New seismic stratigraphy and Late Tertiary history of the North Tanganyika Basin, East African Rift system, deduced from multichannel and high-resolution reflection seismic data and piston core evidence

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ABSTRACT

We present here the initial results of a high-resolution (sparker) reflection seismic survey in Northern Lake Tanganyika, East African Rift system. We have combined these results with data from earlier multichannel reflection seismic and 5-kHz echosounding surveys. The combination of the three complementary seismic investigation methods has allowed us to propose a new scenario for the late Miocene to Recent sedimentary evolution of the North Tanganyika Basin. Seismic sequences and regional tectonic information permit us to deduce the palaeotopography at the end of each stratigraphic sequence. The basin history comprises six phases interpreted to be responses to variations in regional tectonism and/or climate. Using the reflection seismic–radiocarbon method (RSRM), the minimum ages for the start of each phase (above each sequence boundary) are estimated to be: ~7.4 Ma, ~1.1 Ma, ~393–363 ka, ~295–262 ka, ~193–169 ka, ~40–35 ka. Corresponding lowstand lake elevations below present lake level for the last five phases are estimated to have been: ~650–700 m, ~350 m, ~350 m, ~250 m and ~160 m, respectively. The latest phase from ~40–35 ka until the present can be subdivided into three subphases separated by two lowstand periods, dated at ~23 ka and ~18 ka. From the late Miocene until the mid Pleistocene, large-scale patterns of sedimentation within the basin were primarily controlled by tectonism. In contrast, from the mid Pleistocene to the present, sedimentation in Lake Tanganyika seems to have responded dramatically to climatic changes as suggested by repeated patterns of lake level fluctuations. During this period, the basin infill history is characterized by the recurrent association of three types of deposits: ‘basin fill’ accumulations; lens-shaped ‘deep lacustrine fans’; and ‘sheet drape’ deposits. The successive low-lake-level fluctuations decreased in intensity with time as a consequence of rapid sedimentary filling under conditions of declining tectonic subsidence. The climate signal has thus been more pronounced in recent sedimentary phases as tectonic effects have waned.
INTRODUCTION

Lake Tanganyika (3°30′–S 8°50′ S, 29°–31°20′ E) lies in the Western Rift Valley of Central Africa, at an altitude of 773 m (Fig. 1A). It is the longest (650 km) lake of the East African Rift system, and with a maximum depth of about 1470 m; this lake is the second deepest lake in the world after Lake Baikal. The climate of the Lake Tanganyika region is semihumid–tropical, with a variable-intensity rainy season lasting 8–9 months, alternating with a pronounced dry season of 3–4 months duration (June–September) (Peguy, 1961). The average annual rainfall is about 1000–1100 mm.

This equatorial lake is divided into two main basins, north and south, separated by the major Kalamie–Mahali bathymetric shoal (Fig. 1A). The basin geometry and stratigraphy were documented by the multifold seismic reflection studies of Project PROBE (Rosendahl et al. 1988). Regional tectonic investigations have shown that the northern end of the North Tanganyika Basin is divided into four discrete half-grabens (Fig. 1B), each containing a thick (>4-km) sequence of rift-related sediments (Degens et al., 1971; Rosendahl et al., 1986; Morley, 1988; Ebinger, 1989b; Tiercelin & Mondeguer, 1991). The synrift infill consists of three first-order seismic stratigraphic sequences that were identified above a well-defined reflector named the Nyanja Event (NE), interpreted to be prerift basement (Burgess et al., 1988). Age estimates using the reflection seismic–radiocarbon method (RSMR), developed by Cohen et al. (1993), suggest that the central portion of Lake Tanganyika began to form between 9 and 12 Ma, whereas the northern and southern ends formed more recently (7–8 Ma, and 2–4 Ma, respectively) (Fig. 1A).

In this paper, we present the first results of a study that combines three seismic reflection data sets in the tectonically complex area of Northern Lake Tanganyika (Fig. 1C). First, we have used the previously collected multichannel seismic data of Project PROBE (Rosendahl et al. 1986, 1988). We have combined these data with our sparkler seismic data (Project CASIMIR), which document with a better resolution the upper parts of the multichannel sections. Finally, we have incorporated results from the 5-kHz echosounding lines of Project GEFORIFT, which provide details that are the uppermost (∼100 m) part of the lake’s stratigraphy (Bourrouilhe et al., 1991, 1992). From correlations with piston core analysis and 5-kHz seismic facies (Mondeguer et al., 1989; Balthzer, 1991; Bourrouilhe et al., 1991; Mondeguer, 1991; Tiercelin et al., 1994), we infer that most Lake Tanganyika sediments are pelagic oozes, with subordinate fine siliclastics (mud and sand) and, less frequently, coarse-grained sands and gravels. Differences in the deposition of this material can be attributed to differences in tectonic–geomorphic settings such as border fault margins, axial and lateral littoral platforms, midlake structural highs and axial-deep basins (Cohen, 1989, 1990; Tiercelin et al., 1992; Soreghan, 1994). All of these complementary data allow us to develop a tectonostratigraphic framework for the northern end of the North Tanganyika Basin and to understand its Late Tertiary history.

TECTONIC AND HYDROLOGICAL SETTING

The North Tanganyika Basin is structurally controlled by the two main fault trends of the East African Rift system, which trend N00°–20° and N130°–140°. The deep basin structure between 3°20′ S and 4°30′ S consists of four individual half-grabens (Fig. 1B,C). These include, from north to south, respectively: (1) the down-to-the-west North Rusizi half-graben; (2) the down-to-the-east South Rusizi half-graben; (3) the North Kigoma; and (4) East Kigoma half-grabens. The Kigoma half-grabens are separated from the Rusizi half-grabens by the N20°-trending Ubwari horst (Burton’s Bay accommodation zone of Rosendahl et al., 1988), on top of which lies the Magara–Banza ‘perched’ depression (Figs 1C and 2). The southern part of this horst comprises the subaerial Ubwari Peninsula, which is bounded by two main fault systems, the East and West Ubwari Faults (EUF and WUF) intersected by N130°–150°-trending faults called the Kabezi Fault (KAF) and the Kaboge Fault Zone (KFZ).

The modern topography of the study area consists of the rectangular-shaped Bujumbura and Rumonge sub-basins, superimposed on the South Rusizi and North Kigoma half-grabens, respectively (Figs 1A and 2). These two sub-basins are separated by the N20°-trending narrow Magara–Banza depression, which results in the intersection of the steeply dipping Banza slope with the gently dipping Magara slope (Fig. 2). The 300-m-deep Bujumbura sub-basin is bounded to the west by the major Uvira Border Fault System (UBFS) and to the east by the Bujumbura Border Fault System (BBFS) and the West Ubwari Fault (WUF) (Fig. 1C). Strong and deep seismic activity with earthquakes as strong as 6.5 on the Richter scale occurs in this region (Zana & Hamaguchi, 1978), particularly along the Uvira Border Fault System where active sublacustrine hydrothermal fields have been discovered (Tiercelin et al., 1993).

The Bujumbura sub-basin terminates on its southwest margin in the shallower Burton’s Bay (mean water depth 150 m), whereas its northern end is occupied by the wide deltaic system of the Rusizi River axial drainage (Fig. 1C). The Rusizi River flows from Lake Kivu and the rift escarpment uplands to Lake Tanganyika and is a major riverine contributor to Lake Tanganyika’s water and salt budget (Hecky, 1978). Ninety per cent of Lake Tanganyika’s water falls directly on the lake’s surface or is carried to it by small, mostly intermittent, streams draining the rift escarpments. Ninety per cent of the lake’s water loss is through evaporation and only 10% is lost through the Lukuga effluent to the Zaire River. Minor, lateral drainages that are controlled by the gradient of the rift margins also enter the Bujumbura sub-
Fig. 1. A. The main basins and physiographical features of the Lake Tanganyika trough. RSRM basal age estimates of Lake Tanganyika are indicated (from Cohen et al., 1993). B. Major structural and geomorphological elements of the North Tanganyika Basin: 1, North Rusizi half-graben; 2, South Rusizi half-graben; 3, North Kigoma half-graben; 4, East Kigoma half-graben (from Rosendahl et al., 1988; Tiercelin & Mondeguer, 1991). C. Simplified bathymetric map. Tectonic and hydrological settings (from Rosendahl et al., 1988; completed): 1, Bujumbura sub-basin; 2, Burton’s Bay; 3, Rumonge sub-basin; 4, Magara–Banza depression. BBFS, Bujumbura Border Fault System; UBFS, Uvira Border Fault System; CBF, Cape Banza Fault; CMF, Cape Magara Fault; EUF, East Ubwari Fault; KAF, Kabezi Fault; KFZ, Kaboge Fault Zone; WUF, West Ubwari Fault.
Fig. 2. Detailed bathymetric map of the study area, plus seismic line and core locations. 1, Deep Bujumbura sub-basin; 2, Baraka canyon/channel; 3, Capart channel; 4, Magara canyon/channel; 5, Magara slope; 6, Banza slope; 7, Rumonge channel; 8, Rumonge platform. For fault names, see Fig. 1C.

basin (Cohen, 1989; Tiercelin et al. 1992; Soreghan, 1994).

The southern end of Burton’s Bay is marked by the axial and lateral drainages formed by the Kasandjala, Mutembala and Baraka Rivers (Fig. 1C). Overlying the Ubwari horst is the 300-m-deep ‘perched’ Magara—Banza depression, formed at the intersection of the N0–20°-trending Magara and Capart canyons and related channel systems. These channels originate in the Ruzibazi River lateral drainage, flowing from the eastern rift escarpment near Cape Magara, and from the axial Kasandjala-Mutembala Rivers in Burton’s Bay, respectively (Fig. 1C). At the north end of the Ubwari Peninsula, bathymetric contours are strongly asymmetrical (Fig. 2).
Fig. 3. Interpretation of seismic lines in a complex structural zone of Northern Lake Tanganyika. A. Multichannel line 16 drawing (Rosendahl et al., 1988; modified). NE, Nyanja Event; Ru, Rusizi Lower Sequence; RSB, Rusizi Sequence Boundary 1; Ru, Rusizi Upper Sequence; Bu, Banza Lower Sequence; Bg, Banza Sequence Boundary 1; Bu, Banza Upper Sequence; M, Makara Sequence; KMSM, Kigoma-Makara Sequence Boundary; K, Kigoma Sequence. For fault names, see Fig. 1C. B. Simplified sparker line S2 drawing. WFB to WFA, South Rusizi half-graben/Bujumbura sub-basin Sequences; MFB to MFA, Ubwari horst/Magara-Banza depression Sequences; RMB - E to RMA, North Kigoma half-graben/Rumonge sub-basin Sequences. For multichannel and sparker seismic sequence correlations, see Fig. 5. ‘Site 2–3’, ‘Site 10’ and ‘Site 14’ are the localities used for our RSRM age estimation (see text for details). Wd, water depth; TW-TT, Two Way-Travel Time.
The East Ubwari Fault (EUF) is well expressed in the present-day topography and coincides with the Banza shoal, which dips to the north (Coussement et al., 1994; Coussement, 1995). To the west, on the Ubwari horst, a N0–20°-trending 'en relais' fault system is formed by the antithetic Cape Banza (CBF) and Cape Magara (CMF) Faults (Fig. 2). These sublacustrine faults with opposing polarity scarps dip northward and southward, respectively, and, as we will discuss later, appear to teconically control the Magara–Banza depression and the Capart channel (Fig. 2).

The 1150-m-deep Rumonge sub-basin is bounded on its eastern side by a N160°-trending margin that is very steeply sloping in the north, but grades into a gently dipping platform towards the south. Its strongly asymmetric western side corresponds with the EUF. The sub-basin structure also controls the N0–20°-trending Rumonge channel that heads in the Ruzibazi and Murembwe Rivers where they flow off the eastern rift escarpment. The Rumonge channel runs south along the eastern flank of the Ubwari Peninsula down to the deepest point (1310 m) of the Kigotna sub-basin (Figs 1B, 1C and 2).

**SEISMIC AND CORING METHODS**

In this paper, we present the initial results of a reflection seismic survey (lines S2–S9) (Fig. 2) conducted in the framework of the Project CASIMIR (1992) in the North Tanganyika Basin. During this survey, the CENTIPede Sparker (300-J) developed at the Renard Centre of Marine Geology at Ghent University, Belgium, was used as a seismic source. It allows an acoustic penetration of about 300 ms two-way travel-time (TW- TT). Single-channel data were recorded digitally for subsequent processing using DELPH 2 and PHENIX VECTOR software.

We have combined results from the CASIMIR survey, illustrated in this work by the most representative sparker lines S2 and S9, with data from earlier multichannel reflection seismic and 5-kHz echosounding surveys carried out in the same area by Project PROBE (lines 16 and 14) (Rosendahl et al., 1988) and Project GEORIFT (lines 55 and 24) (Bourroulec et al., 1991, 1992). In addition we used data from Kullenberg cores (cores SD 2, SD 3, SD 10 and SD 14) collected in the same area (Fig. 2) by the GEORIFT Project, to establish a correlation between the sedimentary and high-resolution seismic facies analysis.

**SEISMIC STRATIGRAPHY RESULTS**

The synrift fill of the North Tanganyika Basin accumulated above the Nyanga Event (NE), a high-amplitude and well-defined group of reflectors present all over Lake Tanganyika, which is interpreted to mark the surface of prerift basement. On multichannel seismic lines 16 and 14 (South Rusizi half-graben), the post-Nyanga Event
sediment infill comprises the Rusizi Lower and Rusizi Upper Sequences (Rt. and Ru), separated by the Rusizi Sequence Boundary 1 (RSB1) (Figs 3A and 4A). On the Ubwari horst and in its base by a seismic discontinuity defined by virtue of reflector terminations and overall facies changes. These sequences are labelled BuF to BuA in the South Rusizi half-graben/Bujumbura sub-basin, MBF to MA for the Ubwari horst/Magara-Banja depression (where MBE is not imaged), and RMF-E to RMA for the North Kigoma half-graben/Rumonge sub-basin (where RMF and RME are indistinguishable). Sequence boundaries are labelled Bu(f) to Bu(a), MB(f) to MB(a), and RM(f) to RM(a), respectively (Figs 3B, 4B and 5). As a group these sequences correspond to the uppermost seismic sequences identified on the multichannel lines, named Ru, Bu and K Sequences (Figs 3A, 4A and 5). The most recent sequence boundaries (b) and (a) imaged in our sparker lines S2 and S3 have been previously identified on the 5-kHz echosounding profiles (Tercelin et al., 1989; Bouroulec et al., 1991) (Fig. 5). Several additional subsequence boundaries above the (a) sequence boundary that are not clearly visible on the sparker lines were imaged on the 5-kHz echosounding lines, and are referred to here as (x) and (y). These boundaries can be used to divide the upper A Sequence into three subsequences, labelled C1 Upper (C1u), C2 and C3 (Fig. 6).

The three most common seismic facies signatures seen throughout the North Tanganyika Basin can be interpreted to have formed as follows.

1 'Basin fill' units: these form the base of each sequence and spread over the underlying, irregularly shaped unconformity. The associated seismic facies are typically chaotic or weakly stratified, with unit boundaries represented by medium-amplitude reflectors with hyperbolic character.

2 'Deep lacustrine fans': these are lens-shaped units with a seismic facies that is typically chaotic.

3 'Sheet drape' units: these form the upper parts of individual sequences. They are expressed by parallel to subparallel, high-amplitude reflectors that alternate with thin chaotic to subtransparent layers. These units may also be represented by another facies type consisting of subtransparent or transparent layers alternating with thin discontinuous, medium-amplitude reflectors.

South Rusizi half-graben/Bujumbura sub-basin: BU Sequences

Sequence BuF. The BuF Sequence is bounded at its base by the poorly defined acoustic discontinuity Bu(f) (Figs 3B and 4B). On line S3, this limit delineates a combined dome-depression feature that extends from the

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Fig. 5. Multichannel and sparker seismic sequence correlations. For sequence names, see Fig. 3.
Zairean slope across the central part of the South Rusizi half-graben (Fig. 3). This feature, named the Kaboge dome, may be related to the positive Kaboge flower structure that is the subsurface expression of the Kaboge Fault Zone (KFZ) (Fig. 2). East of this structure, Sequence BU\(F\) is 130 ms thick, and is acoustically poorly defined, except at its top, where a few toplap terminations are evident and delineate the overlying BU(e) boundary (Fig. 7A).

**Sequence BU\(E\).** This sequence is bounded at its base by the BU(e) discontinuity (Figs 7A and 8A). BU(E) averages 60 ms thick and forms a dome-shaped structure within the KFZ. Its acoustic character consists of alternating subtransparent to chaotic layers, each of which is continuous and 15 ms thick. Internally, the layers contain high-amplitude, parallel reflectors 5 ms thick, that onlap to the west on the BU(e) boundary (Figs 3B and 7A).

**Sequence BU\(D\).** This sequence is bounded at its base by a strongly irregular surface BU(d), onto which the BU\(D\) reflectors clearly onlap and downlap (Fig. 7A). The BU\(D\) Sequence reaches a maximum thickness of 130 ms. At its base it consists of a chaotic unit (mean thickness 35 ms) that overlies the BU(d) surface across the entire basin. This unit reaches a maximum thickness of 50 ms in the axial part of the South Rusizi half-graben, where it forms the first infill of the Baraka furrow on top of the irregular BU(d) surface (Figs 3B and 7A). Above this is a unit characterized by subparallel and high-amplitude reflectors that alternate with chaotic to subtransparent layers. This unit has a maximum thickness of 80 ms on the east side of the Kaboge dome but thins to 10–40 ms towards the western Bujumbura sub-basin. It disappears to the east where it onlaps the BU(d) surface on the west side of the buried Magara shoal (Figs 3B and 7A).

**Sequence BU\(C\).** The transition from the BU\(D\) to the BU\(C\) Sequence generally corresponds to the BU(e) erosional surface in the Bujumbura sub-basin (Figs 3B and 4B). However, internal reflectors of the BU\(C\) Sequence may be concordant with the top of the BU\(D\) Sequence in the central part of the sub-basin. At its base BU\(C\) is characterized by a 15-ms-thick, acoustically chaotic or weakly stratified layer. This layer is present on the Kaboge dome as well as in the depression located between the dome and the western slope of the sub-basin. In this depression, the stratified reflectors onlap the BU(c) discontinuity (Fig. 7A). On top of the Kaboge dome, the 15-ms-thick layer in turn is overlain by a 30-ms-thick by 3-km-wide, flat-bottomed lens with a typical chaotic acoustic character (Fig. 7A). Above this, a 30-ms-thick unit is formed by subtransparent layers that alternate with discontinuous and high-amplitude reflectors. This overlying unit is very continuous, and has a high-amplitude acoustic character. It thickens to as much as 50 ms thick towards the basin centre (Fig. 3B).

**Sequence BU\(B\).** This sequence is bounded at its base by the BU(b) unconformity, defined by virtue of the truncation of internal reflectors of the BU\(C\) Sequence (Fig. 7A). In the depression between the western slope and the west side of the Kaboge dome, the BU\(B\) Sequence is characterized at its base by a thin (20-ms) layer with parallel or discontinuous dipping reflectors that onlap the BU(b) surface. This layer is partially overlain by a 30-ms-
thick by 3-km-wide, flat-bottomed, chaotic lens, that is developed on the Kaboge dome. Above, a 45-ms-thick unit consists of alternating subtransparent layers, 10 ms thick, and highly continuous reflectors whose amplitude increases eastward. The BU_B Sequence corresponds to both the F_1 chaotic Unit and the C_1 Lower half Subsequence (C_{1u}) defined on 5-kHz echosounding lines (Bujumbura sub-basin) by Bouroullec et al. (1991) (Fig. 6).

**Sequence BU_A.** This uppermost sequence (mean thickness 10 ms) was deposited over the BU(u) sequence boundary. Its base comprises a subtransparent unit overlain by a group of high-continuity, high-amplitude reflectors about 5–10 ms thick that can be easily recognized throughout the northern basin (Bouroullec et al., 1991). Above, a subtransparent unit forms the lake bottom (Figs 7A and 8A). In the Baraka channel, the BU_A Sequence contains major erosional truncations at its base and displays a chaotic seismic facies pattern. This succession of units can be correlated to the C1 Upper half (C_{1u}), C_2 and C_3 Subsequences of Bouroullec et al. (1991), which are separated by the (x) and (y) minor unconformities (Figs 6, 7A, 8A and 9).

**Ubwari horst/Magara–Banza depression: MB Sequences**

**Sequence MB_F.** The MB_F Sequence overlies the MB(f) sequence boundary, which is marked by a high-amplitude and high-continuity reflector that forms the highly faulted top of the Ubwari horst (Fig. 3). This boundary corresponds to the BU(f) discontinuity previously described.
at the base of the succession of BU Sequences in the South Rusizi half-graben/Bujumbura sub-basin (Figs 7A and 8A). The mF Sequence is 80 ms thick and is characterized by low-amplitude and low-continuity reflectors alternating with thin chaotic layers. The top of this sequence displays erosional truncations on the buried Magara shoal (Figs 7B and 8B). Sequence reflectors dip towards the central part of the Magara–Banza depression, then rise eastward on the Banza slope where they onlap the MB(f) discontinuity (Fig. 3B).

**Sequence MBd.** In the Ubwari region along sparker line S2, Sequence mBd is not identifiable. The mBd Sequence (80 ms mean thickness) is developed on the irregular MB(d) discontinuity, which is defined by truncation of reflectors at the top of the mF Sequence (Figs 7B and 8B). The acoustic character of Sequence mBd changes laterally from clinoform reflections with alternation of chaotic layers and a few poorly stratified reflectors on the eastern flank of the buried Magara shoal to fan-shaped, chaotic to subtransparent units separated by thin acoustically stratified units composed of alternately low- and high-amplitude reflectors, in the Magara–Banza depression (Figs 3B and 7B).

**Sequence MBc.** The lower boundary MB(c) of the MBc Sequence is typically expressed as an erosional discontinuity in the Capart channel where it truncates internal reflectors of the mBd Sequence (Fig. 7B). Sequence MBc in this channel consists of two parallel, chaotic to sub-transparent layers, 10–20 ms thick, separated by a group of thin (5-ms), medium-continuity and high-amplitude reflectors. In the Magara–Banza depression, the MB(c) sequence boundary is represented on the Magara slope by a vertical facies change from mBd to MB(c) (Fig. 7B) and by an erosional surface (Fig. 8B), and on the Banza slope by an onlapping configuration with respect to the mBd Sequence (Fig. 3B). Sequence MBc here is 90 ms thick and changes to a divergent configuration consisting of a set of high-amplitude reflectors of average continuity, as well as subtransparent to chaotic layers.

**Sequence MBb.** This sequence is bounded at its base by an erosional surface, MB(b), identified in the Capart
Fig. 7C.
channel (Fig. 7B). It corresponds also to an onlap surface at the crest of the Banza slope (Fig. 3B). In the Capart channel, Sequence mB is characterized at its base by a chaotic lens, 30 ms thick and 1 km wide, deposited on the Mb(b) erosional surface along the western side of the channel. Above, a 10-ms-thick stratified unit covers the axial part of the channel and consists of low-amplitude, low-frequency reflectors alternating with thin subtransparent layers. To the east, the base of Sequence mB consists of a small and thin (15-ms) unit which fills the axial part of the Magara–Banza depression (Fig. 7B). This unit is overlain by a 45-ms-thick and 2.5-km-wide chaotic lens which pinches out to the east and west on the Mb(b) surface. On both sides of the lens a 30-ms-thick stratified unit occurs, which consists of subtransparent layers. This unit pinches out by onlapping on the Mb(b) discontinuity on the Banza slope (Figs 3B and 7B).

**Sequence mB.** The Mb(a) lower boundary of Sequence mB (45 ms maximum thickness) forms an erosional discontinuity on both levees of the Capart channel, and an onlap surface on the eastern side of the Magara–Banza depression (Figs 3B and 7B). In the Capart channel, the Mb(a) boundary corresponds to a vertical facies change from the uppermost stratified unit of Sequence mB to a lower stratified to subtransparent unit 40 ms thick in...
Sequence MBA. Above this, a very thin (2–ms), acoustically stratified layer occurs, composed of high-amplitude and high-continuity reflectors, that in turn is overlain by a transparent layer (5–6 ms thick) that forms the modern channel floor. On both sides of the channel, the same sequence configuration is developed, with the exception that a chaotic unit (20–30 ms in maximum thickness) is developed at the base of some irregularities on the MB(a) surface in the axial part of the Magara–Banza depression (Fig. 7B). The first two units of MBA (chaotic, then subtransparent) can be correlated to the C1U Subsequence defined by Bourouleuc et al. (1991). The two upper layers of MBA (well-stratified, then subtransparent) correspond to the C2 and C3 Subsequences (Figs 6 and 9).

North Kigoma half-graben/Rumonge sub-basin: RM Sequences

Sequence RMFe-E. At the base of line S2 in the North Kigoma half-graben, a high-amplitude reflector is well represented near the East Ubwari Fault (EUF) (Figs 3B and 7C). On the basis of its stratigraphic position and shape, this reflector can be correlated to the KMSB boundary delineated on multichannel seismic lines 16 and 14 (Figs 3A, 4A and 5). Labelled RM(f), it forms the lower limit of a sequence with a poor acoustic definition. Based on its stratigraphic position, shape and thickness, it can be correlated with the two consecutive BbF and BbE Sequences defined in the South Rusizi half-graben (Figs 7A and 8A), the (e) limit not being
apparent in the North Kigoma half-graben (Fig. 7C). We have therefore named this sequence rmF-E.

**Sequence rmD.** This sequence is conformable with the rmF-E Sequence, and is bounded at its base by the discontinuous, high-amplitude RM(d) reflector, interpreted as an erosional surface on the basis of reflector truncation at the top of Sequence rmF-E (Figs 7C and 8C). rmD has a mean thickness of 40 ms, and is characterized by poorly stratified, low-amplitude and low-continuity reflectors alternating with thin chaotic layers.

**Sequence rmC.** This sequence developed on the RM(c) sequence boundary, which cuts the rmD and rmF-E
reflector on the east side of the Rumonge platform (Figs 3B and 7C). To the east, on sparker line S2, RM(c) is displaced by a normal fault with an apparent throw of 50 ms, bounding the eastern side of the Rumonge channel. Sequence rMC is conformable with the underlying sequence and thickens to a maximum of 80 ms towards the East Ubwari Fault. This sequence consists at its base of a thin (15–ms) chaotic layer covering the entire sub-basin, and is overlain by four divergent and chaotic units (15–30 ms in maximum thickness) alternating with 5-ms-thick groups of thin, stratified and discontinuous reflectors. These units are intensely faulted on the Rumonge slope and platform (Figs 7C and 8C). In the Rumonge channel, reflectors of this unit are strongly flexed against the East Ubwari Fault.

**Sequence rMB.** A major erosional surface RM(b) forms the base of the rMB Sequence (Fig. 7C). In the Rumonge channel, RM(b) cuts deeply into the rMC Sequence, as shown by erosional truncations, and forms a flat and horizontal channel floor (Fig. 7C). The rMB Sequence is 100 ms thick in the Rumonge channel, and consists at its base of a 20-ms-thick horizontal, chaotic to subtransparent unit overlain by a deposit of complex fill 80 ms thick. This pattern changes upwards from lens-shaped reflectors to a group of shallow chaotic lenses, mainly present in the channel axis, and interbedded with strongly irregular reflectors (Fig. 7C). Eastward, on the Rumonge slope and platform (Fig. 7C), the rMB Sequence is 90 ms thick and strongly faulted. It consists at its base of a thin (15–ms) chaotic layer that is probably equivalent to the 20-ms-thick chaotic channel-fill unit previously described in the Rumonge channel. This layer dips conformably with the underlying sequence towards the west as it thins, and then disappears to the east on the upper platform. Above, a 75–ms-thick unit is formed by four subtransparent to chaotic subunits separated by three discontinuous, high-amplitude reflectors. Aggradational onflaps are delineated within these units by low-amplitude internal reflectors, thus giving the rMB Sequence a general prograding/aggrading appearance (Fig. 7C). Further north, the rMB Sequence is only 25 ms thick and drapes the steeply sloping Rumonge platform (Figs 2 and 8C).

**Sequence rMA.** This sequence is 25–30 ms thick within the Rumonge channel and about 30 ms on the Rumonge platform (Figs 7C and 8C). Its lower boundary, RM(a), is an erosive surface mainly defined on the Rumonge slope where it cuts the rMB internal units and forms the lake floor (Fig. 7C). In the Rumonge channel, the RM(a) boundary is represented by a vertical facies change from the complex fill facies of the rMB Sequence to a 5-ms-thick flat-bottomed lens with arched, medium-amplitude reflectors at the base of the rMA Sequence. Above, the rMA Sequence consists of two distinctive units separated by a thin layer (2–3 ms thick) with extremely high-amplitude and continuous reflectors. The lower unit, 10 ms thick, shows a lateral facies change from alternating stratified and chaotic layers to the east to subtransparent facies to the west (close to the EUF). The upper unit, 7 ms thick, forms the lake floor. Near the East Ubwari Fault the upper unit is mound shaped and shows a subtransparent seismic facies. It changes to the east to a horizontal unit with both parallel and chaotic facies. On the Rumonge platform, the base of the rMA Sequence consists of a chaotic unit 10 ms thick, overlain by two subtransparent units (each 5–10 ms thick) that are separated by a thin set of high-amplitude, high-continuity reflectors (2–3 ms). This succession of Sequence rMA units can be recognized in both the Rumonge channel and the platform, and is identical to the one identified on lines S2 and S5 in the Magara–Banza depression and the Bujumbura sub-basin (Figs 7A, B and 8A, B). It also corresponds to the succession of C1, C2 and C3 Sub-sequences of Bouroulec et al. (1991) (Figs 6 and 9).

**ESTIMATION OF PALEOLAKE TOPOGRAPHIES AND SEQUENCE BOUNDARY AGES**

For this study, we have deduced the palaeotopographies at the end of deposition of each stratigraphic sequence and estimated the ages of these sequence boundaries along multichannel line 16 and sparker line S2 by combining two complementary methods. First we determined the palaeolake topographies and depositional environments at each sequence boundary based on a seismic facies analysis. This provided a relative-age history of tectonism and deposition of the evolving basin along lines 16 and S2. We then assigned absolute age estimates to the sequence boundaries in these reconstructions, using the reflection seismic–radiocarbon method (RSMR) of Cohen et al. (1993). This method produces an age estimate by extrapolating recent sediment accumulation rates (calculated from piston core data) to the stratigraphic section of interest. The stratigraphic section thickness is calculated from seismic data from the location of the piston core and is then decompacted to allow an age estimate to be made (Fig. 5; Table 1).

**Palaeolake topographic reconstruction and low-lake-level estimation method**

On the basis of their acoustic character, many of the sequence boundaries appear to represent erosional periods related to lake lowstands (the top of an underlying transgressive–regressive sequence). For example, we have used toplap terminations and erosional truncation surfaces as indications of subaerial exposure and nondeposition/erosion. For each sequence boundary we identified the lake lowstand elevations (below the present lake level, bpl) and corresponding basin topography by identifying the points on line S2 where the upper erosion surface of each sequence intersects the lower sequence
boundary, thus characterizing sequence wedges (Figs 3B and 10).

Whenever possible, we have used the elevations of sequence wedges within tectonically quiescent areas (i.e. Bujumbura sub-basin during the BU and BU A Sequences deposition and on Banza slope during the MB A Sequence) (Figs 3B and 10E). We have subsequently extrapolated these elevation estimates for each lake lowstand/erosional period in the area of line S2. In situations where such sequence wedges cannot be observed in tectonically quiescent areas (i.e. Banza slope and shoal), estimating lowstand lake-level elevations is confounded by the difficulty of separating subaerial erosion caused by water level changes from subaqueous or subaerial erosion induced by tectonism. For example, in the palaeo-topographic reconstruction for the (c) lowstand period (Fig. 10C), low-lake-level estimation is deduced from the elevation of the BU(c)/BU(d) and MB(c)/MB(d) intersection points, defined in a relatively tectonically quiescent area on the buried Magara shoal. Between these two intersection points, the absence of the D Sequence probably indicates non-deposition or subaerial erosion. In order to extend this lowstand level estimation to the whole area of line S2, the MB(c)/MB(f) and the RM(c)/RM(d) intersection points identified on the Banza slope and in the eastern side of the North Kigoma half-graben, both considered as tectonically active areas, have to be reverted to the elevation of the intersection points defined westward on the tectonically quiescent area of the Magara shoal.

We have used a similar approach for the (d) lowstand period (Fig. 10B). In the South Rusizi half-graben, which appears to have been tectonically quiescent during this period, significant erosion associated with the (d) lake lowstand excavated the nascent Baraka furrow, thus providing a lake level estimate for the erosional episode (d). On the eastern side of the Ubwarri horst, the intersection point between the MB(d) and MB(f) sequence boundaries is situated on the Banza slope, which underwent uplift throughout the rifting phase, as shown by the fan-shaped morphology of the Magara–Banza Sequences (Figs 3B and 4B). To revert this intersection point, it is necessary to strip the deformation out from this zone. The degree to which subsequent tilting of this uplifted area has affected the MB(d)/MB(f) intersection point is constrained by the elevation of the erosional surface BU(d) in the Baraka furrow (Figs 3B, 7A and 10B).

Age estimation method

In the absence of dated core material from Lake Tanganyika that is more than ~35 ka old, age estimates for events inferred from seismic data can only be generated through sediment accumulation rate extrapolations. We have made age extrapolations in this study using a modification of the RSRM developed for the Lake Tanganyika Basin by Cohen et al. (1993). The original method combined the short-term sediment accumulation rates derived from radiocarbon-dated cores (summarized in Tiercelin & Mondegueur, 1991) with depth-to-basement estimates derived from reflection seismic data (Project PROBE) at or near the same locality to estimate an age to basement (Nyanja Event). Cohen et al. (1993) also examined the assumptions required for their method, and other methods of lake age extrapolation, given the known relation between structure and sedimentation in the East African Rift system. Given our present state of knowledge we consider this a justifiable approach and also see potential benefits in making such extrapolations as a way of generating useful hypotheses that can be tested when long cores eventually do become available.

In this study, we have estimated the age of individual sequence boundaries and the duration of individual depositional sequences within each structural zone of the study area (RL–RIV, BU–BUV and M–K Sequences on multichannel line 16, and F to A Sequences on the sparker line S2) (Fig. 3; Table 1). We have modified the original methodology of Cohen et al. (1993) by allowing for variability in sediment accumulation rates at each study site over time, dependent on varying tectonic conditions (discussed in detail below). In addition to providing a depositional chronology for the lake, these age estimates also permit us to establish a chronology of tectonism along the major faults. Taken together, these results allow us to infer a palaeo-topographic and lake-level history for Northern Lake Tanganyika (Fig. 10).

We used four core localities as the basis for combining short core sedimentation rates with seismic reflection data from the same areas (Figs 2 and 3). These included: core SD 2, located close to the active Cape Magara Fault (CMF); core SD 3, in the central deep Bujumbura sub-basin (South Rusizi half-graben); core SD 10, on the west, steep side of the Capart channel (on the buried Magara shoal); and core SD 14, on the Rumonge platform (North Kigoma half-graben). As shown on Fig. 2, core localities SD 2 and SD 3 are slightly north of line S2.

Fig. 10. A to E. Reconstructed cross-sections along sparker line S2 of successive palaeo-topographies (lake lowstand periods) for the ~1 Ma to ~40–35 ka Northern Lake Tanganyika basin history. The present-day basin morphology is shown in Fig. 3B. Local direction of rift extension is indicated (from Morley, 1988; Coussement, 1995). 1, normal faulting; 2, uplift; 3, subsidence/collapse; 4, rotation zone with no vertical motion; 5, sequence boundary intersection points (wedges), see palaeolake topographic reconstruction section; 6, lowstand lake level (below present lake level, bpl); 7, site = RSRM locality, see age estimation method (Table 1); 8, figures indicate the relative chronology of resumption of sedimentation during the first steps of a transgressive phase.
Table 1 A. Reflection seismic radiocarbon method (RSRM) age estimate calculations, derived from multichannel line 16. For sequence abbreviations, see Figs 3A and 5. B. RSRM age estimate calculations, derived from sparker line S. Tectonostratigraphic conditions experienced throughout the late Miocene to Present Lake Tanganyika Basin history and corresponding sediment accumulation rates are estimated as follows: ‘Open basin’, 0.32 mm y⁻¹; ‘proximal to fault escarpment’, 0.86 mm y⁻¹; ‘steep channel side’, 0.15 mm y⁻¹; ‘platform shallow water’, 0.405 mm y⁻¹; ‘dominance of proximal to fault escarpment’ conditions; ‘’, ‘’; ‘intermediate tectonostratigraphic conditions (for details see age estimation method).

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However, we did not consider this a serious problem for two reasons.
1 Both sparker lines S2 and S3 show the same seismic succession (F to A) in the deep Bujumbura sub-basin (Figs 3B and 4B). Line S2 is close to core localities SD 2 and SD 3 (Fig. 2).
2 The present-day geomorphological characteristics of the SD 2 and SD 3 core localities closely correspond to our inferred palaeomorphologies (deduced from line S2) along the West Ubwari Fault area (WUF) (Fig. 10).

A major fault system similar to that adjacent to coring site SD 2 (Cape Magara Fault) dominated the oldest sequences (F and E) along the eastern side of the South Rusizi half-graben (Fig. 10B). Intermediate between an open basinal setting (core SD 3) and a fault-dominated system (core SD 2) existed during the more recent sequences (D to A) (Figs 3B and 10C–E). Such characteristics allowed us to extrapolate the SD 2 and SD 3 core data (lithology, sedimentation rate) onto line S2 in the West Ubwari Fault area of the South Rusizi half-graben (Figs 3 and 10). By doing this, we generated one RSRM locality named ‘Site 2–3’ in the eastern part of the South Rusizi half-graben, one RSRM locality (‘Site 10’) on the western side of the Ubwari horst and one locality on the Ruminze platform (‘Site 14’) (Figs 3 and 10).

Our first step in estimating sequence ages involved calculating all stratigraphic sequence thicknesses below each chosen key site from the sediment–water interface down to the Nyanja Event on multichannel line 16 and down to the (f) seismic boundary on sparker line S2 (Figs 3 and 5; Table 1). To convert two-way travel-times to thicknesses, we assigned an average wave velocity value to each sequence. These were calculated by linearly interpolating between minimum (1480 m s⁻¹) and maximum (3000 m s⁻¹) velocity estimates for the lake bottom sediments and the deep portion of the section (Nyanja Event), respectively (Fig. 3A). These estimates in turn were derived from estimated velocities for Lake Tanganyikan sediments on the basis of PROBE multichannel seismic data (Rosenhahn et al., 1988; Scholz & Rosenhahn, 1988; Scholz et al., 1989). Because no refraction data are currently available for Lake Tanganyika, the 3000 m s⁻¹ velocity estimate was based on refraction studies in analogous basins in Lakes Malawi and Baikal (Ding, 1991; Hutchinson et al., 1992).

After estimating stratigraphic thicknesses, it is also necessary to apply decompaction corrections to the calculations, to make the results comparable to piston-core-based accumulation rates. For decompaction calculations, we have followed the method of Scattered & Christie (1980). The degree of mechanical compaction is highly dependent upon lithology, which is unknown below maximum core penetration. However, core data analysis suggests that sedimentation is dominated by fine-grained deposits such as muds and diatom oozes. Within the upper interval (C, B and A Sequences) resolved with high-resolution seismic data (sparker and 5-kHz echosounding), we have observed a general 2/3:1/3 proportion between parallel, thin-bedded facies (presumed to be shales) and chaotic facies (presumed to be silt, sand or gravel). We have therefore used this proportion (shale to coarse-grained facies) in calculating decompaction ratios (Table 1).

We have modified Cohen et al’s (1993) original method of extrapolating a single sediment accumulation rate throughout the seismic study interval. For this study we have allowed accumulation rates to vary between sequences, dependent on the tectonic configuration of the study site over time. To do this we have used the mean sediment accumulation rates obtained from the various tectonostratigraphic conditions represented by our coring sites (Fig. 3; Table 1). These rates were then extrapolated over the stratigraphic interval during which the study site experienced those conditions. For this study, we used the following rate estimates (Tiercelin & Monleguer, 1991; Cohen et al., 1993): core SD 2, ‘proximal to fault escarpment’ conditions: mean 0.86 mm yr⁻¹; core SD 3, ‘open basinal’ conditions: mean 0.32 mm yr⁻¹; core SD 10, ‘steep channel side’ conditions: mean 0.15 mm yr⁻¹; core SD 14, ‘platform shallow water’ conditions: mean 9.405 mm yr⁻¹.

From our palaeotopographic reconstructions (Fig. 10) it is evident that for some intervals, certain study sites (chiefly ‘Site 2–3’ and ‘Site 10’) may have experienced intermediate tectonostratigraphic conditions. For example, at ‘Site 2–3’, Sequence Rt was deposited under pre rift, open basinal conditions, with an assumed accumulation rate of 0.32 mm yr⁻¹ (Figs 3A and 10A; Table 1A). Above Sequence Rt, the active rifting phase was marked by the deposition of the BuF and BuE Sequences, deposited very close to the major West Ubwari Fault (WUF), which was active during the maximum subsidence of the South Rusizi half-graben (Fig. 10B). Therefore, for this interval (conditions ‘proximal to fault escarpment’), we have assumed a sediment accumulation rate of 0.86 mm yr⁻¹ according to the core SD 2 value. After the deposition of the BuF and BuE Sequences, subsidence in the South Rusizi half-graben stopped, as suggested by the flat morphology of the BuD, BuC, BuB and BuA Sequences (Fig. 3B). Because ‘Site 2–3’ lies close to the incipient Cape Magara Fault at the time of the deposition of the BuD through BuA Sequences, we assume that depositional rates were intermediate between ‘open basinal’ and ‘proximal to fault escarpment’ conditions (Figs 3B and 10A–C). Therefore, we have used the arithmetic mean between the ‘open basinal’ and ‘proximal to fault escarpment’ values (0.59 mm yr⁻¹) as an accumulation rate estimate for this interval (Table 1B).

On the buried Magara shoal at ‘Site 10’ (Figs 3A and 10A), the Bu Sequence is thought to have been deposited (like the Rt Sequence at ‘Site 2–3’) under pre rift, ‘open basinal’ conditions (0.32 mm yr⁻¹) (Table 1A). Sequences MiB to MiB were deposited under conditions intermediate between the ‘open basinal’ and ‘steep channel side’ states. Therefore, we have assumed a mean value of 0.23 mm yr⁻¹ for this interval. Finally, we have applied
a value of 0.15 mm yr$^{-1}$ to the uppermost MBA Sequence, corresponding to the present-day sediment accumulation rates (core SD 10) on the steep western side of the Capart channel (Figs 7B and 10E). In the North Kigoma half-graben, at 'Site 14' on the Rumonge platform, there is no compelling palaeo-geographic argument for allowing sedimentation-accumulation rate estimates to vary over the site’s post Nyanja Event history (Figs 3 and 7C). This region appears to have experienced ‘platform shallow water’ sedimentation conditions throughout its history (Fig. 10). Therefore, we have assumed a constant sediment accumulation rate at ‘Site 14’ of 0.405 mm yr$^{-1}$, equivalent to the core SD 14 mean accumulation rate.

Based on the above assumptions, we have estimated the duration of sedimentation for each sequence (Table 1). This provides us with minimum ages for the first deposits above each sequence boundary. The possibility exists that at some or all sites, significant errors could have accrued from our method as a result of uncertainty about seismic velocities, sediment composition and compaction, and the duration of sequence boundary erosive events. We acknowledge such possibilities, but in the absence of dated cores can only note that our ‘minimum-age’ method provides the most defensible argument for sequence boundary age estimates currently available. Therefore, we have used these minimum ages as a basis for the chronological reconstruction of the Late Tertiary history of the North Tanganyika Basin.

**DISCUSSION – TECTONIC AND DEPOSITIONAL HISTORY**

The combination of different seismic stratigraphic analyses obtained from the three complementary seismic investigation methods has allowed us to propose a new depositional history for the Northern Tanganyika Basin. This history varies both in space and time between the different structural zones of the study area (Figs 2–4), and comprises six major transgressive-regressive phases (Figs 3 and 10). These can be grouped as follows: ‘Phase RBM’ (period of deposition of the RL, Bt, and M Sequences); ‘Phases F-E to B’ (the period of deposition of the F through B Sequences) and ‘Phase A’ (the period of deposition of the A Sequence, which corresponds to the C$_{1T}$, C$_2$ and C$_3$ Subsequences) (Figs 3B and 9). As a consequence of using different seismic resolution methods (multichannel, sparkler and 5-kHz echosounding), these sequences and phases are of very unequal thickness and duration (Table 1). However, each phase can be interpreted as a response to variations in regional tectonism and/or climate. Lake lowstand conditions characterized the end of each interval, and their corresponding palaeo-geographic reconstructions are shown in Fig. 10.

*Phase RBM* – initial synrift infills
(\~7.4 Ma to \~1.1 Ma)

The oldest sediments of the Northern Tanganyika Basin were deposited over prerift basement (the Nyanja Event). This prerift surface appears to have been very broad and flat (Rosendahl et al., 1988). Our age estimation method applied to multichannel seismic line 16 shows sedimentation commencing in the South Rusizi half-graben \~7.4 Ma (Fig. 3A and Table 1A). This age corresponds to the oldest deposits of the Rusizi Lower Sequence (Rt.) above the NE surface at ‘Site 2–3’. Further east, on the Ubwari horst (‘Site 10’) and in the North Kigoma half-graben (‘Site 14’), deposition over the NE surface appears to have commenced considerably later, at \~4.9 and \~3.6 Ma, respectively (Fig. 5 and Table 1A). Apparently, the Ubwari/North Kigoma structural areas were higher and underwent continued erosion or non-deposition during the early history of subsidence in the South Rusizi half-graben. These age estimates correspond reasonably well with earlier estimates of post-Nyanja Event sedimentation history given by Cohen et al. (1993): minimum age of 7.51 Ma for the South Rusizi area and maximum age of 2.43 Ma in the North Kigoma area.

Our palaeo-geographic reconstruction at the end of ‘Phase RBM’, marked by the (f) lowstand period (Fig. 10A), is mainly derived from data along line 16 (Fig. 3A). The (f) sequence boundary marks a major erosional event possibly related to a prolonged lake lowstand and onlap relationships shown in line 16 (Rosendahl et al., 1988) suggest that the lake may have been reduced to a very small size and shallow depth at this time. We estimate that lake level at this time was \~650–700 m bpll, which would have corresponded to a maximum water depth of \~30 m in the South Rusizi half-graben (‘Site 2–3’) (Fig. 10A). During this period, rift tectonism resulted in the partitioning of the northern part of Lake Tanganyika into two adjacent half-grabens, the actively subsiding South Rusizi half-graben, and the nascent North Kigoma half-graben, separated by the nascent Ubwari horst. The East and West Ubwari Faults (EUF and WUF) developed, and their growth subsequently accelerated along the eastern and western sides of the Ubwari structure at this time (Fig. 10A).

Minimum age estimates for the (f) palaeo-geographic reconstruction and its corresponding lowstand vary between \~1.1 Ma at ‘Site 2–3’ to \~670 ka at ‘Site 14’ and \~550 ka at ‘Site 10’ (Fig. 10A and Table 1A). The \~1.1 Ma age estimate may represent the initiation of the major rift tectonism phase in the area of line S$_2$, with prolonged erosion/non-deposition continuing at ‘Site 10’ and ‘Site 14’ for a considerable length of time after sedimentation resumed at ‘Site 2–3’. This diachronity in the resumption of sedimentation was probably the result of higher elevations in the eastern Ubwari/North Kigoma areas, which were tectonically quiescent until the subsidence of the South Rusizi half-graben commenced.

*Phase F-E* – first synrift deposits
associated with major tectonic activity
(\~1.1 Ma to \~393–363 ka)

Following the (f) lake lowstand, submergence of the study area began with the poorly defined uF Sequence.
that first filled the actively subsiding South Rusizi half-graben. On top of Sequence 6F, the minor BU(e) sequence boundary may correspond to a sedimentary break, possibly resulting from weakly erosive bottom currents during a rapid cessation of rising lake level. The resumption of rising lake level resulted in the deposition of a 50-m-thick succession of alternating organic muds and fine-grained detrital sediments forming Sequence 6E. Motion along the strike-slip, N130°–140°-trending Kaboge Fault Zone (KFZ) appears to have occurred at this time as indicated by the faulted, dome-shaped morphology of the 6H4F and BU sequences reflectors (Figs 2 and 7A).

On top of the Ubwari horst, Sequence 6B4F probably experienced lower sediment accumulation rates as a result of being a topographic high. It comprises ~65 m of stratified, coarse-medium deposits from detrital inputs that may have originated from northern or southern axial drainages (i.e. from either the north or south ends of the Rusizi half-grabens) (Fig. 1C). These deposits progressively filled the subsiding Magara slope and onlap onto the eastern, gently dipping flank of the Banza slope, indicating the occurrence of minor uplift of the Magara shoal and continuous normal faulting along the East Ubwari Fault (Figs 2 and 3B). The brief lake level stillstand (c) previously identified to the west, combined with the high topographic position of the Ubwari horst, could explain the complete erosion or nondeposition of Sequence 6B4F deposits. In the North Kigoma half-graben this (c) sedimentary hiatus may be represented in the 3B4F–E Sequence, but it was not identified on sparker lines 523 and 524.

The end of ‘Phase F–E’ deposition corresponded to a lake lowstand, indicated by the (d) sequence boundary (Fig. 10B). Horizontal and vertical tectonic movement along the Kaboge Fault Zone and West and East Ubwari Faults, and subsequent erosion, resulted in considerable topographic irregularity on the (d) sequence boundary. Relative fault motions between the Magara shoal and the Kaboge dome resulted in the formation of the incipient Baraka furrow. ‘Phase F–E’ therefore represents an important period of tectonism and major reorganization of rift structure.

The reconstruction of the (d) boundary morphology (Fig. 10B) suggests a lake level lowstand of ~350 m bpl1. The minimum age estimates for this lowstand are consistent between the three study sites (~393 ka for ‘Site 2–3’, ~368 ka for ‘Site 10’ and ~363 ka for ‘Site 14’) (Table 1B). This concordance in sequence boundary minimum age estimates suggests that the South Rusizi/Ubwari region was fairly flat topographically, possibly as a result of decreasing subsidence rates at the end of the 6B4F and BU sequences deposition. The South Rusizi/Ubwari areas were emergent during this interval, with a maximum water depth of ~25 m in the Baraka furrow area. At the same time, the topographic relief in the North Kigoma area may have been more pronounced, with significant relief on the uplifted Banza shoal on the Ubwari horst. We estimate that maximum water depths of ~30 m occurred in the North Kigoma half-graben (under conditions of renewed tectonism). The (d) erosional period appears to have resulted from a combination of tectonic and climatic events, while during the previous erosional episode (f) tectonic displacements were probably prevalent, thus masking climatic influences.

‘Phase D’ – sedimentation in a mature rift basin morphology (~393–363 ka to ~295–262 ka)

The initiation of Sequence D deposition in the North Tanganyika Basin corresponded to a major lake transgression. In the South Rusizi half-graben, the beginning of this lake level rise was marked by the excavation and infilling of BU deposits and widespread sediment irregularities, particularly in the Baraka furrow, by a body of coarse-grained sediments up to 40 m thick (Figs 7A and 10B). At ‘Site 2–3’, a minimum age of ~393 ka is proposed for the primary Sequence D sediments (Table 1B). These sediments onlapped onto the western sides of the South Rusizi half-graben and Ubwari horst. Their detrital component was probably derived from south-west of the South Rusizi half-graben by the Baraka River (Figs 1C and 2). This south–north drainage probably formed simultaneously with the development of the Cape Banza Fault (CBF) and the formation of Burton’s Bay. Cohen et al. (1993) have previously obtained an RSRM age estimate for Burton’s Bay of ~400 ka, which would be consistent with this scenario. The upper part of Sequence D consists of up to 65 m of alternating proximal and distal turbidites. These sediments were derived from nearby southern and distant northern axial drainages, respectively. The architecture of Sequence D in the South Rusizi half-graben indicates that subsidence rates decreased over time. Only strike-slip movements along the Kaboge Fault Zone are indicated by the smooth dome shape of Sequence D reflectors in this area (Fig. 7A).

On the Ubwari horst and in the North Kigoma half-graben, between 65 and 35 m, respectively, of sediments accumulated during deposition of Sequence D (Figs 7B and 8B). These consist of alternating distal (dominant) and proximal (subsidiary) turbidites. The proximal turbidites deposited in the horst were probably derived from the incipient Burton’s Bay. Thinner proximal turbidite layers in the Kigoma half-graben were derived from lateral drainage of the eastern rift escarpment (Fig. 1C). The distal turbidites in both the Ubwari and Kigoma areas probably originated from a northern axial drainage (Fig. 2).

Discontinuity (c) marks a renewal of lake lowstand conditions within the North Tanganyika Basin (~350 m bpl1), associated with erosional phenomena as shown by erosional truncations at the top of the D Sequence on the buried Magara shoal (Figs 7B and 10C). The (c) sequence boundary minimum age estimates vary between ~295 ka at ‘Site 14’, and ~262 ka at ‘Site 10’
and ‘Site 2–3’ (Fig. 10C and Table 1B). In the South Rusizi area, the lake floor appears to have been relatively flat following the end of active subsidence, presaging the morphology of the modern Bujumbura sub-basin. Emergent conditions (dry or swampy areas) probably existed throughout the sub-basin, and a small lake or lakes with a maximum ~25 m water depth developed within a nascent Magara–Banza depression on top of the Ubwari horst (Fig. 10C). Continuous subsidence of this depression resulted in the collapse of the central part of the Ubwari horst (along the Cape Banza Fault), which in turn promoted the formation of the Capart channel on the Magara slope. During the (c) period, on the gentle slope forming the Kigoma half-graben floor, Sequence D experienced minor but regular erosion, as indicated by discrete toplap terminations (Fig. 7C). Our lake level reconstruction (Fig. 10C) suggests a maximum water depth of ~30 m in the North Kigoma half-graben along the East Ubwari Fault. The opposite part of the half-graben appears to have been subaerially exposed, since the entire Sequence RMD and parts of Sequence RM2-E were removed at this time by active erosion (Figs 3B and 10C).

‘Phase C’ – the predecessor to the present North Tanganyika Basin
(~295–262 ka to ~193–169 ka)

This phase corresponds to the deposition of Sequence C. It began with a lacustrine transgression and the deposition of ‘basin fill’ deposits in the nascent Bujumbura sub-basin similar to those deposited during ‘Phase D’. The earliest deposits of ‘Phase C’ in the Bujumbura sub-basin comprise a 12-m-thick coarse-grained ‘basin fill’ unit, which could have been derived from rapid accumulation of detrital inputs from northern and/ or southern axial drainages. Above this, coarse clastic material transported by gravity-driven flows forms a 3-km-wide, 25-m-thick lens on the relatively flat bottom of the Bujumbura sub-basin. We interpret this unit to have been a deep lacustrine fan developed at the outlet of the Baraka River, connected to the Baraka channel excavated on top of the buried Baraka fanow during the early part of the transgressive period of Phase C (Figs 2 and 7A,B). In Lake Malawi, sedimentation adjacent to border faults is similar although typified by the development of fan deltas during lowstands. These fans emerge from small, incised drainages high on the rift shoulders (Scholz et al., 1990; Scott et al., 1991).

Above the coarse detrital fan and to the east as far as the Capart channel (Figs 3B and 7A,B), Sequence C is characterized by ‘sheet-drape’-type deposits similar to those sampled in recent sequences in the Bujumbura sub-basin (Bourrouille et al., 1992). They may have formed from the alternating accumulations of autochthonous organic matter and fine-grained detrital sediments, resulting in the deposition of up to 40 m of stratified organic mud and clay. This ‘sheet drape’ sedimentation pattern provides evidence for high lake stand conditions, when sedimentation was dominated by organic deposits (organic matter and biogenic material) with subordinate but recurrent detrital inputs. Presumably, the detrital inputs were mainly derived from an active northern axial drainage. This input may have been augmented by minor, southern lateral and axial drainage from Burton’s Bay through the Capart and Baraka channels (Figs 1C and 2).

In the Capart channel, Sequence C sediments are dominated by ~25 m of detrital sediments (Fig. 7B), cut by a unit that may be contemporaneous with the deep lacustrine fan previously defined at the base of BUC in the Bujumbura sub-basin. In the Magara–Banza depression, Sequence C comprises only the ‘sheet drape’ deposits, with the origin of detrital inputs restricted to the northern axial drainage. The fan-shaped configuration of the C Sequence in the western side of the depression suggests continuous axial subsidence along the Cape Banza Fault, whereas on its eastern side, the dome-shaped C Sequence indicates an active strike-slip motion along the Kabezi Fault (Figs 1C, 7B and 10D). In the Rumonge sub-basin, Sequence C is 65 m thick (Fig. 7C) and shows the same seismic character as in the Magara–Banza depression (Fig. 7B), suggesting a northern axial origin for the fine-grained detritus in both areas. Active faulting along the East Ubwari Fault (EUF) is marked by divergent and bent Sequence C reflectors. On the Rumonge slope, concurrent anticlinal faulting resulted in the formation of a small graben that enhanced the development of the N0°-trending Rumonge channel (Figs 2 and 7C).

The main discontinuity (b) atop Sequence C represents the most recent major erosive event in the North Tanganyika Basin. It is comparable to the (A) and (W) discontinuities previously defined in the Southern Mplulungu and Northern Bujumbura sub-basins, respectively (Fig. 6). We interpret discontinuity (b) as indicative of a lake level fall to ~250 m bpl (Fig. 10D). This value is close to the ~300 m bpl previously proposed for the (A) and (W) discontinuities by various authors (Tiercelin et al., 1989; Mondeguer, 1991; Tiercelin & Mondeguer, 1991; Bourrouille et al. 1992). Our minimum age estimates for resumption of sedimentation above discontinuity (b) range from ~193 ka (‘Site 10’) to ~191 ka (‘Site 2–3’) to ~169 ka (‘Site 14’) (Table 1B). These results are also comparable to the proposed ages of between ~150 ka and ~200 ka for the (A) and (W) erosional periods proposed by Tiercelin & Mondeguer (1991) and Bourrouille et al. (1992) (Fig. 6). Moreover, the (b) unconformity seems to correlate with the discontinuity at the base of the Songwe Sequence (10 to >100 m thick), the uppermost sequence observed on the multichannel and high-resolution seismic lines in the nearby Lake Malawi (Scholz & Finney, 1994; De Batist et al., in press). These authors estimated a maximum age of ~150 ka to ~120 ka for the Songwe Sequence.

A reconstruction of the palaeotopography at the (b)
lowstand period suggests the existence of a flat basin floor in the Bujumbura sub-basin and a more pronounced topography in the subsiding Magara–Banza depression. We estimate a maximum water depth of ~75–100 m within these two areas (Fig. 10D). At this time, 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 Kidoma half-graben was coupled with the uplift of the Banza shoal and probable rollover of the basin margin represented by the shallow Runumonge platform. Margin rollover systems have been previously described in similar half-grabens of Lake Malawi by Ebinger et al. (1993). This mechanism resulted in the formation of the present-day narrow Runumonge platform, with its large, steep slope (Figs 2, 3B and 4B). The lower part of the slope probably subsided as a consequence of tectonic movements along a N0–20°-trending fault, antithetic to the main East Ubwari Fault (Figs 1C and 10D). This subsidence also resulted in the deepening of a small graben that subsequently became the Runumonge channel. We estimate that the maximum water depth in the Runumonge sub-basin was ~155 m during the (b) lowstand, with the shoreline coinciding closely with the shelf break.

‘Phase B’ – development of the extent sub-basins (~193–169 ka to ~40–35 ka)

Renewed transgression in the North Tanganyika Basin was marked by the accumulation of Sequence B. The early stages of lake level rise were characterized by extensive erosion, particularly in the Capart and Runumonge channels, where the earlier Sequence C is strongly eroded (Fig. 7B,C). The Capart channel was probably fluvially excavated by an axial drainage flowing northward from the emergent Burton’s Bay (Figs 1C and 2). Erosion in the Runumonge channel was directly related to coarse detrital inputs derived from the lateral Ružibari River (Fig. 2) and diverted southward during this period by the N0–20°-trending Cape Magara Fault, which was becoming active at this time (Figs 2, 7C and 8B).

As in the previous ‘Phase C’, deposition during ‘Phase B’ in the Bujumbura sub-basin began with detrital inputs resulting in a 15-m-thick, coarse-grained fill (Fig. 7A). This fill is limited to a gentle depression at the foot of the western slope of the sub-basin. These sediments correspond to the F1 ‘basin fill’ Unit of Lake Malawi (Figs 1C and 6B), and are also similar to the ones described at the base of the previous D and C Sequences (Figs 7A and 8A). Above this unit, a coarse clastic lens, 3 km wide and 25 m thick, developed on the relatively flat bottom of the axial part of the sub-basin. This lens-shaped unit can be interpreted, as in the previous C Sequence, 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 (Fig. 7A), the upper part of Sequence B&B is characterized by ‘sheet drape’ sedimentation resulting in the deposition of a unit up to 25 m thick that provides evidence for a subsequent lake highstand.

To the east, in the Capart channel, the base of Sequence B&B comprises a narrow (1-km-wide), 25-m-thick, coarse-grained, lens-shaped body (Fig. 7B) that probably resulted from detrital inputs issued from the axial Kasandjala and Mutemba Rivers at the southern end of Burton’s Bay (Fig. 1C). These deposits are overlain in the channel axis by an 8-m-thick unit of alternating autochthonous organic sediments and thin detrital units. In the Magara–Banza depression, both regional (pelagic deposits) and local processes (coarse-grained gravity flows) resulted in the same sedimentary succession as exists in the Bujumbura sub-basin. Coarse deposits mainly accumulated on the western flank of the depression as a large (35 m thick by 2.5 km wide) lens which pinches out towards the east, probably as a consequence of a reduction of subsidence rate in the Magara slope and continuous uplift of the Banza slope (Figs 3B and 7B). This lens locally forms the entire Sequence B&B and may be interpreted as a ‘deep lacustrine fan’ formed by coarse inputs derived from the Ružibari River and diverted southward along the active sublacustrine Cape Magara Fault escarpment (Figs 2, 3B and 4B).

In the deepest part of the Runumonge sub-basin, Sequence B comprises a 75-m-thick deposit forming the primary infill of the Runumonge channel (Fig. 7C). The base of this deposit consists of a relatively flat and thin (15-m) coarse detrital layer similar to the lower part of Sequence B in the Bujumbura sub-basin and the Magara–Banza depression (Fig. 7A,B). Above this basal layer is a unit 60 m thick. At its base, a wide lens-shaped coarse-grained deposit probably corresponds to a ‘deep lacustrine fan’ formed at the outlet of the Ružibari River. Overlying this, small, coarse detrital lenses with cut-and-fill features formed as narrow channel infills, and are interbedded with thin stratified layers which we interpret as distal turbidites (Fig. 7C). As a whole Sequence B provides evidence for another cycle of lake level rise. On the Runumonge slope and platform, Sequence B consists of a thin (12-m) coarse layer overlain by a 60-m-thick unit of prograding clinoforms and aggrading strata (Fig. 7C). This succession corresponds to the ‘basin-fill’/’sheet-drape’-type deposits identified in the Bujumbura and Magara–Banza areas.

The flat shape of Sequence B reflectors in the Runumonge channel suggests that the subsidence of the North Kidoma half-graben stopped during the earliest stage of the B deposition. However, in Lake Malawi, Ebinger et al. (1993) suggested that marginal rollover as a response to earlier subsidence may have continued on this type of half-graben even after the termination of active subsidence. Thus, the Runumonge platform probably continued to undergo uplift during the deposition of Sequence B. The prograding/aggrading configuration of Sequence B (Fig. 7C) may have resulted from a combination of

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vertical tectonic movements on the Rumonge platform and lake level rise. Both of these processes would have resulted in a stabilization of relative lake level, and a major increase in sediment input as suggested by the 75-m-thick sediment accumulation in the Rumonge channel. Lake level rise and increased sediment input may also have resulted from a period of increased runoff. The prograding clinoforms developed on the Rumonge platform are probably the result of a nearby river delta (C. A. Scholz, personal communication, 1995), possibly the lateral Murembe River (Fig. 2).

The end of Sequence B deposition is marked by the (a) sequence boundary, which we interpret as evidence of a regressive period that can be correlated to the (A) boundary defined on the 5-kHz echosounding lines in the SouthernMpulungu sub-basin (Tiercelin et al. 1989; Mondeguer, 1991) (Figs 6 and 9). Combining data from core SD 10 (5.55 m length, 275 m water depth) and 5-kHz echosounding line 24 (Figs 2 and 9B), we observe that the base of core SD 10 penetrated the (a) boundary. This basal horizon is represented by clayey sands with gravels and numerous mud balls that have been interpreted as products of an intense erosive phase at the site of core SD 10 (Baltzer, 1991). On the basis of palynological studies, discontinuity (a) appears to be older than ~35 ka (A. Vincens, personal communication, 1994). Our minimum age estimates for the event (a) range from ~40 ka (‘Site 10’) to ~39 ka (‘Site 2–3’) to ~35 ka (‘Site 14’) (Table 1B). We estimate that lake level fell at this time to ~160 m bpl (Fig. 10E). These estimates are consistent with the ~150-m-bpl lake-level estimate for the (A) period in the Southern Mpulungu sub-basin (Mondeguer, 1991; Tiercelin & Mondeguer, 1991). 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 (Fig. 10E). We estimate a maximum water depth of ~220 m to ~130 m for the entire line S2.

‘Phase A’ – the modern Northern Tanganyika sub-basins (~40–35 ka to Present)

The most recent phase of deposition in Lake Tanganyika is represented by Sequence A, which is characterized by a succession of detrital and organic-rich deposits. In the Bujumbura sub-basin and Magara–Banza depression, the oldest Sequence A deposits comprise three organic to organodetrital units corresponding to the 5-kHz Subsequences C1u, C2, and C3 (Figs 6 and 9A). These sub-sequences are sequentially separated by two lake lowstand unconformities, designated (x) and (y) (Bourrouille et al., 1991). These subsequence boundaries are probably a result of minor lake level fluctuations related to small-scale regional climate and tectonic changes during the last ~40 kyr (Bourrouille et al., 1992). In the Rumonge channel, Sequence A is 20–25 m thick. Its base comprises coarse deposits that grade upwards into mixed organic/detrital deposits (Fig. 7C). Deposition of Sequence A (25 m thick) on the Rumonge platform seems to have started slightly later at ‘Site 14’ (~35 ka) than deposition of Sequence A at ‘Site 2–3’ (~39 ka) and at ‘Site 10’ (~40 ka), probably because the platform’s higher topographic position induced a delay in the resumption of sedimentation above unconformity (a) (Fig. 10E and Table 1B).

The ‘C1u Subphase’

This subphase corresponds to the deposition of the C1u Subsequence that can be identified both on sparker lines S2 and S1 (Figs 7 and 8) and 5-kHz echosounding lines 55 and 24 (Fig. 9A,B). In the Bujumbura sub-basin and Magara–Banza depression, the C1u base comprises a 15-m-thick unit of coarse detrital deposits that were locally derived and that filled irregularities on the (a) sequence boundary (Fig. 7B). These ‘basin fill’ deposits were probably formed during the first stages of transgression, which were marked by strong erosion. Overlying this first unit is a 15-m-thick ‘sheet drape’ deposit that resulted from quiet water deposition of organic sediments, periodically disturbed by the input of detrital fines. Such influx decreased over time as lake level rose and then stabilized. Studies of 5-kHz echosounding lines by Bourrouille et al. (1991) showed that this C1u Subsequence draped the entire palaeotopography of the Northern Tanganyika Basin to a mean thickness of 22 m.

In the Rumonge channel, the base of the C1u Subsequence is characterized by a large and flat coarse detrital lens up to 4 m thick (Fig. 7C). This unit may correspond to the distal portion of a ‘deep lacustrine fan’ that developed during the early stages of the transgression. The fan was fed by the lateral Kuzibazi River and subsequently diverted southward along the sublacustrine Cape Magara Fault escarpment (Figs 2, 7C and 8B). The upper part of the subsequence is an 8-m-thick accumulation of organic-rich ‘sheet drape’ deposits on the eastern flank of the channel, that grades laterally to the west into a succession of turbidites. On the relatively flat Rumonge platform (‘Site 14’), an 8-m-thick basal deposit of coarse detrital sediments resulted from processes similar to those acting in the Bujumbura sub-basin, also resulting in ‘basin-fill’-type deposits. Above this, ‘sheet drape’ organic deposits are interbedded with thin detrital layers both on the Rumonge platform and Rumonge slope (Fig. 7C).

The end of deposition of the C1u Subsequence corresponds with a minor regression and lake lowstand. This regression was accompanied by slight erosion, as shown by discrete toplap terminations which define the (x) unconformity (Bourrouille et al., 1991, 1992). These authors correlate this (x) unconformity with the (A) discontinuity defined in the Southern Mpulungu sub-basin (Fig. 6). From our detailed sparker seismic interpretations we propose that the south sub-basin (A) discontinuity is correlative with the northern basin (a)
unconformity (Fig. 6), making the (x) erosional event younger than the (A)/(a) event.

On the western side of the Capart channel, the C_{12} Subsequence pinches out towards the channel axis. This resulted in the superimposition of the (a) and (x) boundaries, as shown on 5-kHz echosounding line 24 (Figs 7B and 9B). At its base core SD 10 intersects these superimposed (x) and (a) boundaries (Fig. 9B). In the core the (x) boundary is marked by a 0.5-m-thick alternation of grey, quartz-rich fine sands and a few thin layers of black clay overlying the very coarse deposits marking the (a) boundary. These sediments have yielded a radiocarbon date of \textasciitilde 23 ka (Baltzer, 1991). This minor lake lowstand in the Northern Tanganyika Basin corresponds to the intermediate lake level conditions suggested for the 26–21.7-ka period in the Southern Mpu lungu sub-basin by Gasse et al. (1989).

The 'C2 Subphase'

Following the (x) lowstand, the thin (3–5-m) C_{2} Subsequence was deposited. Its particular seismic character (a dark grey line evident on all the 5-kHz seismic lines) can be easily recognized throughout the Northern Tanganyika Basin (Bourrouillec et al., 1991, 1992) (Fig. 9A). It also occurs throughout most sparker lines except on the Rumonge slope (Fig. 7C), where the steep and irregular morphology appears to have induced slumping towards the Rumonge channel. This strongly layered subsequence is thought to consist of numerous 'sheet drapes' organodetrital deposits with detrital inputs dominating at the base of the subsequence, possibly indicating a humid climate. This succession is also evident in the lower half of core SD 10 which shows alternating decimetre-thick layers of grey clayey silt and grey to black platy clays (Baltzer, 1991) (Fig. 9B). Bourrouillec et al. (1991) suggested that the C_{2} Subsequence deposits formed during a relatively brief and low-amplitude level rise followed by a major regressive period ('C_{2} Subphase'). This pattern seems to conform to the regional lake level pattern observed by Street & Grove (1979) for intertropical Africa between 22 and 21 ka (lake highstands) and between 21 and 18 ka (lowstands). In the Southern Mpu lungu sub-basin, the uppermost 5-kHz echosounding subsequence contains a 4-m-thick unit termed the β Unit, described by Tiercelin et al. (1988) and Tiercelin & Mondeger (1991), and characterized by abundant allochthonous teregenous inputs (Fig. 6). Radiocarbon dating and palaeomagnetic studies show that this β Unit was deposited between 22 and 18 ka (Tiercelin et al., 1988; Gasse et al., 1989; Williamson et al., 1991), and thus may correlate with the C_{2} Subsequence of the Northern Basin (Fig. 9A).

The end of the 'C2 Subphase' is marked by a lake lowstand represented by the (y) unconformity. Facies analysis of core SD 10 shows that the (y) unconformity is bound by silt and sand layers (Fig. 9B). Based on its seismic characteristics (toplap figures are scarcely observed), the (y) unconformity appears to have been an erosional event that lasted longer than the previous (x) event (Bourrouillec et al., 1991). Maximum lake lowstands in the late Pleistocene occurred during a period of great cooling and severe aridity, centred at 18 ka in the Southern Tanganyika Basin (Tiercelin et al., 1988; Gasse et al., 1989; Williamson et al., 1991; Vincens et al., 1993; Chalié, 1995), as well as in most of the intertropical African lakes (Street & Grove, 1979; Talbot & Johannessen, 1992). Comparable climatic conditions also characterized the Northern Tanganyika Basin (Vincens, 1989, 1993) and the nearby Burundi highlands (Bonnetfille & Riollet, 1988) before 15 ka, and resulted in the lake level fall marked by the (y) event. A major exception to this regional picture of late Pleistocene maximum lowstands occurred in nearby Lake Malawi, where the most recent lake lowstands (100–150 m bpll) date from 6 to 16 ka (Johnson & Davis, 1989; Finney & Johnson, 1991; Finney et al., in press), and are thus out of phase with the Tanganyika record.

The 'C3 Subphase'

Above the (y) boundary, the C_{3} Subsequence transparent layers correspond to laminated or homogeneous diatomaceous muds overlain by dark-green muds sampled at the top of cores SD 10 and SD 14 (Baltzer, 1991) (Fig. 9B,C). These sediments are the product of the most recent lake transgression resulting from the increased rainfall evidenced in cores from the Southern Tanganyika (Tiercelin et al., 1988; Gasse et al., 1989; Williamson et al., 1991; Vincens et al., 1993; Chalié, 1995) and Northern Tanganyika (Vincens, 1989, 1993) Basins, as well as in most basins of tropical Africa (Street & Grove, 1979; Stree-Perrut et al. 1989). These highstands began in the latest Pleistocene (~13 ka) and reached their maximum in the early Holocene, about ~9.9 ka (Haberyan & Hecky, 1987).

**CONCLUSIONS**

Our seismic stratigraphic analysis of multichannel, sparker, 5-kHz echosounding, and piston core data as well as the use of the reflection seismic-radiocarbon method (RSMR) for age estimation, has allowed us to propose a scenario of late Miocene to Recent tectonic and sedimentary evolution for the North Tanganyika Basin. Eight well-defined hiatus surfaces and associated seismic sequences identified in the lake's stratigraphy represent major environmental and depositional changes. Multichannel seismic data from the PROBE Project have allowed us to define the oldest 'Phase RBM' of geological evolution of the North Tanganyika Basin from sediments lying above the pre rift basement (Nyanja Event). The initiation of sedimentation during this phase began with deposition on top of basement rocks in the nascent South Rusizi half-graben at \textasciitilde 7.4 Ma. A substantial delay in the onset of sedimentation until \textasciitilde 4.9 Ma in the Ubwari.
area and ~3.6 Ma in the North Kigoma region resulted from a general westward tilting of the study area at this time, producing higher elevations in what would become the Ubwari/North Kigoma structural zones.

From end of the ‘RBM Phase’, five major transgressive-regressive phases (the ‘F-E to B Phases’, and ‘Phase A’ subdivided into C₁₀, C₁ and C₂ Subphases) and their separating lake lowstand periods have been identified using Project CASIMIR sparker and Project GEORIFT 5-kHz echosounding seismic and core data. This study has permitted us to reconstruct the details of the Northern Tanganyika Basin tectonosediary evolution from ~1.1 Ma to the present. The seven successive lake lowstand periods have been dated as follows: (f) ~1.1 Ma; (d) ~393–363 ka; (c) ~295–262 ka; (b) ~193–169 ka; (a) ~40–35 ka; (x) ~23 ka; and (y) ~18 ka. Corresponding palaeotopographic reconstructions have permitted us to estimate lake levels (bplll) at the end of each phase/lowstand period: ~650–700 m for (f); ~350 m for (d); ~350 m for (c); ~250 m for (b); and ~160 m for (a).

The late Miocene to Recent tectonic history of the Northern Lake Tanganyika Basin involved the following events. (i) Starting at ~7.4 Ma, an initial phase of rift tectonics affected a broad and flat prerift surface and resulted in the partitioning of the northern part of Lake Tanganyika into two adjacent half-grabens, the actively subsiding South Rusizi half-grabens to the west, and the nascent North Kigoma half-graben to the east, separated by the Ubwari structural high. (ii) From ~1.1 to ~393 ka, a major phase of reorganization in rift basin structure resulted in the accelerating subsidence of both the South Rusizi and the North Kigoma half-grabens, with major vertical movements along the West and East Ubwari Faults bounding the Ubwari horst. This tectonic movement resulted in the diachronacy of sedimentation in the two areas. (iii) Starting at ~393–363 ka, the study area acquired a clear rift-basin morphology, with well-defined half-grabens. During this time subsidence stopped in the South Rusizi region but continued in North Kigoma. Asymmetric uplift on the eastern side of the Ubwari horst (East Ubwari Fault) and subsidence along the Cape Banza Fault resulted in the collapse of the top of the horst and the formation of both the Burton’s Bay, the Magara–Banza depression and the nascent, fault-controlled Capart channel. (iv) Starting at ~295–262 ka the morphology began to conform to that of the present-day North Tanganyika Basin. Of particular importance was the formation in the west of the Bujumbura sub-basin. Asymmetric uplift continued along the Ubwari horst and vertical movements on the Cape Banza and Cape Magara Faults resulted in the deepening of the Magara–Banza depression. To the east, along the East Ubwari Fault, associated subsidence of the North Kigoma half-graben and antithetic faulting resulted in the formation of the Rumonge channel. (v) From ~193–169 ka until the Present the morphologies of the modern North Tanganyika sub-basins took shape. This period was mainly marked by an end to subsidence in the North Kigoma half-graben and a rollover response of the eastern rift margin represented by the Rumonge platform. Continuous collapse occurred in the Magara–Banza depression along the active Cape Banza Fault. The Cape Magara Fault became very active and well expressed in the sublacustrine topography around the Cape Magara area at this time, thus influencing the north–south axis drainage of the Rumonge channel.

The infill history of Lake Tanganyika over the past ~400 kyr has been characterized by a recurrent association of three types of deposits that repeat in each stratigraphic sequence. (a) Chaotic ‘basin fill’ deposits that form the lower parts of the successive sequences and accumulate during the first steps of each transgression. Such deposits result from the accumulation of tremendous volumes of heterogeneous, clastic material derived from both lateral and axial drainages. They were deposited rapidly over lake or stream valley floors on top of the underlying erosional surfaces from prior lake lowstands. (b) Lens-shaped ‘deep lacustrine fans’ that fill locally elongate depressions (Magara–Banza depression and western Bujumbura sub-basin) or deep channels (Capart and Rumonge channels). They probably formed from large, energetic coarse gravity flows generated from successive floods of lateral and axial drainages. (c) Uniform ‘sheet drape’ deposits generally form the upper two thirds of each sequence. They were formed under lake highstand conditions by alternating accumulations of autochthonous organic muds and allochthonous, fine-grained detrital material related to intermittent turbidity flows from both axial and lateral drainages.

Our data show that from the late Miocene (~7.4 Ma) until the mid-Pleistocene (~400 ka), deposition in the sub-basins of Northern Lake Tanganyika was mainly controlled by tectonism. Starting at about ~1.1 Ma, a series of lake-level fluctuations began, which decreased in intensity over time as a consequence of rapid sedimentary filling under conditions of declining tectonic subsidence. Since ~400 ka, Lake Tanganyika seems to have responded dramatically to climatic changes as suggested by repeated patterns of lake-level fluctuations. The climatic signal has thus been more pronounced in recent sedimentary phases as the importance of tectonism has waned. Notwithstanding the different resolutions of the seismic methods we used, within-basin tectonic activity appears to be insufficient to explain the degree and number of lake-level fluctuations observed from the middle mid-Pleistocene to the present.

This series of reconstructions for Northern Lake Tanganyika has important implications for the tectonic, climatic and biological evolution of this rift basin that will be presented in a companion paper.

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