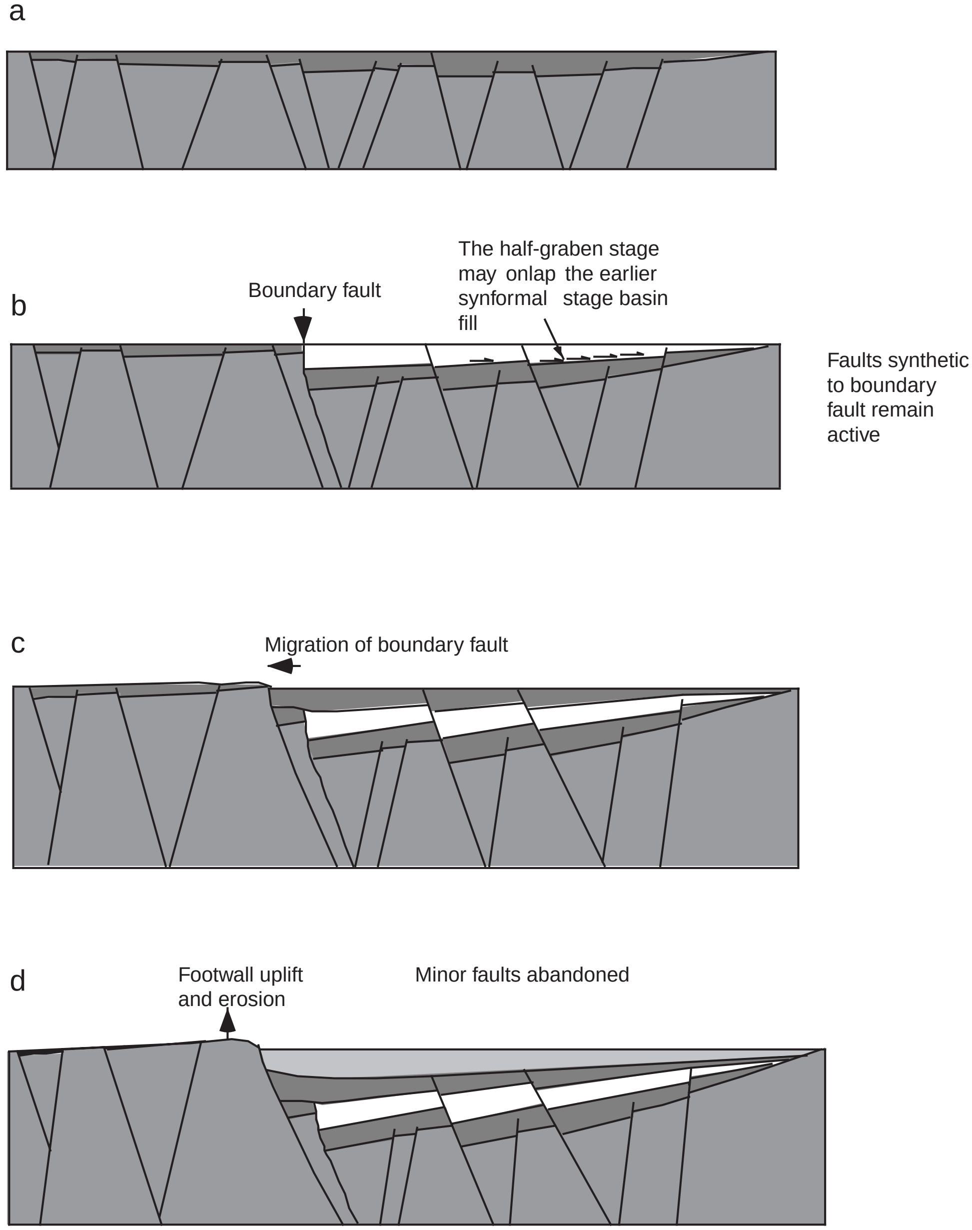


Figure 6. Schematic cross sections illustrating an idealized evolution of a half graben and highlighting the main structural stages commonly found in East African rifts. (a) Early rift stage, synformal depression (e.g., Usangu Flats). (b) Early half-graben stage. (c) Mature half graben. (d) Late-stage half graben.

have developed during a late rotation in extensional stresses from a more east-west extension direction to northwest-southeast (Ebinger et al., 1989; Ring et al., 1992). Because of its location on the Archean craton, the effects of preexisting fabrics on fault development might be less in the Usangu Flats than in other parts of the East African rift system.

Other large boundary faults in East Africa do not necessarily display a well-developed Usangu Flats–type stage, and evidence for early fault linkage must be found in the fault patterns themselves. Fault striations are not commonly found in the Turkana area, but where they are present, show normal dip-slip to oblique dip-slip orientations (generally indicating east-west extension) (see also Strecker et al., 1990). The Lokichar fault of northern Kenya is a rare example; the geometry of early fault linkage can be seen both in the early basin fill and in the fault geometry (Morley, 1999a) (Figure 7). The fault is characterized by several fault splays that extend into the hanging wall. These splays were active relatively early in the basin history and were deactivated later in favor of a fault segment that lies on the footwall side of the splay. The splays represent the earlier geometries of two separate faults that are separated by a synthetic transfer zone. Later, the two faults became linked across the transfer zone

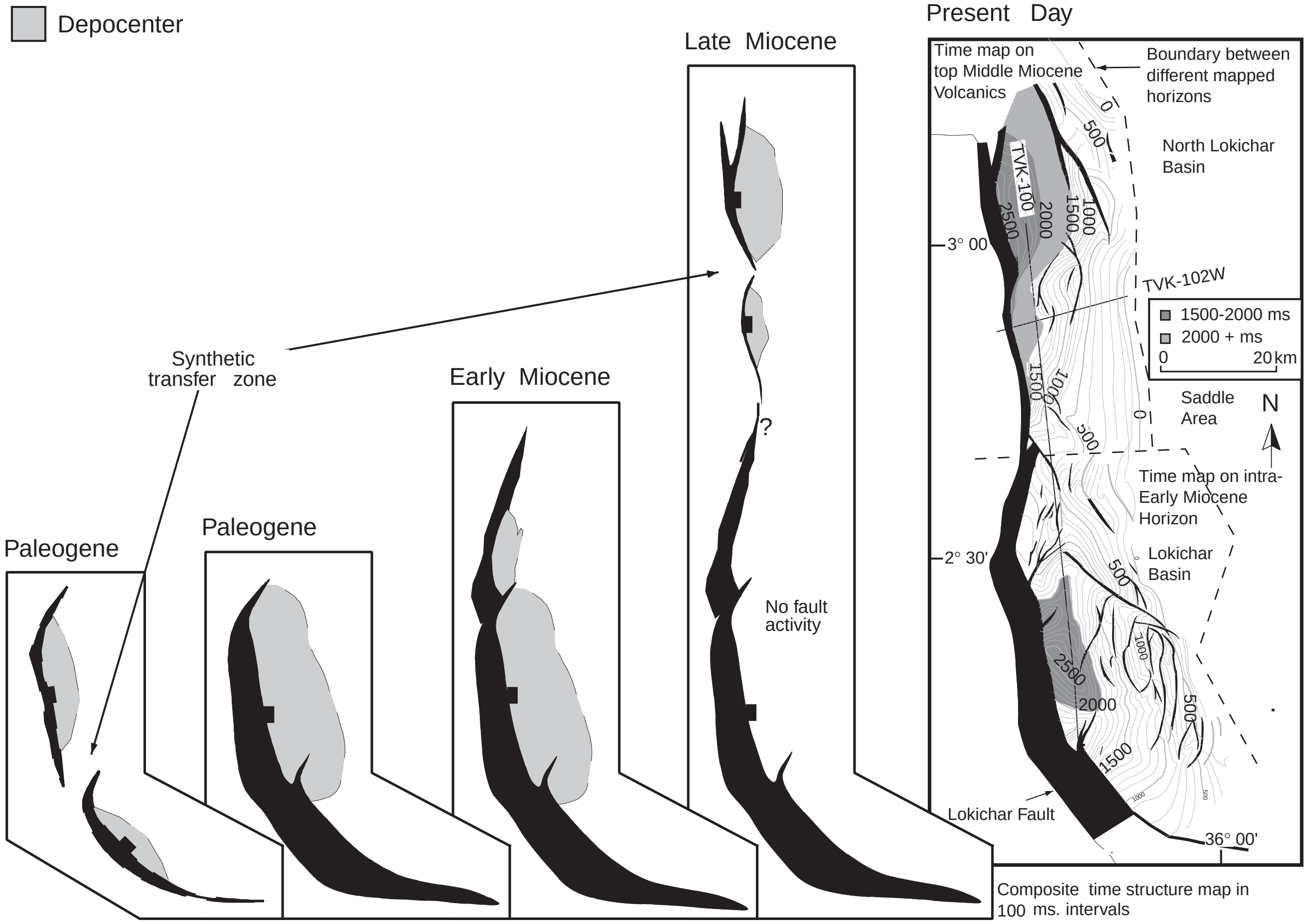
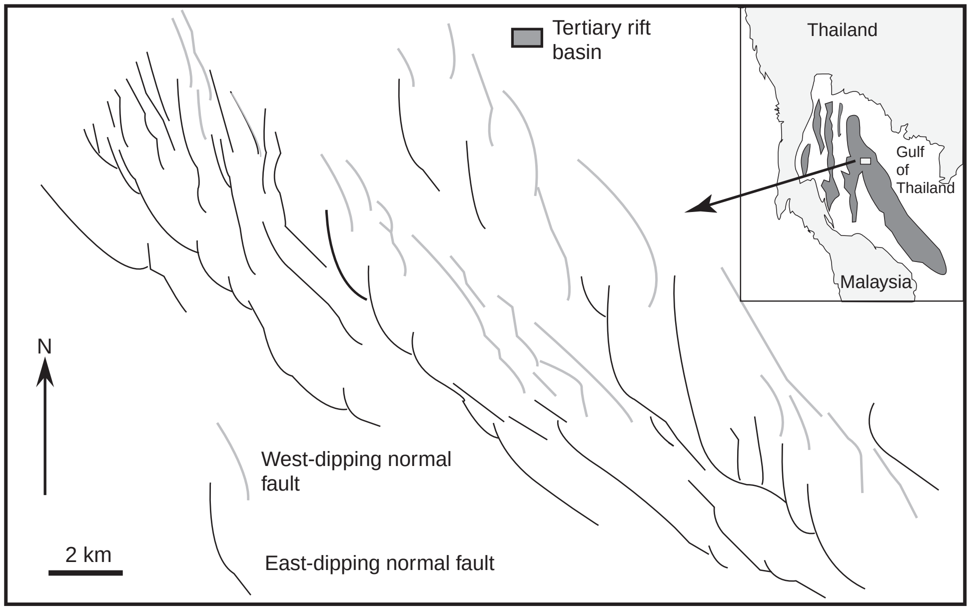


Figure 7. Map-view evolution of the Lokichar fault, northern Kenya. The fault zone appears to have evolved by amalgamation of two initially separate faults in the southern part of the fault zone, the associated basins joined, and half grabens developed during the Paleogene and early Miocene. The northern part of the fault zone developed during the late Miocene–Pliocene and joined with the largely inactive Lokichar basin segment of the fault. The structure map is based on seismic reflection data presented in Morley et al. (1999a).

(Figure 7). This is a common early linkage geometry (e.g., Griffiths, 1980; Morley et al., 1990; Peacock and Sanderson, 1991; Cartwright et al., 1995).

The linkage geometry described for the Lokichar fault (Morley, 1999a) (Figure 7) is a common feature in rifts. An example described by Dawers and Underhill (2000) for the Statfjord East area, northern North Sea, showed that changes in isopach geometries documented the fault-linkage evolution. One extreme example of the linkage geometry having splays into the hanging wall comes from the northwestern Malay basin (Bongkot field area) in the Gulf of Thailand (e.g., Duval et al., 1995), where faults zones are commonly composed of as many as 10 (and in some places 20) joined, curving fault splays (Figure 8). Displacement maxima on structure contour and isopach maps at the center of each curved splay confirm that the fault zones were composed initially of isolated faults that subsequently joined. The faults are composed of northwest-southeast– and north-south–striking segments, and the influence of preexisting basement fabrics on promoting the curved splays is suspected strongly (e.g., Watcharanantakul and Morley, 2000). The extreme variations in timing of fault linkage with respect to basin development are present in these examples. In the Lokichar basin, the splays are evident, but the basin isopach is so overwhelmed by the later displacement history that the early unlinked-displacement patterns cannot be detected. In the North Sea and northwestern Malay basin examples, however, the individual prelinkage fault-displacement patterns can still be seen. In the case of the Bongkot field, the fault segmentation is clearly reflected in the distribution of individual hydrocarbon accumulations within the field.

Figure 8. Map of the Bongkot field area, northwestern part of Malay basin, Thailand, illustrating a fault pattern characterized by curved faults that originally were isolated faults that then linked to form considerably longer faults having numerous splays curving into the hanging wall (map redrawn from Duval et al. [1995]).



Not all rift geometries conveniently fit into the pattern of evolution discussed previously. Some seismic lines show very little evidence for the early development of numerous minor faults. Line 38 from Lake Tanganyika (e.g., Burgess et al., 1988) shows two large boundary faults that have a flat-lying graben floor between them where only one small fault can be seen. This line indicates that, in some circumstances, fault linkage must occur very early and rapidly so as to inhibit almost completely minor fault development (Morley, 1996).

DEVELOPMENT OF BOUNDARY FAULTS

Seismic data from East Africa show that once boundary faults develop, their tips appear to remain fixed for long periods of time (Figure 9). The simplest basin-fill patterns parallel to the strike of a fault show maximum thickness in the approximate center of the fault and gradual thinning toward the fault tip (Figure 1b). If the boundary fault had propagated during basin development, the basin fill would onlap the prerift basement (Figure 1a) (Schlische, 1991).

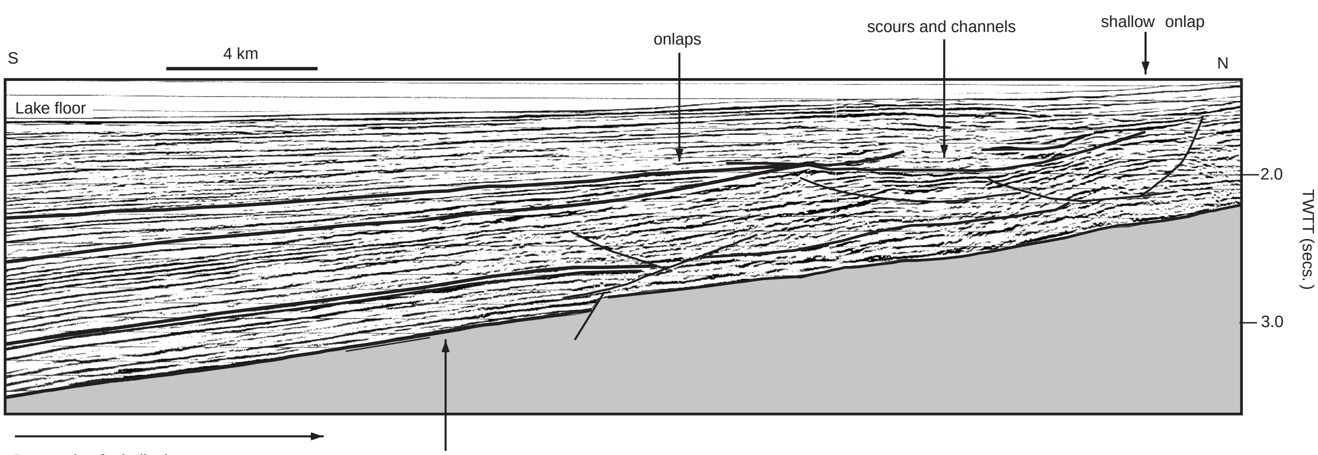


Figure 9. Portion of Project PROBE line 11, reprocessed by Amoco, illustrating a strike section through part of the hanging wall of the East Kigoma fault, Lake Tanganyika. Note the gradual thinning of the reflection packages toward the fault tip and the absence of onlap terminations onto Precambrian basement.

A common pattern of half-graben development is for faults synthetic to the boundary fault to develop into the dominant faults (e.g., Morley, 1995). Many of the early faults, particularly those antithetic to the boundary fault, become inactive (Figure 6). This may be a result of reduction of stresses in a halo around faults that formed early, which tends to inhibit the development of antithetic faults and promote synthetic fault formation, as described by Price and Cosgrove (1990). Later in the development of a half graben, secondary faults also may become inactive; in this case, extension becomes concentrated on the boundary fault (Figure 6). Similar patterns of fault development have been modeled numerically in simulations of fault propagation (Cowie, 1998). The stress fields set up around active normal faults either enhance or relax stresses on nearby faults (King et al., 1994; Hodgkinson et al., 1996). In general, faults lying along strike and slightly overlapping each other tend mutually to enhance stresses and promote failure, whereas colinear faults are in stress shadows and tend to be less active. Such modeling predictions are supported by rock experimental data; for example, Brace and Bombolakis (1963) found that systems of en echelon cracks propagated under a fraction of the applied stress required to cause growth of similar isolated cracks. The results of the studies discussed previously indicate that as certain faults become larger, they become more dominant and lead to inactivity of increasingly more of the smaller faults (as reviewed by Cowie [1998]). Initial results suggest that the scaling of these models to geological time scales is reasonable (e.g., Gupta et al., 1998). Numerical modeling shows that there is no need to infer changes in strain rate to accomplish changes in fault linkage geometry, such as slow subsidence in the rift-initiation phase to rapid subsidence in the rift-climax phase. Fault propagation and linkage under a constant strain rate can achieve the same effect (Gupta et al., 1998).

Although there is no need to infer changes in strain rate, faults almost certainly display cyclic activity over a range of time periods. Changes in strain rate on individual faults may reflect both regional changes in strain rate and the way strain is distributed locally on fault arrays. Considerable evidence points to pulsed fault activity over different time spans as follows. (1) Faults display pulses of earthquake activity at time scales of tens to hundreds of years (Machette et al., 1991; Marone, 1998), and this episodicity can be described by frictional laws (e.g., Scholz, 1998). (2) Evidence for pulsed activity on the scale of thousands to tens of thousands of years is just beginning to be gathered (Morley et al., 2000). (3) Sequence stratigraphy in rifts shows that for long time periods (hundreds of thousands to millions of years), pulsed structural activity, together with the effects of climate and sediment supply, can exert an important control on sedimentary sequence geometries and their bounding surfaces (e.g., Sellwood and Netherwood, 1984; Purser et al., 1990; Underhill, 1991a, b; Gawthorpe et al., 1994; Ravnas and Steel, 1998; Morley, 1999b).

Typically, boundary faults in the East African rift are associated with a maximum thickness of synrift section of 7–8 km, and the duration of fault activity

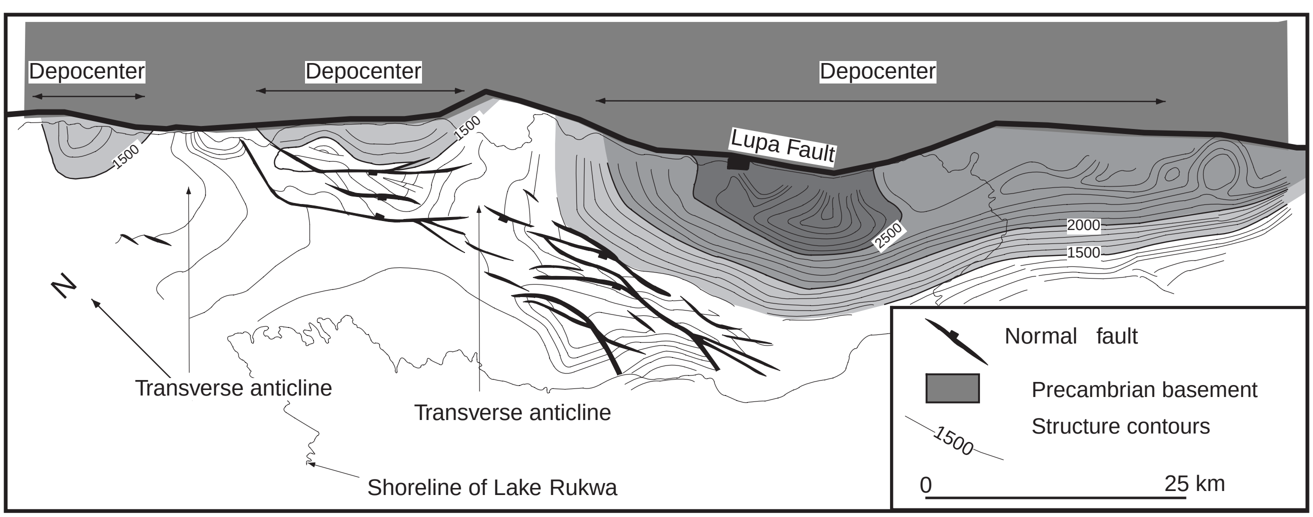


Figure 10. Structure map of the central and southeastern Rukwa rift on the base of the Lake Beds (late Tertiary) based on seismic reflection data (modified from Morley et al., 1999b).

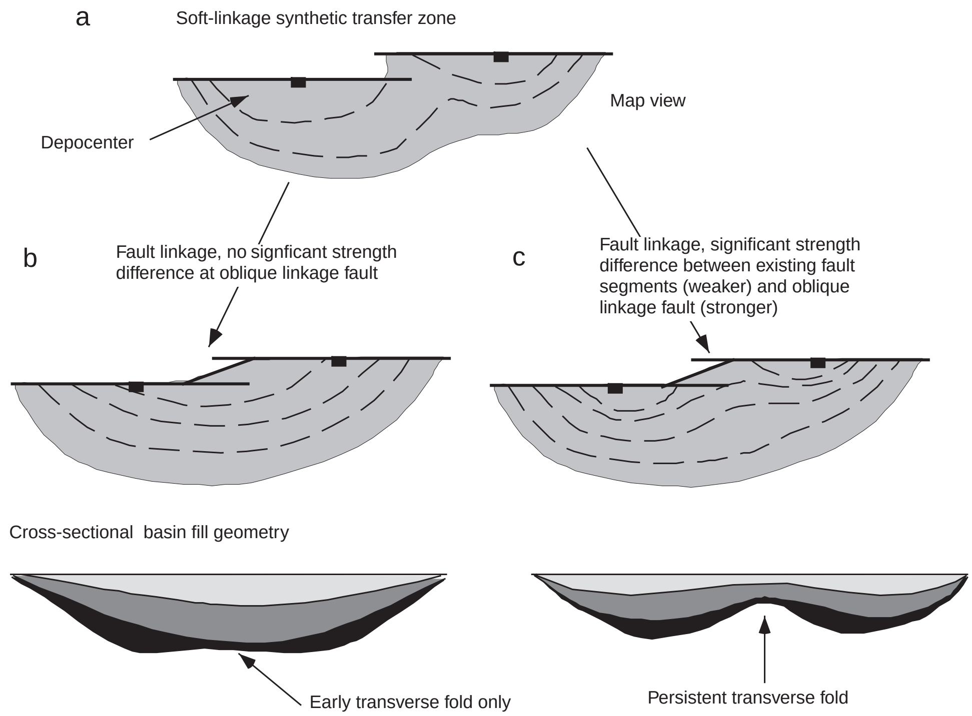


Figure 11. Schematic evolution of faults amalgamated at a synthetic transfer zone (a). Once joined, the displacement on the fault zone may alter, so that the displacement maximum lies at the approximate center of the newly amalgamated fault (b). The older, smaller depocenters can be seen only at the base of the basin fill, where a transverse anticline may underlie the later depocenter. Alternatively, the old displacement patterns may prevail (c) despite the new fault length, which leads to the development of a transverse anticline that is persistent throughout the basin history.

typically is as much as 10 m.y. Hence, maximum time-averaged throw-displacement rates on major normal faults in East Africa are about 1 mm/yr (e.g., Lupa fault, Rukwa rift; Lokichar fault, Turkana [Morley, 1989]). For the Rukwa rift, however, high-resolution seismic data suggest that for short periods of time (thousands to ten of thousands of years), fault-throw displacement rates may be as high as 6 mm/yr (Morley et al., 2000). Therefore, to produce an average of 1 mm/yr displacement over the approximately 7 m.y. life span of the fault would require considerable periods of inactivity to compensate for pulses of rapid displacement.

EVOLUTION OF RIFT BOUNDARY FAULTS

Once a boundary fault has been established, its subsequent history can be highly variable. The shortest major boundary faults in East Africa are about 50–70 km long. Larger ones, such as the Lokichar and Lupa faults, are 150–200 km long. Fault evolution tends to be more complex the longer a fault remains active, which is typically between millions and tens of millions of years. Displacement patterns also are highly variable, ranging from fairly static growth of displacement (by repeating similar patterns throughout its life) (Morley, 1999a; Morley et al., 2000) to marked

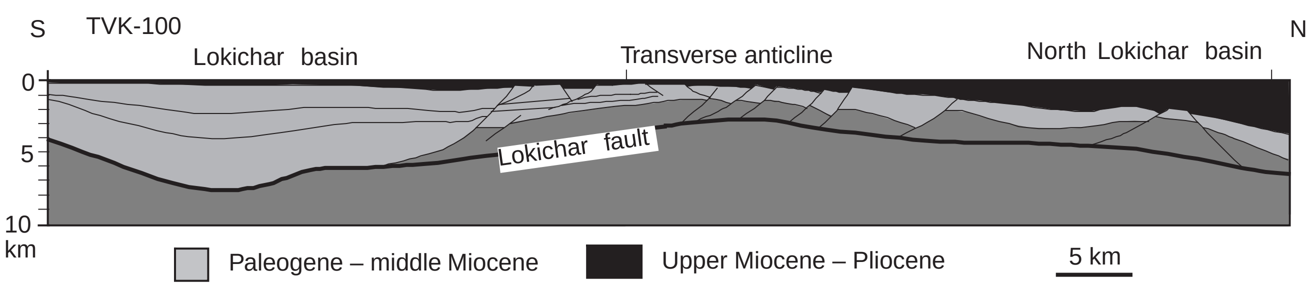


Figure 12. Geological cross section of a strike section along the Lokichar fault based on strike seismic line TVK-100. It illustrates two separate depocenters created at different times, having a very deep Paleogene–middle Miocene depocenter in the south and an upper Miocene–Pliocene depocenter in the north separated by a high area (modified from Morley, 1999a). See Figure 7 for location.

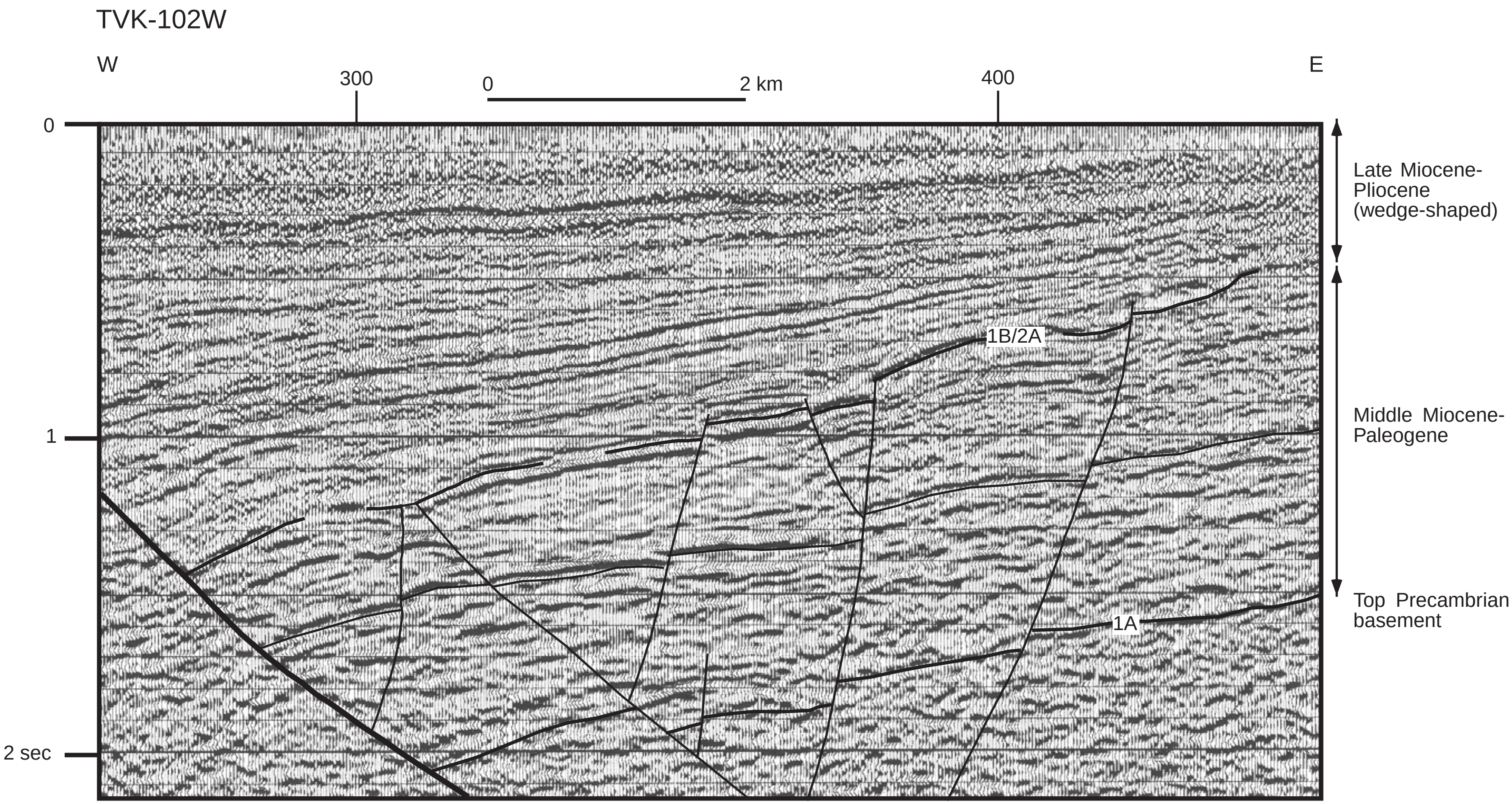


Figure 13. Line 102 part of TVK-102W, western Turkana, illustrating an expanding wedge of synrift basin fill of late Miocene–Pliocene age; its geometry is controlled by the Lokichar fault. The base of this sequence, which lies above crystalline basement, is marked 1A. Underlying the westward expanding wedge is a middle Miocene–Paleogene sequence that expands in the opposite direction. This lower wedge is affected by minor faults that do not affect the higher wedge and that terminate at the unconformity 1B/2A. The transition between the two wedges is marked by a sequence that appears to infill the older wedge topography. See Figure 7 for location.

changes in the location of displacement and further propagation and linkage events (Schlische and Anders, 1996; Morley, 1999b; Contreras et al., 2000; Dawers and Underhill, 2000; McLeod et al., 2000).

After the initial propagation and linkage phase, the fault pattern in some rifts remains relatively static, as displacement tends to keep building in the same places. The Rukwa and Tanganyika rifts in the western branch of the East African rift systems are two such examples. The early linkage geometries are apparent in places as transverse anticlines along boundary-fault trends (Figure 10) that correspond to areas of relatively low displacement along the boundary faults (e.g., Morley et al., 1990, 1999a, b; Morley 1999a). These anticlines probably correspond to the sites of old synthetic transfer zones. The Lupa fault in the Rukwa rift shows well-developed transverse anticlines (Figure 10), which persist in the present sedimentation pattern, approximately 8 m.y. from the start of extension (Wescott et al., 1991). Nothing indicates that the fault zone developed a simple displacement pattern after linkage (following the models in Figure 11b). This implies that regions of the fault remained persistently stronger than other areas (e.g., Figure 11c). The fault zone cannot have undergone slip repeatedly along its entire length; if it had, the early displacement pattern of transverse folds would have been eliminated from the later displacement pattern. The persistence of the transverse folds suggests that only fault segments or patches of faults have slipped at any one time (e.g., Cowie and Scholz, 1992; Burgmann et al., 1994).

Some rifts, particularly those having a long history of activity (10–30 m.y.), show a variety of changes in boundary-fault geometry. In East Africa, depocenter relocation was relatively abrupt, and no detectable progressive changes occurred in depocenter location. Instead, a new depocenter is marked by an abrupt shift in depocenter location (Morley 1999a) (Figures 7, 11). Thus, the impression from such data is that faults propagate very rapidly, and most basin formation occurs after a phase of rapid propagation. Such an example is the Lokichar fault, which is a composite of two faults of different ages. The northern segment of the fault is characterized by the development of a late Miocene–Pliocene half-graben depocenter (North Lokichar basin); the southern segment formed a half graben of Paleogene–middle Miocene age. The result is a transverse anticline in the center of the Lokichar fault that separates the two depocenters (Figure 12). Note the difference in the formation of the transverse anticline

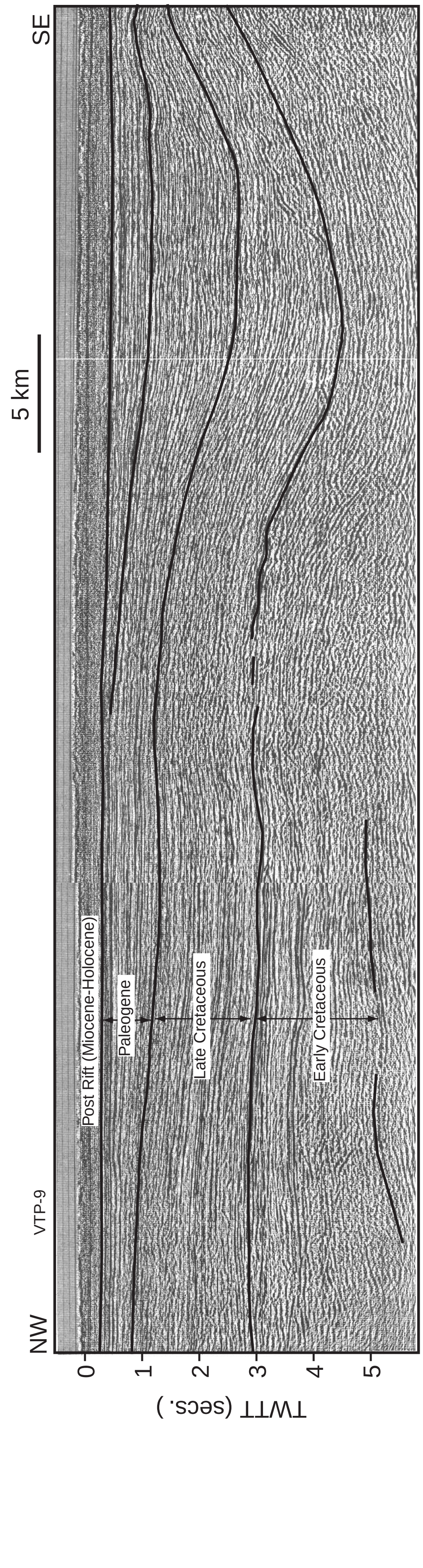


Figure 14. Strike seismic line parallel to the Lagh Bogal fault, Anza graben, illustrating a marked shift to the south of the Paleogene depocenter with respect to earlier, more uniformly distributed Cretaceous units. The Paleogene depocenter represents localization of fault slip to the southeast as fault activity dies out.

along the Lupa fault of the Rukwa rift and the Lokichar fault. In the Rukwa rift, the anticline is formed in sedimentary rocks of the same age on each flank of the anticline, whereas the transverse anticline in the center of the Lokichar fault separates fault segments of different ages (the northern anticline flank is composed of younger sedimentary rocks than the southern flank). Hence, the Rukwa transverse anticline appears to be related to the mechanics of the faulted zone, whereas the Lokichar transverse anticline is a result of reorganizations of fault zones within a long-lived and evolving rift system.

Reorganizations of boundary fault geometries typically are associated with the development of angular unconformities within the synrift section. In particular, they are prominent where secondary faults associated with the activity of a boundary fault are switched off approximately synchronously with the cessation of activity along the boundary fault. This leaves abandoned minor faults and tilted fault blocks below an unconformity. The example shown in Figure 13 is a dip seismic line across the North Lokichar basin. The unconformity marks the cessation of the earlier (Paleogene–middle Miocene) minor fault activity followed by westward tilting of the basin toward the northern segment of the Lokichar fault when it became active in the late Miocene–Pliocene. This event marks a major reorganization of boundary-fault activity. On a shorter time scale (thousands to tens of thousands of years), minor-fault activity can be pulsed and synchronous with pulsed boundary-fault activity (Morley et al., 2000).

Long-lived (>10 m.y.) boundary faults commonly display displacement patterns that vary over time from one part of the fault to another. For example, the Lupa fault in the Anza graben shows considerable change in the location of Cretaceous and Paleogene depocenters (Figure 14). A similar evolution can be

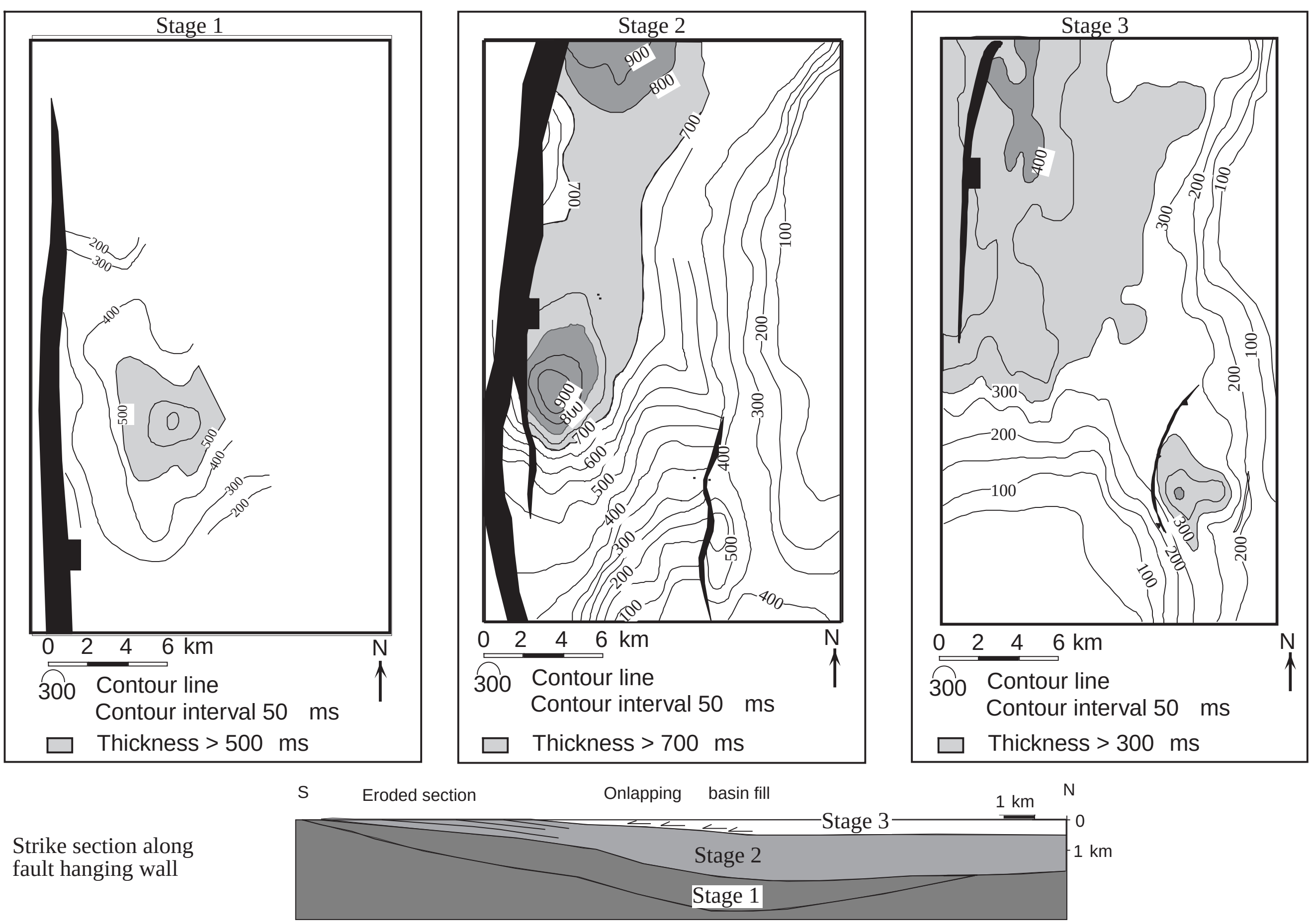


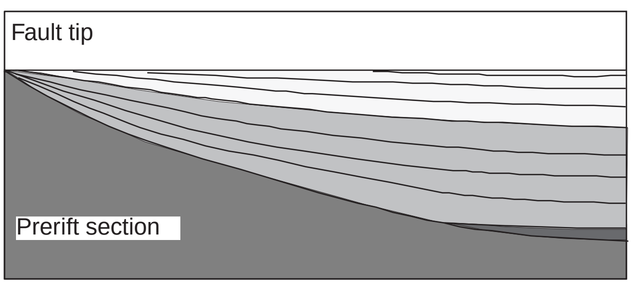
Figure 15. Evolution of a Tertiary rift-boundary fault from southeastern Asia. Isopach maps for three different stages of the synrift section illustrate that the early depocenter lay in the southern area of the fault. Later displacement occurred over a longer region and finally retreated to the northern segment of the fault. The duration of fault activity is about 15 m.y.

seen in a boundary fault from a rift in southeastern Asia (Figure 15). The early displacement on the fault is concentrated in the south; as the fault grows, displacement moves northward, and the southern tip first becomes abandoned, then eroded. It was not a progressive relocation of displacement. The jump appears to have been rapid, because the second fault-related depocenter is characterized by onlap of the basin fill onto the earlier synrift section (Figure 15). Had displacement shifted progressively northward, there would be no onlap and marked shift in depocenter location. Subsequently, the abandoned, southern, fault-segment hanging wall was partially eroded.

FINAL STAGES OF FAULT ACTIVITY

Fault activity may end abruptly because of a major external event, such as tectonic inversion. In boundary faults that die out more gradually as other faults in a basin become active, displacement appears to become progressively more restricted to one segment, either in the center of the fault or at one end (Figures 14, 15, 16).

SCHEMATIC STRIKE SECTION



STAGES OF FAULT GROWTH

(1) Initial fault linkage and basin formation (extent of basin fill does not necessarily indicate extent of fault)

(2) Nonpropagating development of fault

(3) Contraction of active area on fault, commonly uplift and erosion near fault tip

Figure 16. Summary diagram illustrating a typical basin filling pattern for a strike section in the hanging wall immediately adjacent to a boundary fault. The early rift-basin fill related to the fault linkage stage tends to be very thin or absent. Most of the basin fill thins toward the fault tips; there is little indication of basin propagation accompanying fault propagation. As the fault dies out, the latest synrift fill occupies an increasingly narrow part of the fault.

Commonly, faults display late activity restricted to the center of the fault, whereas the hanging-wall section at the fault tips is eroded. The cause of the erosion can be attributed to several mechanisms: (1) late inversion on parts of the fault; (2) the active part of the fault creates a depression so that abandoned parts of the fault passively form flanks that are elevated relative to the depression; consequently, they may become a region of net erosion; and (3) isostatic uplift.

One problem associated with interpreting rift sequences is assessing whether the basin fill is associated with an active fault or just infilling the abandoned, sediment-starved inactive hanging wall (Prosser, 1993; Rattey and Hayward, 1993). This has become a complex argument, and solutions have been sought in the lithology and dip-section geometry of the hanging-wall fill (Prosser, 1993). Erosion of a subaerial, inactive footwall block may cause boundary-fault fan deposits to extend farther into the footwall block than during the active phase (e.g., Blair, 1987). Similar sequences, however, may be caused by the footwall block in an active rift changing from a submerged to a subaerial setting (Figure 17). Hence, the sedimentary criteria for identifying the timing of fault activity remain somewhat equivocal. Another means of determining whether a sequence infilled an inactive- or active-fault hanging-wall depocenter is to check the strike geometry of the infill. If the fault was active, the beds should display typical curved, infilling patterns, thinning toward the fault tips (Figure 17b), whereas in an inactive fault, the depression would be infilled by subhorizontal, onlapping strata (Figure 17a). The fault-strike parallel geometry of strata infilling a decreasingly active fault produces a narrowing upward of the basin fill (Figures 14, 17b), whereas infill of an inactive hanging wall produces a broadening-upward geometry (Figure 17a). Such patterns are applicable to simple extensional basins but cannot be considered useful criteria where significant changes in paleostress direction occurred (for example, changing from extension to strike-slip deformation).

The final-stage fault geometry is also dependent upon whether the setting is submarine or continental. Many Late Jurassic extensional faults in the North Sea ceased activity in deep-marine conditions. Subsequent postrift sedimentation began by infilling the inactive submarine-rift topography (e.g., Rattey and Hayward, 1993). In a continental setting, the transition to the postrift phase is different. Decreasing fault activity tends to result in rapidly shallowing lacustrine environ-

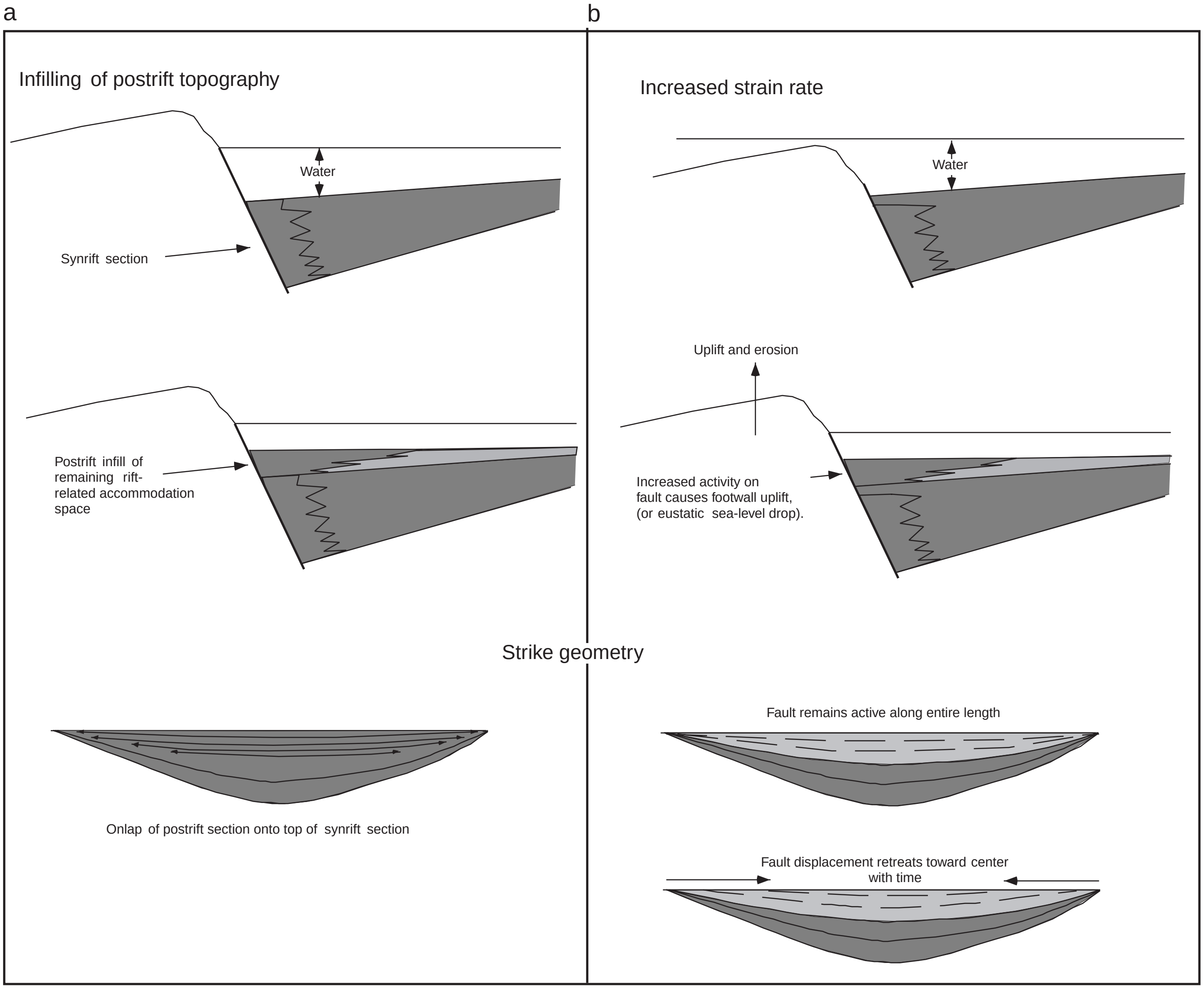


Figure 17. Schematic illustration of the problems associated with discerning late synrift infill of a half graben (b) vs. postrift infill (a) of abandoned rift topography. On a dip section alone, it may be difficult to determine whether a coarse clastic pulse is caused by erosion of inactive rift topography (a) or by active uplift that caused the topography to rise from a submarine to a subaerial position (b). The pattern of basin fill parallel to fault strike might help to distinguish synrift from postrift fill.

ments and a transition from lacustrine shales to prograding fluviodeltaic deposits (e.g., Lambiase and Bosworth, 1995). This infilling history, plus the subaerial exposure of the footwall block, results in boundary faults associated with continental sedimentation that tend to have a somewhat different late-infill geometry compared with marine settings. In particular, boundary faults that die out in marine conditions may have considerable accommodation space remaining that is filled by postkinematic sedimentary rocks for millions of years after fault activity has ceased. Faults in continental rifts are unlikely to have any accommodation space preserved for long after they cease to be active, because of erosion of the footwall geometry and rapid infilling of the hanging wall.

CONCLUSIONS

Fault development can be divided into several stages: (1) early linkage, (2) evolution of established boundary faults, and (3) termination of fault activity. Early linkage patterns involve joining of en echelon faults at synthetic transfer zones. Differences between examples of early fault linkage center around the timing of basin development with respect to the establishment of a major boundary fault from the linkage of minor faults. One end member of the continuum is very rapid fault linkage to form a boundary fault prior to significant basin development (Morley, 1999a) (<1 m.y.?). The other end member is the creation of an extensive synformal basin by numerous minor faults and the subsequent development of boundary faults after several million years of early rift sedimentation (e.g., Nukhul Formation, Gulf of Suez [Patton et al., 1994; Gupta et al., 1998]; Usangu Flats, Kenya, Figure 4). In between comes the development of a basin during boundary fault propagation (Schlische, 1991).

For many faults, the postlinkage displacement pattern is simple; maximum displacement occurs in approximately the center of the fault. In some faults, however, displacement remains persistently less at sticking points along the fault (at the sites of former synthetic transfer zones) and creates transverse anticlines. The subsequent evolution of boundary faults can produce marked changes in depocenter location. Amalgamation of two faults of different ages also can generate transverse anticlines. Such changes tend to accompany major structural reorganizations in rifts, and intra-synrift unconformities are one product of this process.

Faults can die out in a variety of ways. Cessation of activity can be abrupt, or it can be a more prolonged decrease in activity that results in displacement restricted to increasingly narrower areas of the fault. The pattern created by hanging-wall fills during the late synrift stage as seen on strike sections may help determine whether sedimentation infilled abandoned rift topography during the postrift stage or during the late synrift deposition (Figure 17).

The early stages of fault development can be understood in terms of propagating fractures and rock mechanics (e.g., Burgmann et al., 1994; Cowie, 1998; Gupta et al., 1998). Once a large boundary fault is established, its subsequent evolution is dependent upon a complex mixture of influences. Fault activity and geometry can be affected by strain-hardening mechanics within the lithosphere (e.g., Kusznir and Park, 1987), by volcanic activity, by changes in regional stresses, and by reorganizations caused by fault propagation and upper crustal strain hardening or softening.

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