New palaeogeographic and lake-level reconstructions of Lake Tanganyika: implications for tectonic, climatic and biological evolution in a rift lake

A. S. Cohen,* K.-E. Lezzar,† J.-J. Tiercelin† and M. Soreghan‡

*Department of Geosciences, University of Arizona, Tucson, AZ 85721, USA
†CNRS URA 1278 'Domaines Océaniques', Groupe Riftogénese Est-Afrique, Université de Bretagne Occidentale, B. P. 809, 29285 Brest Cedex, France
‡School of Geology and Geophysics, University of Oklahoma, Norman, OK 73019, USA

ABSTRACT
Palaeogeographic and lake-level reconstructions provide powerful tools for evaluating competing scenarios of biotic, climatic and geological evolution within a lake basin. Here we present new reconstructions for the northern Lake Tanganyika subbasins, based on reflection seismic, core and outcrop data. Reflection seismic radiocarbon method (RSRM) age estimates provide a chronological model for these reconstructions, against which yet to be obtained age dates based on core samples can be compared. A complex history of hydrological connections and changes in shoreline configuration in northern Lake Tanganyika has resulted from a combination of volcanic doming, border fault evolution and climatically induced lake-level fluctuations. The stratigraphic expression of lake-level highstands and lowstands in Lake Tanganyika is predictable and cyclic (referred to here as Capart Cycles), but in a pattern that differs profoundly from the classic Van Houten cycles of some Newark Supergroup rift basins. This difference results from the extraordinary topographic relief of the Western Rift lakes, coupled with the rapidity of large-scale lake-level fluctuations. Major unconformity surfaces associated with Lake Tanganyika lowstands may have corresponded with high-latitude glacial maxima throughout much of the mid- to late Pleistocene.

Rocky shorelines along the eastern side of the present-day Ubwari Peninsula (Zaire) appear to have had a much more continuous existence as littoral rock habitats than similar areas along the north-western coastline of the lake (adjacent to the Uvira Border Fault System), which in turn are older than the rocky shorelines of the north-east coast of Burundi. This model of palaeogeographic history will be of great help to biologists trying to clarify the evolution of endemic invertebrates and fish in the northern basin of Lake Tanganyika.

INTRODUCTION
Palaeogeographic and lake-level reconstructions have proven to be immensely valuable tools in visualizing lake histories and evaluating alternative hypotheses of palaeohydrology and palaeoclimate (e.g. Teller, 1985; Morrison, 1991). Because of the important role that many large, tectonically formed lakes have for biological evolution, such reconstructions have considerable potential to resolve important debates among evolutionary biologists as well (e.g. Owen et al., 1990). However, the use of lacustrine reconstructions in this application has been much more limited. In this paper we present a series of palaeogeographic reconstructions for Northern Lake Tanganyika with important implications for the tectonic, climatic and biological evolution of this rift basin. Lake Tanganyika is the largest, and probably the oldest extant lake of the East African Rift System (Fig. 1A). The palaeogeographic synthesis of Northern Lake Tanganyika presented here is based on several prior reflection seismic surveys and coring studies (Tiercelin & Mondeguer, 1991; Bouroullec et al., 1991; Cohen et al., 1993a), also discussed in detail in the companion paper to this article (Lezzar et al., 1996), as well as on outcrop and oil wells studies from the region (Attou, 1988).

METHODS
Seismic stratigraphy and age estimations
The data and interpretations that we present here are based primarily on results from Project CASIMIR's 1992
seismic reflection survey, which employed a 300-joule CENTIPEDE Sparker system as a seismic source. This system (developed at the Renard Centre of Marine Geology, University of Ghent, Belgium) allows an acoustical penetration of \( \approx 300 \text{ ms (twtt)} \), an excellent scale for the interpretation of modern rift–lacustrine stratigraphic sequences. Nine seismic lines, totalling 250 km, were shot in the North Tanganyika Basin during the 1992 CASIMIR survey. Figure 1B shows the key lines (Sparker Line S2 and Project PROBE Multichannel Line 16) used for the chronological and sequence stratigraphic interpretations proposed in this paper. Other seismic line locations and technical specifications for this survey, as well as detailed discussions of seismic facies observations, are given in a companion paper by Lezzar et al. (1996).

Sparker system data were supplemented in this study by data from earlier seismic (multichannel and subbottom echosounding) and piston coring studies by Projects PROBE and GEORIFT (Rosendahl et al., 1988; Tiercelin et al., 1989; Tiercelin & Mondeguer, 1991; Boursollec et al., 1991).
et al., 1991), by comparison with exploration well data from the Rusizi River Delta (Attou, 1988) and our personal observations of the limited number of outcrops exposed around the north end of the lake.

In the absence of dated core material older than $\pm 35$ ka from Lake Tanganyika, age estimates for events inferred from seismic data can only be generated through sediment accumulation rate extrapolations. Such estimates serve an important function in developing hypotheses about rates and controls on geological, climatic or evolutionary events, or corroborating independently developed chronologies. They are also falsifiable hypotheses, in the sense that multiply derived estimates of the same ‘event’ should yield approximately the same value. However, they must be clearly differentiated from radio metrically constrained age dates, since they depend on large extrapolations. In order to estimate sequence boundary ages we have used a modification of the reflection seismic – radiocarbon method (RSRM, developed previously to estimate the age of the oldest synrift sediments in Lake Tanganyika (Cohen et al., 1993a). This method can be used to generate age estimates for sedimentary horizons in stratigraphic packages with high degrees of completeness. The sumps of hydrologically closed basins in deep half graben lakes have been shown elsewhere often to display these types of highly predictable accumulation rates, with rate ‘constancy’ at the $10^5–10^7$-year resolution, measured over $10^2–10^3$-year time scales (Sadler, 1981; Williams, 1993; Reynolds, 1994; Moutoua, 1995; Olsen et al., 1996). We used $\delta^{13}$C-derived mean sediment accumulation rates for late Pleistocene unconsolidated sediments from four coring sites (Fig. 1B) to infer the chronology of decompacted stratigraphic sequences from the same localities, and deposited under similar tectonostratigraphic conditions (proximity and activity of adjacent faults). In a modification of earlier RSRM methods, rate estimates varied between sites and over time at sites, depending on the tectonostratigraphic setting existent at each site at the time. Details of the RSRM interpretive methodology and phase terminology are presented in Lezzar et al. (1996).

**Palaeogeographic reconstructions**

Our primary method for making the palaeogeographic sketch maps presented here was to trace lowstand lake-level estimates based on onlapping relationships observed in reflection seismic profiles (Projects PROBE, GEORIFT and CASIMIR) and, to the extent possible, relate these to the existing onshore exploration well data and the modern geomorphology of surrounding watersheds. The clear correlations between erosional truncation surfaces evident in three structurally distinct northern subbasins along seismic line S4 formed the basis for subsequent extrapolations of the extent of the lake(s) at various time intervals (Lezzar et al., 1996, figs 3 and 10).

Given the limited number of high-resolution seismic lines in the northern basin, we have extrapolated tural, stratigraphic and geomorphic trends to the south only as far as the extent of major faults or other structural features that are imaged by sparker seismic lines. This corresponds with the location of the East and West Ubwari Faults, on both sides of the Ubwari Peninsula at the south end of Burton’s Bay. For these reasons we caution the reader against attributing more precision than is warranted to shoreline locations.

Our facies interpretations are based on correlations between near-surface high-resolution profiles and litho stratigraphic data from cores, discussed more fully in Bouroullec et al. (1991, 1992) and Tiercelin et al. (1992, 1994). Facies interpretations for the more recent II and A seismic sequences are based on the facies maps of Bouroullec et al. (1992). The estimation of lake highstand surface elevations is problematic in Lake Tanganyika because of the absence of widespread highstand deposits, particularly above modern lake level. However, given the extremely steep-sided shoreline of Lake Tanganyika, an error in the estimate of lake highstand surface elevation does not translate into an unacceptably large error in shoreline longitude/latitude position. For example, assuming a mean topography similar to that of the present, a 100-m error in lake surface elevation would only introduce an error of c. 3.1 km ($\pm 38$) in shoreline position and a $30\%$ error in lake area. Also the complete absence of Lake Tanganyika deposits outside of the confines of the current structural basin, or even for that matter higher than 55 m above present lake level (the highest reported undeformed lake bed outcrops), constrains the maximal highstand extent of the lake considerably (Ilunga, 1988). Finally, two oil exploration wells drilled by AMOCO during the mid-1980s on the Rusizi River plain (north end of Lake Tanganyika) provide some control on the age and duration of flooding events for the lake outside the confines of the modern lake area (Attou, 1988). It seems likely that the spillway elevation for the lake has been close to its modern elevation for much of the lake’s Late Tertiary/Quaternary history and we have used this assumption in delimiting highstand lake margins.

We inferred palaeoshoreline substrates from a number of clues. First, if basement uplifts are associated with lake margin border faults we assumed that the shoreline was rocky. This assumption holds true today along all northern Lake Tanganyika shorelines, in the sense that either rock or mixed rock/sand coastlines are present in these areas (Soreghan & Cohen, 1996). This assumption would only likely be violated in the case of antecedent drainages that cross-cut major rift-bounding structures. In Lake Tanganyika, the Malagarasi River inflow – Lukuga River outlet system is the only strong candidate for a river system entering and leaving the lake that is antecedent to the formation of the rift (Coulter, 1991). In the case of subsequent drainages, their associated sandy beaches are mostly portions of small-scale ($1 \text{ km}^2$ or less) fan deltas (Soreghan & Cohen, 1996).

Conversely, we have assumed that all shoaling margins
of half-grabens or structural platform shorelines were made up of soft substrates. Given the size and nearshore wave energy around a large lake such as Tanganyika, this would be winnowed to a sandy shoreline in all but the most protected areas. Again this assumption would be correct today, with the exception of carbonate-cemented beachrock outcrops and a few isolated and small rocky outcrops along the central Burundi (Resha area) and northern Zaire coasts. Finally, we assume that significant marshlands existed around all major active axial drainage deltas, as is the case today in most rift lakes regardless of lake stand condition (Butzer, 1971; Cohen, 1982; Tiercelin et al., 1992).

For purposes of clarity we have put the locations of all the major structural elements discussed below (major faults, geomorphic features and half-graben structures) only on Fig. 1B.

**PALAEOGEOGRAPHIC RECONSTRUCTIONS**

**Initial synrift infills – Phase RBM (RSRM estimated age $\approx 7.4$ Ma to $\approx 1.1$ Ma)** (Figs 2A–C & Fig. 3)

The oldest sediments of the Northern Tanganyika Basin were deposited in the South Rusizi half-graben over a precrift surface (the Nyanja Event), which appears to have been very broad and flat (Rosendahl et al., 1988). RSRM age estimates for this event from both Cohen et al. (1993a) and Lezzar et al. (1996) suggest these sediments are $\approx 7.4$ Ma. Structural highs developed in response to footwall uplift of the West Ubwari Fault, defining the eastern limit of the present basin. The South Rusizi half-graben was the major depocentre of the North Basin during this period, with major sediment accumulation along the western side of the active West Ubwari Fault (central segment north of Cape Banza) (Fig. 1B and 2A). The principal sediment source for this basin appears to have been from an axial, 'proto-Rusizi' drainage to the north, lateral drainages from the western shoaling margin of the basin, based on stratral geometries imaged in PROBE lines 84–200 and 14 (Rosendahl et al., 1988), and probably from lateral drainages from the West Ubwari Fault escarpment. A lake extended from the northern part of the nascent North Rusizi half-graben almost as far south as modern Cape Banza. The northern extent of the lake is delimited by the presence of late Miocene marginal lacustrine deposits in the Buringa I and Rusizi I wells (Attou, 1988).

At the northern end of the North Rusizi half-graben, tectonic activity began at this time along a nascent Uvira Border Fault System. This may be related to the initiation of volcanic doming (the 'Kivu–Rusizi Local Volcanic Dome') that began to develop in the Kivu and Rusizi Basins at about this time (Ebinger, 1989a,b; Coussement, 1995), corresponding to the first stage ($\approx 7.8$ to $\approx 5$ Ma) of the second cycle of doming activity (Bellon & Poulet, 1980; Kampunzu et al., 1983; Pasteels et al., 1989). This initiation of tectonic activity in the northern Lake Tanganyika region may be coincident with the start of subsidence in northern Lake Malawi (the deposition of the Nyasa-Baobab Sequence (estimated at $\approx 8.6$ to $\approx 2.5$ Ma) (Schulte et al., 1989; Ebinger et al., 1993)). Project PROBE profiles 83-06A and 84-228 (Rosendahl et al., 1986; Morley, 1988) suggest that there was little or no activity along the Bujumbura Border Fault System (i.e. the eastern side of the North Rusizi half-graben) at this time. A topographic high appears to have existed along the nascent Uvira Border Fault System, delimiting sediment transport towards the south into a major depocentre.

Further south, in the Kigoma region, Versfelt (1988) and Rosendahl et al. (1988) have suggested that the early, post-Nyanja event history of the lake was characterized by the presence of a broad lake, with little or no activity along the major, modern border faults ('a water-covered cratonic basin developed on a peneplain (Nyanja Event)' according to Rosendahl et al., 1988, p. 26). During the RBM Phase, the West Ubwari Fault became tectonically active long after the deposition of these initial post-Nyanja Event sediments. RSRM estimates suggest this occurred $\approx 4.9$ Ma. A major depocentre developed at this time in the hangingwall of the West Ubwari Fault in the subsiding South Rusizi half-graben (Fig. 2B). The Ubwari horst began to develop as a major structural feature at this time, along with the two transverse zones: the Rusizi accommodation zone and the Kaboge Fault Zone. By this time the lake had expanded slightly in area, extending further to the east (as indicated by sedimentation over the nascent western Ubwari horst) than in the immediate post-Nyanja Event period (for cross-sections in the areas described see figs 3 and 4 in Lezzar et al., 1996).

Apparently the Ubwari/North Kigoma structural areas were high and underwent continued erosion or nendposition during the early history of subsidence in the South Rusizi half-graben. At the north end of the modern lake, subsidence probably accelerated along the Uvira Border Fault System, resulting in a nascent North Rusizi half-graben that underwent rapid sedimentary infilling, most likely from sources to the north (Fig 2B,C). RSRM estimates suggest this occurred between $\approx 4.9$ Ma and $\approx 3.6$ Ma.

Subsequently, activity along the Uvira Border Fault System expanded considerably to the south along with subsidence within the North Rusizi half-graben. Figure 2C illustrates the state of the basin at $\approx 3.6$ Ma (RSRM estimate). Sediment accumulation within the North Rusizi half-graben was concentrated along its rapidly subsiding western margin, with some accumulation also in the modern Rusizi delta region. Much of the sediment discharged from the axial ‘proto-Rusizi’ River appears to have also been ponded against the north-west side of the West Ubwari Fault. An accommodation zone linking the North Rusizi and South Rusizi half-
Fig. 2. (A to N) Palaeogeographic/palaeotectonic sketch maps of the northern Lake Tanganyika region from the early Nyanga Event to the present. Maps are based on a combination of seismic, outcrop, well and geomorphic data. Lake boundary placements are approximate, guided by lapout boundaries and the position of major structures known to be active during each time interval. Fault name abbreviations are the same as for Fig. 1B. The southern limit of the mapped area represents the furthest extent of structures that cross-cut seismic lines for which we have chronological control data. bpll = elevation of lake surface below present lake level. RSRM = reflection seismic radiocarbon method.

Onset of regional erosion and focused rifting – Phase F–E (RSRM estimated age 1.1 to 0.39 Ma) (Figs 2D–G & 3)  

A major change occurred in the northern end of the Lake Tanganyika basin with the onset of focused rifting and the development of associated regional erosion surfaces (Fig. 2D). We estimate this occurred about 1.1 Ma, corresponding to the initiation of Phase F–E of Lezzar et al. (1996). This surface, variously referred to as the Rusizi Sequence Boundary 1 (RSB1), the Banza Sequence Boundary 1 (BSB1) or the Kigoma–Makara Sequence Boundary (KMSB), depending on location, resulted from a combination of a major relative lake level decline (~650–700 m below present lake level, bpll, based on seismic reflection geometries), as well as extensive uplift along the developing East and West Ubwari Faults and the Uvira Border Fault System (Morley, 1988; Rosendahl et al., 1988; Scholz & Rosendahl, 1988; Lezzar et al., 1996). Rosendahl et al. (1988) showed that after the formation of the KMSB, rifting within the Kigoma Province became focused, and its major border faults and modern basin asymmetry developed. Deposition of the early Phase F–E lacustrine sediments (Lezzar et al., 1996) was restricted at this time to small and probably saline terminal lakes in the depocentres west of the West Ubwari Fault and east of the East Ubwari Fault. Activity on the Kivu–Rusizi Local Volcanic Dome had resumed by about 1.9 Ma, the second stage of the second cycle of doming by this time, with a major depocentre forming in the deep basin east of the Ubwari Border Fault System. Major activity on the East and West Ubwari Faults at this time created a separation between the South Rusizi half-graben and the North Kigoma half-graben, referred to as the higher Ubwari horst (Fig. 2D).  

Differeniatating climatic from tectonic causes for the early Pleistocene contraction of lakes in the Tanganyika basins is difficult, if not impossible given our present state of knowledge. It is possible that the reduction in lake area began at ~2.5–2 Ma as a response to the late Pliocene regional aridification in East and Central Africa (de Heinzelin, 1955; Bonnefille, 1976; Cerling & Hay, 1986; Cerling et al., 1988; Pickford et al., 1991; deMenocal, 1995). This is also consistent with a significant change in lithologies in the Rusizi Plain wells, from Pleistocene lacustrine deposits to Pleistocene fluvial deposits (probably separated by an unconformity) (Atzou, 1988). Unpublished seismic data acquired by AMOCO over the Rusizi plain indicates the presence of a pronounced angular unconformity in this area. This may reflect ongoing tectonism during this low-lake period. Under this scenario, low stand and desiccation would have persisted in the northern Lake Tanganyika region for a considerable part of the latest Pliocene and early Pleistocene. Rosendahl et al. (1988) have independently suggested that the deep scour evident in some areas on the KMSB surface argues for a very long interval of erosion. deMenocal (1995) has suggested that increased aridity occurred in Africa at about 1 Ma, based on dust variability records. It is clear that lake level increased in the basin at some time during this interval, although the precise timing of this event, as well as its magnitude, is not well constrained. Shortly after the lowstand event that we estimate occurred ~1.1 Ma, sedimentation first resumed in the actively subsiding South Rusizi half-graben, but it was not until much later (RSRM estimate ~670 ka) that sedimentation resumed on the eastern North Kigoma half-graben margin.  

By the middle of Phase F–E, at around 670 ka by RSRM estimate (Fig. 2E), lake levels had risen substantially (though the magnitude of this rise is unknown). However, the Ubwari horst continued to separate the northern and southern terminal lakes (Lezzar et al., 1996). Motion along the strike-slip Kaboge Fault Zone (KFZ) appears to have occurred at this time as indicated by the faulted, dome-shaped morphology of the F and E Sequence reflectors (fig. 7A in Lezzar et al., 1996). Erosion occurred throughout most of the North Rusizi half-graben at this time. Subsequently, the deposition of Phase F–E sediments across the Ubwari horst indicates that a single lake overtopped this feature. The state of knowledge of the basin at ~550 ka (RSRM estimate) is depicted in Fig. 2F. Based on their seismic reflection characteristics, Sequence F deposits on the Ubwari horst appear to comprise stratified, coarse- to medium-grained detrital deposits that were derived axially from either the north or south ends of the Rusizi half-grabens. These deposits progressively filled the subsiding Magara slope and onlap onto the eastern, gently dipping flank of the Batanza slope, indicating the occurrence of minor uplift of the Magara shoal and continuous normal faulting along the East Ubwari Fault (Banza Shoal).  

Much of the North Rusizi half-graben had also flooded by this time, with a major depocentre forming in the deep basin east of the Ubwari Border Fault System. During this time, motion along the strike-slip Kaboge Fault Zone appears to have also accelerated, whereas subsidence in the South Rusizi half-graben, west of the West Ubwari Fault, appears to have slowed. Sedimentation in this same region also slowed at this time as a consequence of sediment trapping to the north. No structural or seismic data are available to constrain the timing of motion along the Bujumbura Border Fault System (Fig. III), or whether its formation is associated...
Palaeogeography of Lake Tanganyika

Fig. 2. (continued)
with the ancient West Ubwari Fault or the more recent Cape Magara Fault. Deposition during the latter part of Phase F–E (Fig. 2G) occurred throughout almost the entire extent of the modern northern Lake Tanganyika (except for Burton’s Bay and the extreme north-west). Rapid subsidence continued in the North Rusizi half-graben and very slow subsidence continued in the South Rusizi half-graben. Motion slowed on the West Ubwari Fault at this time, but began along the Cape Banana Fault (CBF), on the west side of the modern Ubwari Peninsula (Figs 1B and 2G).

**Sedimentation in a mature rift basin and the development of the modern northern Tanganyikan subbasins – Phases D to A (RSRM estimated age 390 KA to present) (Figs 2H–N & 3)**

Sedimentation during this major phase of northern Lake Tanganyika history has been modulated by a series of large, but declining-amplitude lake-level fluctuations. A detailed discussion of these deposits and their seismic facies characteristics is presented in the companion paper to this paper (Lezzar et al., 1996). The frequency and stratigraphic regularity of these fluctuations, in comparison with earlier ones, strongly suggests that they are climatically driven, although there is clearly a strong tectonic overprint to the morphology of the lake throughout this time interval. Accompanying these lake-level fluctuations are cyclical patterns of sedimentation (discussed more fully in the ‘Facies expression of lake-level fluctuations’ section).

At the end of Phase F–E renewed lake lowstand conditions, represented by the (d) erosional period, resulted in a major reduction in the area of Lake Tanganyika (Fig. 2H). Reconstruction of the (d) boundary morphology suggests that lake level declined to ≥350 m bpl (RSRM age estimate 390–360 Ka). Local uplift of the northern part of the North Rusizi half-graben suggests that renewed activity on the Kivu–Rusizi Volcanic Dome occurred at about this time, or possibly slightly earlier (Pouret, 1976; Bellon & Poulet, 1980; Ebinger, 1988a). This doming may have contributed to the erosion on the (d) sequence boundary at the north end of the study area. Significant horizontal and vertical motion along the Kaboge Fault Zone and both the East and the buried West Ubwari Faults (Figs 1B and 2H), along with subsequent erosion, resulted in considerable topographic irregularity on the (d) sequence boundary (figs 3 and 7A in Lezzar et al., 1996). Relative fault motions between the buried Magara shoal and the Kaboge Dome resulted in the formation of the incipient Baraka furrow. This, along with the extension of the Cape Banana Fault to the south, initiated the formation of the present-day Burton’s Bay.

The return of lake highstand conditions represented by Sequence D (Fig. 2I) was accomplished by a slowing or cessation of subsidence in the South Rusizi half-graben. RSRM estimates suggest this occurred between ∼360 and 290 ka. Burton’s Bay continued to develop, extending further south as fault activity expanded on the Cape Banana Fault and the southern part of the Uvira Border Fault System. The Baraka furrow acted as a major conduit for sediments shed off the western side of the developing Burton’s Bay, although axial sources from further south in the developing Burton’s Bay also contributed sediments to this region. These basin fill sediments are expressed seismically as chaotic to weakly stratified unit boundary reflectors (Fig. 4), with medium-amplitude reflectors displaying a hyperbolic character (figs 3A and 7A in Lezzar et al., 1996). They infill irregular, erosional topographic lows on the prior (d) lowstand surface. We infer that they comprise coarse-grained alluvial and clival deposits. Later, this style of sedimentation gave way to the extensive deposition of a very thick sequence of sheet-drape deposits. Seismically, these sheet-drape deposits comprise laterally extensive bundles of parallel to subparallel, high-amplitude reflectors, interlayered with thin chaotic to subtransparent layers (Fig. 4). We infer these deposits to comprise predominantly pelagic oozes and hemipelagic muds (figs 7A and 9B in Lezzar et al., 1996). In the north-eastern part of the lake, subsidence began along the Cape Magara Fault. Strike-slip movement is evident at this time along the Kaboge Fault Zone, as indicated by the smooth dome shape (positive flower structure) of Sequence D reflectors in the South Rusizi half-graben.

Several major sediment sources that generated deep-water lacustrine fans were active at this time in the north; the Proto-Rusizi drainage (providing sediments derived from the upper Rusizi Valley), lateral drainages from the nascent Cape Magara Fault region and lateral drainages on the north-west margin of the modern Burton’s Bay. We recognize these fans seismically by their chaotic internal structure and lens-shaped geometry, extending over kilometres to tens of kilometres. Upper Sequence D sediments on the Ubwari horst and in the North Kigoma half-graben consist of two seismic facies: dominant, chaotic units that are thin and continuous, and that we interpret as gravity flow deposits (presumed to be predominantly distal turbidites), and subsidiary, thicker but more localized, chaotic units, that we also interpret as gravity flow deposits (presumably proximal turbidites). The proximal turbidites deposited on the buried western side of the horst were probably derived from the incipient Burton’s Bay (see rivers in Fig 1B) and the eastern Ruzibazi River, near Cape Magara. Thinner, proximal turbidites in the Kigoma half-graben were derived from lateral drainages of the eastern rift escarpment (mainly the Ruzibazi and Murembwe Rivers). The distal turbidites in both the Ubwari horst and the North Kigoma half-graben areas probably originated from a northern axial source. The Cape Magara Fault appears to have acted to divert the growth of a northerly sourced axial fan towards the south-west. Elsewhere throughout the
Fig. 2. (continued).
Fig. 2. (continued).

Lake low stand (b)
RSLM Estimate: 190-170 Ka
(> 250 m bgl)

Lake low stand (a)
RSLM Estimate: 40-35 Ka
(> 160 m bgl)

For present-day lake basin morphology see Fig. 1B
Unconformity (c) marks a renewal of lake lowstand conditions within the North Tanganyika Basin (∼350 m bpl) (Fig 2J). RSRM age estimates suggest this occurred ∼290–260 ka. At various times during this interval, a northern, slightly deeper lake along the Uvira Border Fault System may have periodically been separated from the main lake across a very shallow region at the south end of the South Rusizi half-graben. The lake floor in the South Rusizi area appears to have been relatively flat following the end of active subsidence, presaging the morphology of the modern Bujumbura subbasin. Emergent conditions (dry or swampy areas) probably existed throughout much of the subbasin, with shallow lake conditions (∼25 m maximum water depth) developed within a nascent Magara–Banza depression perched on top of the Ubwari horst. Continuous subsidence of this depression resulted in the collapse of the central part of the Uvira horst (along the active Cape Banza Fault), which in turn promoted the formation of the Capart Channel on the buried Magara shoal. A submerged rocky shoal may have crossed the lake between modern Cape Banza and the Kabezi Fault on the east side of the lake at this time.

Minor but widespread erosion of Sequence D sediments occurred during the (c) period on the gentle slope forming the Kigoma half-graben. Lake level reconstructions for the North Kigoma half-graben (NKHG) suggest a maximum water depth of ∼30 m along the East Ubwari Fault. Simultaneously the eastern side of the NKHG was subaerially exposed and undergoing active erosion.

Lake level rise during Lezzar et al.’s Phase C (Fig. 2K) resulted in the flooding of essentially the entire modern northern part of Lake Tanganyika, with the possible exception of the southernmost portion of Burton’s Bay, which may not have formed by this time. RSRM age estimates suggest this occurred between ∼260 and 190 ka. In the early part of the transgression, basin-fill deposits infilled topographic irregularities on the (c) erosion surface in the nascent, present-day Bujumbura subbasin. These sediments may have been derived from either northern (proto-Rusizi) or southern (proto-Kasandjala and Mutembala) axial drainages (Figs 1B and 2K). As the transgression proceeded, these gave way to the deposition of extensive deep-water lacustrine fan deposits that developed near the major sediment discharge point. A deep lacustrine fan also developed at the outlet of the Ilaraka River. Sheet drape deposits, comprising fine-grained detrital clays and organic oozes, overlies the fan deposits. Most of the clastic material in these well-stratified deposits probably was derived from the Capeart Channel on the buried Magara shoal. In the Capart Channel, Sequence C sediments appear to be dominated by coarser detrital sediments (based on their seismic character). In both the Magara–Banza depression (on the Uvira horst) and the North Kigoma half-graben, Sequence C comprises only sheet drape deposits. A fan-shaped geometry within Sequence C along the western side of the Magara–Banza depression suggests continuous.

Fig. 4. Idealized Capart Cycles from the northern Lake Tanganyika basin. Facies interpretations relating core and high-resolution seismic data, average cycle and facies thicknesses are based on Tiercelin & Mondeguer (1991), Baltzer (1991), Bouroullec et al. (1991, 1992) and Lezzar et al. (1996). The seismic line blowups illustrating the seismic facies characteristics for each part of the cycle are from CASIMIR Sparker Line S₂, in the Magara–Banza depression (Ubwari horst and the southern Bujumbura subbasin–South Rusizi half-graben (see Lezzar et al., 1996, fig. 7A).

The subsidence along the Cape Banza Fault whereas, on its continued subsidence of the Kigoma half-graben was coupled with the uplift of the Banza shoal and probable eastward motion along the Kabezi Fault (figs 3 and 7 in Lezzar et al., 1996). At the end of Phase C, renewed regression resulted in a lake level fall to approximately 250 m bpl. This value is close to the previously proposed for the (A∞) and (W) discontinuities deepening of a small graben that subsequently became the nascent, present-day Rumonge Channel (Tiercelin et al., 1989; Mondeguer, 1991; Tiercelin & Mondeguer, 1991; Bouroullec et al., 1992). The reconstruction of the palaeotopography at the lowest stand suggests the existence of a flat basin floor in the Bujumbura subbasin, and more pronounced relief in the subsiding Magara–Banza depression (see Lezzar et al., 1996, fig. 10C). We estimate a maximum water depth of about 75–100 m in both of these areas. A large part of Burton’s Bay was probably emergent and/or occupied by swampy areas, an idea consistent with earlier observations of Tiercelin & Mondeguer (1991). To the east, continued subsidence of the Kigoma half-graben was coupled with the uplift of the Banza shoal and probable rollover of the basin margin, represented by the shallow Rumonge platform (see Lezzar et al., 1996, figs 7C and 10, evolution of Rumonge platform). These rollover structures are comparable with those described in Lake Malawi by Ebinger et al. (1993). The lower part of the slope probably subsided as a consequence of movement along the N-20°-trending fault, antithetic to the main East Ubwari Fault. This subsidence also resulted in the deepening of a small graben that subsequently became the nascent, present-day Rumonge Channel (≈ 150 m maximum water depth) during the (b) lowstand, with an eastern shoreline closely coinciding with the shelf break.

Sequence B was deposited under a renewed episode of transgression (Fig. 2M). RSRM age estimates suggest this occurred between ≈170 and 40 ka. The earliest stages of lake level rise were associated with extensive erosion, particularly in the Capart and Rumonge channels. The Capart Channel was probably fluvially incised by an axial drainage flowing northward from the emergent Burton’s Bay. Erosion in the Rumonge channel was
Palaeogeography of Lake Tanganyika


directly related to coarse detrital inputs from the lateral Ruzibazi River, which was diverted southward at this time by the very active Cape Magara Fault (Figs 1B and 2M).

Deposition during Phase B in the Bujumbura subbasin began with detrital inputs, and coarse-grained basin fill (Fig. 4; fig. 7A in Lezzar et al., 1996). Above this unit, a coarse clastic lens developed on the relatively flat bottom of the axial part of the subbasin. This lens-shaped unit is interpreted as a deep lacustrine fan, formed by coarse-grained gravity flows at the outlet of the lateral Baraka River in Burton's Bay. Above and on both sides of this fan, the upper part of Sequence B is characterized by sheet drape sedimentation during highstand conditions. This same sedimentary succession is duplicated in the Magara–Banza depression (Fig. 4; fig. 7A in Lezzar et al., 1996). Coarse deposits mainly accumulated on the western flank of the depression as a large lens that pinches out towards the east, probably as a consequence of a reduction in subsidence rates on the Magara slope and continuous uplift of the Banza slope.

In the Capart channel, the base of Sequence B comprises a narrow (1 km wide) coarse-grained, lens-shaped body that probably resulted from detrital inputs from the axial Kasandjala and Mutembala Rivers at the southern end of Burton’s Bay. These deposits are overlain in the channel axis by a unit of alternating autochthonous organic sediments and thin detrital units.

In the deepest part of the Rumonge subbasin, Sequence B comprises a thick deposit forming the primary infill of the Rumonge channel (Fig. 4) (fig. 7A in Lezzar et al., 1996). The base of this deposit consists of a relatively flat and thin, apparently coarse-grained detrital layer. Above this unit is a wide, lens-shaped deposit that we infer to be coarse grained, which probably corresponds to a deep lacustrine fan formed at the outlet of the Ruzibazi River. Overlying this are small lenses of apparently coarse-grained material, with cut-and-fill channel features. These channels are interbedded with thin stratified layers that we interpret as distal turbidites.

On the Rumonge slope and platform, Sequence B consists of a thin, coarse-grained layer overlain by a unit of prograding clinoforms and aggrading strata. The clinoforms are probably associated with a nearby river delta (C. A. Scholz, personal communication, 1995), possibly the lateral Murembwe River (see Lezzar et al., 1996, fig. 7C). Pleistocene hightand terraces in the Rusizi River Plain located at 830 m a.s.l. (i.e. 55 m above present lake level) probably formed during this period, suggesting that the lake was somewhat more extensive (particularly to the north and in Burton’s Bay) than it is today. However, no Pleistocene lacustrine deposits occur in either of the AMOCO wells, suggesting that when transgressions reached beyond the modern lake boundaries, they did so only briefly, or were poorly preserved.

Our youngest reconstruction, Fig. 2N, is for the (a) unconformity. The RSRM age estimate for this boundary is ~40–35 ka. This lake level fall (~160 m bpll) was insufficient to divide the lake into separate water bodies, but did result in the near-total desiccation of Burton’s Bay, and probably the Rumonge platform and north-east Lake Tanganyika. The (a) sequence boundary can be correlated to the (A) boundary defined in the Southern Mpulungu subbasin (Tiercelin et al., 1989, Mondeguer, 1991). The ~160 m lake level fall estimate for the northern lake is consistent with a ~150 m estimate for the Southern Mpulungu subbasin. The basin floor topography reconstructed at the time of erosional event (a) appears to be nearly identical to the present morphology, except for the Banza sublacustrine shoal, which was probably somewhat lower than at present (fig. 10 in Lezzar et al., 1996). We estimate a maximum water depth of ~220 m to ~130 m for the subbasins crossed by the sharp line S2. Two subsequent subsequence boundaries, termed (e) and (J) by Bouroullec et al. (1991), have also been observed in both seismic and short core data. These boundaries correspond to lake level falls dated radiometrically at ~25 ka and ~18 ka, respectively. These surfaces are not figured here, since they were primarily driven by changes in the lake’s water budget, and by this time all of the major modern tectonic elements of the modern lake were in place. Thus after the formation of the (a) boundary lake-level changes more or less have followed modern bathymetry.

FACIES EXPRESSION OF LAKE-LEVEL FLUCTUATIONS

In addition to the marked low lake stand erosional surfaces evident in the northern Lake Tanganyika reflection seismic records, seismic facies analysis allows us to infer some characteristics of the vertical stacking of sediments within each stratigraphic sequence (Bouroullec et al., 1991; Lezzar et al., 1996). These sequences themselves appear to be strongly cyclical, an inference that receives support from the uppermost ‘A’ sequence, which has been partially to completely penetrated by gravity, Mackereth and piston cores (Tiercelin et al., 1988, 1989; Baltzer, 1991; Tiercelin & Mondeguer, 1991).

We refer to these cycles as Capart Cycles (Fig. 4), in honour of André Capart, who first recognized from echosounder profiles the probability that Lake Tanganyika had undergone significant fluctuations in water level through the Pleistocene (Capart, 1952). Reflection seismic profiles and seismic facies interpretations for these cycles are also shown in Fig. 4 and in Lezzar et al. (1996, figs 7 and 8). Capart Cycles consist of a basal erosional surface, overlain by a variable thickness (0 to several metres) of coarse-grained siliciclastic deposits (gravels and mud balls) with very localized distributions. These ‘basin fill’ units are expressed seismically by chaotic or weakly stratified/hyperbolic reflectors of medium amplitude and with hyperbolic sequential surfaces. These presumed coarse clastics both infill and overtop localized depressions on the underlying erosion surfaces and appear to have formed under relatively low...
lakes, at the beginning of a lake transgressive phase. The ‘basin fill’ units are typically overlain by the deposits of ‘deep lacustrine fans’, lens-shaped bodies, typically with chaotic internal structure and hyperbolic sequential surfaces. They are interpreted to represent stacks of metres-thick packages of high-energy, coarse-grained gravity flow deposits, generated by successive floods of lateral or axial drainages, and similar in organization to those described from the Early Cretaceous Reconeaou Basin rift-lacustrine turbidites (Medeiros & Ponte, 1981; Magnavita & Da Silva, 1995). In total these ‘basin fill’ stacks average between 20 and 40 m thick in each cycle. The upper half to two-thirds of each cycle (20–40 m) consists of parallel to subparallel, high-amplitude reflectors that alternate with thin, chaotic or subtransparent layers. These ‘sheet drape’ units are interpreted to be the rising to lake highstand mudtrcks that were deposited contemporaneously over vast areas of the lake floor. In sequence A, where they are penetrated by cores, they are dominantly laminated to massive diatomaceous silts (‘oox’).

Cyclical patterns of deposition have been described previously from other rift lake basins but at a scale and with lithofacies packages that are very different from what we observe in the Capart Cycles of Lake Tanganyika. Most notably, the Van Houten Cycles of the Newark Supergroup basins (Olsen, 1986, 1990) also reflect repeated lake-level fluctuations. Van Houten and Capart Cycles differ significantly in their overall thickness (>10 m for a typical Van Houten Cycle, vs. 50–100 m for a typical Capart Cycle), and probably their duration. Additionally, the strongly developed and laterally continuous lake-lowstand, fine-grained palaeosols do not have a parallel in the Capart Cycles, at least for the lowstand portions observed in cores to date. Instead lowstands are expressed as profound erosion surfaces overlain by relatively coarse-grained deposits. Although palaeosols almost certainly developed on exposed highs and interfluves during Lake Tanganyika lowstands, they apparently were subject to subsequent erosion and therefore much lower preservation potential than in the case of Van Houten Cycle palaeosols. In some respects the Capart Cycles resemble the deposits of the more humid, southern basins of the Newark Supergroup, the ‘Richmond-Type Facies Complexes’ (Olsen, 1990), although the thickness of the Capart Cycles appears to be greater than those typical of the Richmond Basin. The differences between how major lake-level fluctuations are expressed in palaeosoles that generate Van Houten Cycles (e.g. the Newark and Hartford Basins) (Olsen, 1986) vs. those that produce Capart Cycles may result from two factors: (1) differences in temporal scale (we infer that Capart Cycles are of considerably longer duration than Van Houten Cycles) or (2) differences in topographic relief (and, secondarily, relative subsidence rate) between the two types of lakes. We consider the former explanation unlikely, since Van Houten Cycles are themselves bundled into longer-duration cycles. These longer-duration cycles probably approximate Capart Cycles in timing, but nevertheless retain the facies elements unique to Van Houten Cycles (i.e. the proportions of facies vary but not the facies suite), rather than taking on characteristics similar to Capart Cycles. The second explanation for differences in cyclicity seems more likely. Lake Tanganyika is (and apparently has been) an extremely steep-sided and very deep lake. Lake-level fall under such conditions leads to very little lateral facies migration where slopes are steep, but potentially extensive erosion. Conversely, the Newark Basin has been suggested to have been a relatively broad topographic basin with little topographic relief (P. Olsen, personal communication, 1989). Additionally, the Newark Basin lakes were probably shallow and relatively low-energy environments (D. Reynolds, personal communication, 1996). Under such conditions extensive palaeosols might be developed during low lake stands without the pervasive deep erosional surfaces typical of Lakes Tanganyika and Malawi (Scholle & Rosendahl, 1988; Bourrouilh et al., 1991). Field observations and computer modelling experiments on the Miocene Rubielos de Mora (RDM) Basin, Spain, support this conclusion (Anadón et al., 1988, 1991; Cohen et al., 1993b). Van Houten cycles occur in the western portion of the RDM basin, which has been interpreted to have been the low-relief, low-subsidence portion of this half-graben lake. This Van Houten type cyclicity disappears in the eastern portion of the RDM palaeolake, which appears to have been a much steeper sided and rapidly subsiding portion of the lake basin. Facies patterns in the eastern part of the RDM palaeolake are similar but not identical to the Capart Cycles of northern Lake Tanganyika.

REGIONAL CLIMATIC AND TECTONOVOLCANIC IMPLICATIONS OF THIS STUDY

Reflection seismic radiocarbon method (RSMR) age estimates of lake-level fluctuations provide intriguing hints that lake-level lowstands in Lake Tanganyika may have corresponded with high-latitude glacial maxima throughout much of the Quaternary. Abundant evidence exists that correlates the most recent Late Pleistocene, lake-level lowstands in Africa with high-latitude glacial conditions during O stage 2 (e.g. Street-Perrott & Harrison, 1985), although the causal relationship (or lack thereof) between the two events has been controversial (deMenocal et al., 1993, and references therein; deMenocal, 1995). However, there have been few attempts to estimate directly the timing of lacustrine highstands and lowstands in Africa during the Quaternary, prior to ±40 ka, from the lakes themselves. A considerable body of evidence suggests that African climate change during the Quaternary has been strongly cyclical, and some of this evidence relates indirectly to lake-level fluctuations. For example, deMenocal et al. (1993) have shown that the period of freshwater Meliosira
diatom valves deflated from dry lake beds and blown as aeolian dust into the deep sea shows a strong 23–19-ka cyclicity at ODP core site 663, off the west coast of Africa, whereas aeolian dust and phyolith records from the same core are strongly cyclical at 100- and 41-ka intervals. Evidence from the Arabian Sea also shows strong Milankovitch-band cyclicity, with dominant frequencies varying between onshore and offshore coring localities (Clemens et al., 1993; Prell et al., 1992; Anderson & Prell, 1993). Modelling results suggest that eastern Africa would be expected to respond more strongly to northern European ice cover, whereas western Africa would respond more directly to North Atlantic sea surface temperatures (de Menocal & Rind, 1993). Lake Tanganyika, lying as it does on the boundary between these climatic domains, is well situated for examining their interactions in regulating central African climate.

In Fig. 3 we summarize our recently developed sequence boundary chronology for the northern basin of Lake Tanganyika, along with our RSRM age hypotheses for these boundaries, and our interpretation of the probable causal mechanism driving each major erosional event. The distinction between early (pre 6.4 Ma) tectonic causes and more recent climatic causes of major unconformities must be interpreted cautiously, given the better seismic resolution of events in the upper portion of the lake's sediment fill. Tectono-volcanic events are almost certainly responsible for many of the major events recorded in the Miocene to mid-Pleistocene stratigraphic record of northern Lake Tanganyika. It is conceivable that this distinction is simply an artefact of the penetration capabilities of the sparker system, and that improved seismic resolution at depth would reveal a pattern similar to that observed in the D–A Sequences. However, we think this is probably not the case. The profound stratigraphic heterogeneity between basins prior to Sequence D and the similarities after Sequence D argue for earlier tectono-volcanic controls as opposed to later climatic ones. Furthermore, Project PROBE seismic line 16 (which penetrated to basement across nearly the same transect as CASIMIR Sparker Line S3) (Fig. 1B) clearly shows a distinction between a lower package (prior to our D sequence) of reflectors, with an overall fan-shaped morphology, separated from an upper (D–A sequence) package of comparatively flat-lying reflectors (figs 2 and 3 in Lezzar et al., 1996). These lines of evidence, of course, do not eliminate the possibility that climatically controlled lake-level fluctuations occurred prior to the deposition of the D Sequence, but only that if such events did occur, their stratigraphic signature was completely overshadowed by major tectono-volcanic events.

Lake-level forcing mechanisms, related to the lake's hydrological budget, are probably responsible for the pronounced cyclicity of the stratigraphic record since the deposition of the D Sequence (which we estimate at the past 400 ka). The widespread nature of the four major and two minor unconformities that we observe (labelled (d), (c), (b), (a), (x) and (y), respectively) provides strong evidence that these events are the consequence of a drop in the lake's surface elevation. The concordance of RSRM age estimates for each of these boundaries between several subbasins (table 1 in Lezzar et al., 1996) is also consistent with this hypothesis. Furthermore the (x) and (y) events can be correlated with similar surfaces at the south end of Lake Tanganyika (Tiercelin & Mondegue, 1991). Two plausible explanations can be offered for these lake-level declines. First, they may have resulted from episodes of stream capture for major influent drainages, caused by such factors as fault growth, migration or propagation, stream diversion related to seismic events, volcanic dome or damming or competitive fluvial aggradation (Leeder & Jackson, 1993). This type of lake-level fluctuation control has been documented for other large fault-bounded lakes, for example Walker Lake, Nevada (Bradbury et al., 1989). Second, they may have resulted from climate change associated with relative drops in P/E ratios within the Lake Tanganyika watershed. Of these two possibilities we consider the second much more likely. Judging from the relatively small thicknesses of the deeper basinal sequences correlative with the unconformities, it does not appear that the unconformities represent very long intervals of time. A tectonic model implying 'stream piracy' events therefore would have to invoke repeated episodes of watershed discharge reversal, with the Tanganyika discharge always being of longer duration than the drainage to outside the lake. This scenario is highly improbable. Furthermore, the three most recent of the unconformities can be correlated with widespread lake-level lowstands known from elsewhere in intertropical Africa during the late Pleistocene (see Regional and global climatic events section below).

**Regional tectono-volcanic events**

The timing of Neogene tectonism in the northern Lake Tanganyika area is consistent with other tectonic events occurring throughout the western rift region (Fig. 1A). From the mid- to late Miocene, volcanic eruptions occurred in the Kivu–Rusizi region north of Lake Tanganyika (Bellon & Pouclet, 1980; Kampunzu et al., 1983; Pasteels & Boven, 1989; Pasteels et al., 1989). This period corresponds to the initiation of what would become the Kivu–Rusizi local volcanic dome (Pouclet, 1976; Ebinger, 1989a,b; Coussement, 1995). Simultaneously, in Northern Lake Malawi, faulting and volcanism resulted in a broad asymmetric lake basin (Ebinger et al., 1993). These ages coincide with the minimum RSRM age estimate of 7.4 Ma for the initiation of the Northern Lake Tanganyika Basin (Cohen et al., 1993a; Lezzar et al., 1996). The initiation of rifting and formation of the Lake Kivu Basin and the nearby Rusizi Basin occurred between 7.5 and 4.4 Ma, with a culmination of volcanic activity centred between 6 and 5.5 Ma (Ebinger, 1989a; Pasteels et al., 1989). These events probably correspond to the beginning of the RBM Phase.
Regional and global climatic events

Changes in lake levels in Lake Tanganyika during the Neogene may be related to regional and even global climatic events. From about 7.0 Ma and 5.5 Ma, a sharp decline occurred in global temperature, along with a pronounced lowering of sea level. These events affected Africa profoundly, and resulted in major changes in continental biota associated with cooler temperatures (Van Zinderen Bakker & Mercer, 1986). This cool period was followed by rapid warming after 5.0 Ma and a global rise in sea level (Van Zinderen Bakker & Mercer, 1986; Krantz, 1991; Webb & Harwood, 1991; Harwood & Webb, 1993). An expansion of proto-Lake Tanganyika occurred at about 3.6 Ma, either as a result of increased pre-rift downwarping or alternatively as a consequence of wetter climatic conditions. A short-lived cold episode has been identified worldwide at about 3.6 Ma, followed in the southern hemisphere by warmer conditions which prevailed until 2.6 Ma (Stackebrandt & Kennett, 1975; Mercer, 1983, 1984). In the East African Rift, a major expansion of lacustrine conditions in the Malawi Basin is recorded from the Chlwondo beds during the mid-Pliocene, somewhat prior to 4.0 Ma, based on biostratigraphic correlations with radiometrically dated, fossiliferous sediments in East Africa (Bromage, 1995). Evidence for lake expansion and wetter climates elsewhere in Central and East Africa during the mid-Pliocene also exists (Gautier, 1970; Cerling et al., 1977; Pickford, 1990; Verniers & de Heinzelin, 1990) and indicates a wetter climate may have been responsible for the expansion and freshening of lakes in the Afar region (Djibouti) during the same period. Evidence also exists for increased runoff and precipitation in the Sahara between 3.7 and 3.5 Ma (LeRoy & Dupont, 1993). Between 2.6 and 1.8 Ma, temperature dropped worldwide, associated with the reestablishment of both the Northern and Southern hemisphere ice sheets (Harwood & Webb, 1993). Simultaneous with this late Pliocene climate warming, regional aridification appeared in North, East and Central Africa (de Heinzelin, 1955; Bonnefille, 1976; Cerling & Hay, 1986; Cerling et al., 1988; Pickford et al., 1991; Suc et al., 1993; de Menocal, 1995). In North Africa, aridification appears to have begun intermittently, by 3.2 Ma in the north-west Sahara (LeRoy & Dupont, 1993) and by 2.6 Ma in the eastern Sahara (Suc et al., 1993). Aeolian dust accumulation, indicative of regional aridification, also increases in the ocean basins surrounding Africa starting at about 2.8 Ma (de Menocal, 1995). A cooler and drier climate is evidenced from the Ethiopian Rift uplands at 2.5-2.35 Ma (Bonnefille, 1983) and rainfall decreased dramatically in the Turkana Rift at 2.0-1.8 Ma (Cerling et al., 1977). For the period between 1.9 and 1.0 Ma, core data off the coast of west Africa indicate relatively constant climatic conditions (Stein & Sarthein, 1984). Given these major global and regional climatic changes throughout the late Neogene, some of the main unconformities observed in the RBM Phase deposits in the Lake Tanganyika fill (Rosendahl et al., 1988; Lezzar et al., 1996; RSRM estimated age range 7.4-1.1 Ma) could be climatically driven, even if tectono-volcanic events were of primary importance.

Phase F-E of the Tanganyika basin evolution (RSRM estimated age 1.1 Ma to ~0.39 Ma) corresponded to a period of significant tectonism (Lezzar et al., 1996), and the dominant signature in the stratigraphic record of this period can be resolved using the CENTIPEDE marker system appears to be tectonic. From Phase D to the present, the four most recent of the unconformities, (d) to (a), may be correlated with evidence for climatic shifts known from elsewhere in intertropical Africa during the late Pleistocene. Regionally cool climates and a lowering of sea level from ~250 to 200 ka (Van Zinderen Bakker & Mercer, 1986) may be correlated with the (d) Tanganyikan low lake stand (Fig. 2H). Between ~350 and 300 ka, a long-term climatic warming in Africa has been identified from sediment cores of the Zaire deep-sea fan (Jensen et al., 1984; Gasse et al., 1989). This event may correspond to the D sequence highstand in Lake Tanganyika (Fig. 2I). At ~270 ka, a major dry episode has been defined in the Congo Basin (Gasse et al., 1988), which may correlate with the (c) low lake level (Fig. 2J). Pronounced increases in arid-climate indicators (phytolith concentrations, aeolian dust and diatoms from deflated lake beds) occur in deep-sea sediments off the west coast of Africa between 300 and 260 ka (deMenocal et al., 1993). In nearshore core records from the Arabian Sea, evidence for a weak Asian monsoon also occurs between 300 and 250 ka (Anderson & Prell, 1993). At ~250 ka, high lake stands described in the Lake Magadi–Natron basin of the southern Kenya Rift (Hillaire-Marcel et al., 1986; Sturchio et al., 1993) may correlate with the beginning of the transgressive C Phase in Lake Tanganyika (Fig. 2K). Between ~250 and 180 ka deep-sea evidence from both the East and West African coasts indicates
increases in intensification of the Asian monsoon, an increase in precipitation and probably higher lake levels (based on fewer deflated freshwater diatoms) (Anderson & Prell, 1993; de Menocal et al., 1993). At 180 ka, Gasse et al. (1989) have identified another dry episode from the Congo Basin that can be correlated with the (b) low lake level in Lake Tanganyika (Fig. 2L). Deep-sea evidence from both coasts of Africa support generally more arid conditions between 200 and 140 ka (Anderson & Prell, 1993; deMenocal et al., 1993).

Highstand conditions characterize Phase B in northern Lake Tanganyika, which we estimate occurred between 170 and 40 ka (Fig. 2M). At 145–120 ka, highstand conditions also prevailed in Lake Turkana in the northern Kenya Rift (Butzer et al., 1969, 1972), in the Magadi–Natron basin in the southern Kenya Rift (Hillaire-Marcel et al., 1986) and in the northern Kenya Rift (Sturchio et al., 1993). Phytoplith abundance data from the West African deep-sea record suggest a strong shift in the position of the African tall-grass savannah between 140 and 40 ka (deMenocal et al., 1993). In the Araban Sea, a renewed intensification of the Asian monsoon is indicated starting at about 150 ka, and continuing until 40 ka (Anderson & Prell, 1993). Lézine & Casanova (1991) have argued for a series of humid phases in Africa at 140–118 ka, 105–96 ka, 92–73 ka, 52–44 ka and 12–2 ka, based on pollen and dinocyst records from the eastern Atlantic.

At Lake Malawi the record is more complex. The base of the Songwe Sequence (a major lowstand indicator in central Lake Malawi) has been assigned highly variable age estimates at different locations, between 150 and 44 ka (Scholz & Finney, 1994). This sequence bears intriguing similarities to the combined B–A Sequence of Lake Tanganyika, both in terms of similar seismic facies characteristics and thicknesses at comparable water depths (Scholz & Finney, 1994, Lézine et al., 1996) have suggested that lake-level fluctuations in Lakes Malawi and Malawi may be out of phase with one another during when lake levels were 250 m lake level drop recorded for Lake Malawi between 40 and 28 ka (Scholz & Finney, 1994), although Finney et al. (1996) have suggested that lake-level fluctuations in Lakes Tanganyika and Malawi may be out of phase with one another during the Late Pleistocene.

Following the (a) low lake level, several transgressive–regressive periods have been identified within the most recent phase (Phase A) in Northern Lake Tanganyika’s history, by our estimates from 40 ka to the present. These are variously estimated (RSRM) or dated radio-metrically from 40–35 ka to 23 ka, from 23 ka to 18 ka and from 18 ka to the present (Tiercelin & Mondeguer, 1991; Bourrouilh et al., 1991; Lezzar et al., 1996). This pattern of lake-level fluctuation is similar to that identified for much of intertropical Africa (Street & Grove, 1979; Van Zinderen Bakker, 1982; Bonnefille & Riollet, 1988; Tiercelin et al., 1988; Gasse et al., 1988; Street-Perrot et al., 1989; Vincens, 1989, 1993; Vincens et al., 1993; Chalié, 1993), although a major exception to this regional pattern is known from Lake Malawi, where early Holocene major lake lowstands date from 10 to 6 ka (Johnson & Davis, 1989; Finney & Johnson, 1991). The RSRM age estimates for the (d), (c), (b) and (a) unconformities (Lezzar et al., 1996) bear a striking relationship to the timing of marine oxygen isotope stages 10, 9, 8, 6 and 2, respectively (Pisias & Leinen, 1984) (Fig. 5). Previous workers have demonstrated a correspondence between low lake levels in the tropics/subtropics of East and North Africa, and high-latitude, late Pleistocene glacial events (late Weischelian/Wisconsin, oxygen isotope stage 2) (Street & Grove, 1979; Flook & Nicholson, 1980; Lattman, 1989). Gasse et al. (1996) have further argued that transitions from arid to humid con-

---

**Fig. 5.** RSRM age-estimated lake-level chronology for Lake Tanganyika superimposed over the marine oxygen isotope chronology for the north-western Pacific (Pisias & Leinen, 1984). The highstand elevations for Lake Tanganyika are constrained only by the fact that deposits in exploration wells on the Rusizi plain, near the present-day lake margin, are almost entirely composed of either fluvial or marginal lake deposits. The ‘B’ Sequence is thought to have been deposited when lake levels were > 35 m higher than present, based on a probable correlation with nearshore lacustrine deposits at this elevation in the Rusizi River valley.
ditions are synchronous across a broad swath of both the Saharan and the Sahelian regions of Africa. This correspondence has been interpreted in various ways (deMenocal et al., 1993). Numerous authors have stated that there are strong interconnections between glacial conditions at high latitudes (resulting in global cooling of sea surface temperatures) and an intensification of coastal upwelling along the west coast of Africa, with a contemporaneous weakening or elimination of monsoonal circulation across equatorial Africa (Parkin & Shackleton, 1973; Muller & Suess, 1979; Flish & Nicholson, 1980; Lattman, 1989; Gasse et al., 1990). During the late Pleistocene and Holocene, the intensification of the African monsoon may have been forced by the strengthened northern hemisphere summer insolation that occurs every 23–19 ka as a result of the Earth’s orbital precession (Hillaire-Marcel et al., 1986, Prell & Kutzbach, 1987, Lizine & Casanova, 1991).

The dominant periodicities of climate fluctuations during the late Neogene in Africa may have varied over time. Using deep-sea dust records, deMenocal (1995) has inferred that 23- and 19-ka (orbital precession) cycles were dominant during the period from 4.5 to 2.8 Ma. Between 2.8 and 1.0 Ma, 41-ka cycles dominate the dust record, coinciding with the onset of similar duration cycles at high latitudes. Since 1.0 Ma, however, deMenocal shows that 100-ka cycles have dominated the African dust record, corresponding to the large-amplitude glacial cycles of high latitudes. However, the absence of extensively and continuously dated lacustrine sequences of middle Pleistocene age in Africa has prevented palaeo-limnologists from evaluating this chronology directly.

Our RSRM age estimate results, tied to the geographically discontinuous but better dated evidence from elsewhere in Africa, support the notion that cyclicity at the \( \approx 100\)-ka frequency has in fact been in place for at least the past 0.4 Ma. It is also consistent with the temporal arguments for orbital forcing mechanisms controlling tropical lake levels at an orbital eccentricity time scale (Olsen, 1990), though no shorter duration cyclicity is evident from our data. Elsewhere, Reynolds (1992, 1994) has argued on theoretical grounds and through the generation of synthetic seismic models that a 100-ka Milankovitch eccentricity cyclicity should be identifiable in reflection seismic profiles (depending on seismic source signature) from an equatorial lake like Tanganyika. Our RSRM age estimate results are consistent with this prediction. It is important to note, however, that our results do not provide a positive evidence against higher frequency cyclicity in lake-level fluctuations, simply because such cycles are not evident in a seismic stratigraphic record. Major lake-level excursions that occur over brief intervals of time may go unreflected in the record of major unconformities and sequence geometry in large lakes, or may be unresolvable given the seismic acquisition parameters we have used. Evidence for these types of events may only leave their mark through more subtle sedimentological or palaeoentological evidence that must be deduced from cores.

**IMPLICATIONS FOR THE EVOLUTION AND BIOGEOGRAPHY OF LAKE TANGANYIKA ENDEMIC SPECIES**

The history of Lake Tanganyika fragmentation and creation of patches of littoral rocky substrates along major border faults has important implications for understanding the evolution of endemic organisms in the lake (primarily cichlid fish, ostracods and gastropods), many of which are restricted to these habitats (Coulter, 1991; Michel et al., 1992, SturmBacker & Meyer, 1993). Similarly, the origin of major regions of soft-bottom substrates within the littoral zone can be traced to their teconostrophic setting in the lake and the lake-level conditions prevailing at the time. These soft-bottom benthic habitats house their own endemics and act as potential substrate barriers to dispersal for rock-substrate dwellers. Our palaeogeographic reconstructions indicate repeated episodes of desiccation for the northernmost part of Lake Tanganyika (Fig. 2A–N). Small and probably closed basin lakes occupied the topographic sumps of the evolving structural subbasins in this region, in locations that are not directly predictable from modern bathymetry. As indicated by our palaeoshoreline reconstructions (Fig. 6A–N), the northern shoreline of Lake Tanganyika can be subdivided into a series of discrete segments of rocky or sandy substrate, each with a separate history. These include the following regions, from south to north.

**Ubwari Peninsula, East Coast.** This rocky coast today is formed by the East Ubwari Fault, the western escarpment margin of the North Kigoma half-graben (Fig. III). The East Ubwari Fault has been active since synrift deposition began in this area (by our estimate, about 3.6 Ma in the north Kigoma region). Since the (f) lowstand (by our estimate \( \approx 1.1 \) Ma) the northern part of the East Ubwari Fault appears to have been inundated along the coastline (Figs 2C,D and 6C,D). Thus rocky habitat has persisted in this area for at least this long.

**Burton’s Bay.** The present-day Burton’s Bay formed starting at the time of the (d) sequence boundary (RSRM estimate \( \approx 390–360 \) ka) following the formation of the Baraka furrow, through southwards-directed rift propagation between the southern extension of the Uvira Border Fault System (UBFS) and the Cape Banza Fault (CBF) (Figs 2H,L and 6H,L), resulting in progressively younger maximum ages of the bay toward the south (Lezzar et al., 1996). The bay was probably completely dry during the (a) boundary lake-level lowstand (RSRM estimate \( \approx 40–35 \) ka) (Fig. 6N). The east side of Burton’s Bay comprises a rocky coastline that follows the Cape Banza Fault, and would have formed a steep but dry valley margin following the Ubwari Peninsula during late Pleistocene periods of lake-level lowstand. This area of
Periodically been connected to the central/southern sandy shoreline during extreme lake-level lowstands, Pleistocene, with changes in local volcanic doming activity history cannot be deduced from the CASIMIR data set, half-graben. This coastline has migrated on and off shore has been active periodically since the Nyanja Event, and the South Rusizi half-graben (Fig. 6L), which have probably been intermittently active throughout the history of rifting in the area. However, these faults appear to have undergone a period of quiescence prior to the formation of the (d) boundary (RSRM estimate about 390–360 ka). Resumption of faulting would have renewed the existence of rocky areas during or slightly before Phase C (RSRM estimate 190–190 ka), and Fig. 6L, RSRM estimate 170 ka, respectively the area of the Cape Magara Fault would have been above lake level. Thus environments suitable for rock-dwelling endemics have probably been in continuous existence along the north-eastern coast of Lake Tanganyika since some time after 170 ka (i.e. since the resumption of high lake stand conditions during Sequence B). Regardless of the accuracy of the RSRM estimates, the relative sequence chronology demonstrates that this is a shorter continuous duration of rocky habitat than has existed along the other northern lake rocky coastlines. Rocky corridors connecting Cape Magara (Burundi) to Cape Banza (Zaire) may have been developed briefly during lake level fall at the end of Phase B and again early in Phase A (Fig. 6MN), although this would probably have only lasted a few thousand years.

Burton’s Bay could have been recolonized during the past 35 kyr, presumably from the east side of the Ubwari Peninsula.

The south side of Burton’s Bay forms the southern axial margin of the South Rusizi half-graben. By analogy with present-day conditions around most major axial drainage deltas, this area has probably been a zone of either marshland or valley cutting during lowstand conditions, with intermittent sandy coasts, for its entire history.

Zaire coast north of the Baraka River. This is the western shoaling margin of the South Rusizi half-graben (Fig. 1B). It appears to have been a sandy/muddy coast for most of its history. During the early, active subsidence phase of the South Rusizi half-graben (Fig. 2A–H) it probably sloped steeply offshore, but the offshore gradient became flatter following the end of active subsidence at the (d) boundary (RSRM estimate about 390–360 ka).

North Zaire coast along the Ubwari Border Fault System (UBFS). This coast is formed by the UBFS, the western escarpment margin of the North Rusizi half-graben. It has probably been an area of rocky outcrops for most of the duration of the North Rusizi half-graben, starting at 6–5 Ma (Ebinger et al., 1989a,b; Coussement, 1995). This region was completely above lake level during the (c) boundary low lake stand event (RSRM estimate 290–260 ka, Figs 2J and 6J). Thus its current phase of colonization by lacustrine species must postdate that interval. Since that time it has probably been in continuous existence as a rocky habitat.

The Rusizi River deltaic environment is located at the extreme end of the North Tanganyika Basin. The thick (700–800 m) fluviodeltaic sequence present in wells drilled by AMOCO on the modern Rusizi River delta suggests it has been in continuous existence as an axial drainage throughout the history of Lake Tanganyika, although unconformities in its record suggest a discharge point at various times (Attou, 1988). The delta front has probably been marshy for most of its history. Consequently, the Rusizi represents one of the most permanent, although not particularly wide, barriers between rocky substrates along the lake’s northern coastline.

Burundi sandy coast. This coast, part of the Bujumbura platform (Fig. 1B), developed on the rollover zone of the shoaling margin of the active North Rusizi half-graben. This coastline has migrated on and offshore with changes in lake level and, prior to the middle Pleistocene, with changes in local volcanic doming activity (Kivu-Rusizi local dome) (Figs 2A and 6A) and it has periodically been connected to the central/southern Burundi sandy coasts (most recently during the (b) lake lowstand – RSRM estimate 190–170 ka), when the South Rusizi half-graben formed a separate lake (Fig. 6L).

North Burundi rocky coast. This coast represents the north-eastern escarpment margin of the South Rusizi half-graben, close to the present-day Cape Magara Fault. This sublacustrine high and fault system may have formed as a consequence of uplift/subsidence motion along the Ubwari horst and the South Rusizi half-graben (Fig. 6C,J), which have probably been intermittently active throughout the history of rifting in the area. This coast is formed by the UBFS, the western escarpment margin of the North Rusizi half-graben. It has probably been an area of rocky outcrops for most of the duration of the North Rusizi half-graben, starting at 6–5 Ma (Ebinger et al., 1989a,b; Coussement, 1995). Since the resumption of high lake stand conditions during Sequence B, the area of the Cape Magara Fault would have been above lake level. Thus environments suitable for rock-dwelling endemics have probably been in continuous existence along the north-eastern coast of Lake Tanganyika since some time after 170 ka (i.e. since the resumption of high lake stand conditions during Sequence B).

Central Burundi sandy coast. This coast formed as the shoaling margin of the North Kigoma half-graben (Fig. 1B). Sandy corridors have existed along this coast continuously since the (f) limit for the North Kigoma half-graben (which we estimate to be at least 550 ka), although parts of the coastline have been reduced substantially during lake-level falls. Rocky habitats may have existed sporadically along this coast, particularly during periods of activity along the normal or strike-slip faults on the Rumonge Platform (Figs 1B and 2N).

South Burundi rocky coast. The present-day South Burundi rocky coast is a northern extension of the East Kigoma Border Fault System. Project PROBE data (Rosenblatt et al., 1988) suggests that this fault system has been active periodically since the Nyanja Event, although the precise timing of activity is unknown. Its history cannot be deduced from the CASIMIR data set, except to the extent that this region would have formed a sandy shoreline during extreme lake-level lowstands, when the border fault system in this area would have been exposed on land.
Fig. 6. (A to N) Palaeoshoreline reconstructions for northern Lake Tanganyika for the time frames/intervals discussed in the text (see also Fig. 2A–N). Palaeoshoreline substrate designations are intended to show the dominant substrate types occurring at the time in the nearshore environment. Fault name abbreviations are the same as for Fig. 1B. The width of the substrate band is not significant. bpll = elevation of lake surface below present lake level. RSRM = reflection seismic radiocarbon method.
Fig. 6. (continued)
CONCLUSION

In this study we have outlined a series of working hypotheses about the chronology of tectonic, geographical and climatic events in the northern Lake Tanganyika region. Our conclusions are broadly consistent with what was previously known about regional tectonic and climatic history, giving us considerable confidence that RSRM age estimation provides a useful approach for understanding lake basin histories. As such it may provide a model for application in other lake basins. This study has also produced important testable hypotheses about the timing of events of critical importance for fields as disparate as palaeoclimatology and evolutionary biology. Our challenge now, and the challenge for others interested in the geological history of the African Great Lakes, is to obtain long records through deep drilling programmes that can confirm or refute the hypotheses we present here. In many ways our ideas about chronology are analogous to those derived in planetary geology from crater counting. We will only know the full story when we finally get a chance to visit and sample the rocks.

ACKNOWLEDGEMENTS

Research authorizations were provided by the Ministry of Higher Education and Scientific Research of the Republic of Burundi, the Ministry of Energy and Ministry of Scientific Research and Technology of the Republic of Zaire and the Tanzanian Commission for Research and Technology. The 1992 high-resolution seismic programme in the North Tanganyika Basin was supported by the CASHIR Project (Belgium), Elf-Aquitaine Production (France), the Ministry of Foreign Affairs (France) and INSU-CNRS (France). Thanks to Marc De Batist and the RCMG Team (Renard Centre of Marine Geology, University of Ghent, Belgium) for the sparker data acquisition and processing. Thanks to Hervé Bellon for his great help in constructing the tectono-volcanic correlation sections, and to Bill Wescott and Denise Stone, and AMOCO, for making well logs, samples and unpublished data available to us in the preparation of this paper. Special thanks to Gashagaza Masta Mukwaya, Gaspard Ntakimazi, Laurent Ntahuga and Pontien Ndabaniree for facilitating this research, to Bernadette Coleno for assistance with illustrations, and to Michael Talbot, Chris Scholz, Dave Reynolds and Kathleen Nicoll for their many useful suggestions on ways to improve this paper. This is publication #50 of the International Decade of East African Lakes (IDeAL) programme.

REFERENCES


Boxbull, J. L., Rehult, J. P., Rolet, J., Terekhin, J. &
Palaeogeography of Lake Tanganyika


Vincens, A., Cauqué, F., Bonnierle, R., Guot, J. &


Manuscript received 18 December 1995; revision accepted 4 September 1996.