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GEOMETRIC AND KINEMATIC DEVELOPMENT OF BORDER FAULTS AND ACCOMMODATION ZONES, KIVU-RUSIZI RIFT, AFRICA

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Abstract. Three representative basins in the Western rift system of East Africa are bordered along one side by high-angle normal faults with 2- to 5-km throws (border faults). In plan view ~100-km-long systems of linear border faults form curvilinear border fault segments bounding the East Kivu, West Kivu, and Rusizi basins. The opposite sides of these asymmetric basins are bounded by lower relief faulted monoclines or en echelon ramps. The largely unfaulted rift flanks have been uplifted 2 km above the 1.3-km-high East African plateau, with uplift narrowing basins during Quaternary time. Maximum estimates of ~E-W crustal extension within basins are less than 25% (< 16 km), and planar border faults may penetrate the crust. The East Kivu and West Kivu basins are linked across the rift valley by a horst that serves as a hinge for subsidence in both basins. The westward tilted East Kivu and eastward tilted Rusizi border fault segments are linked along the rift by oblique-slip transfer faults that also accommodate along-axis differences in elevation. Upper Miocene-Recent eruptive volcanic centers within the comparatively high-strain interbasinal region (accommodation zone) generally coincide with the tips of border fault segments and transfer faults. The orientations of Miocene-Recent dip-slip and oblique-slip faults show little correlation with Precambrian shear zones or foliations in metamorphic basement. Differences between the East Kivu, West Kivu, and Rusizi basins in the age of initial faulting, subsidence, and age/composition of volcanic products suggest that border fault segments developed diachronously and propagated along the length of the rift. This along-axis border fault propagation and the crosscutting geometry of transfer faults contribute to the segmentation of the Western rift valley.

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INTRODUCTION

The tectonically active continental rift system of East Africa consists of two limbs, the Kenya (Gregory) rift valley and the more remote Western rift valley, which is filled by deep lakes along much of its length (Figure 1). The results of recent studies in the great lakes of the Western rift valley show that the narrow rift valley is segmented along its length into a series of approximately 100-km-long extensional basins [e.g., Crossley and Crow, 1980; Rosendahl and Livingstone, 1983; Ebinger et al., 1984]. The predominantly half-graben basin morphologies and alternating basin asymmetries observed along the length of the Western rift are similar to patterns of uplift and subsidence observed in the Kenya rift and in other continental rift systems [e.g., King, 1978; Chenet and Letouzey, 1983; Logatchev et al., 1983; Reynolds, 1984; Smith and Bruhn, 1984; Bosworth, 1985]. However, the spatial and temporal linkage of individual rift segments along the length of the Western rift valley was not addressed in earlier interpretations of widely spaced seismic reflection profiles from Lakes Tanganyika and Malawi [e.g., Rosendahl et al., 1986; Ebinger et al., 1987; Morley, 1988]. Likewise, the extent of faulting outside the lake regions of the Western rift and across the broad uplifted flanks of the rift valley have been poorly understood; existing geological maps from many areas are at a scale of 1:500,000 to 1:2,000,000.

Tholeiitic and alkalic basalts erupted in late Cenozoic time are found within parts of the Kivu and Rusizi basins and in the region between these extensional segments [Holmes, 1940; De Paepe and Fernandez-Alonso, 1981; Tack and De Paepe, 1983]. Previous petrological and radiometric analyses of these volcanic rocks provide information on the timing and evolution of the South Kivu volcanic provinces, one of four isolated volcanic provinces within the Western rift [e.g., Guibert, 1977a,b; Bellon and Pouclet, 1980; De Paepe and

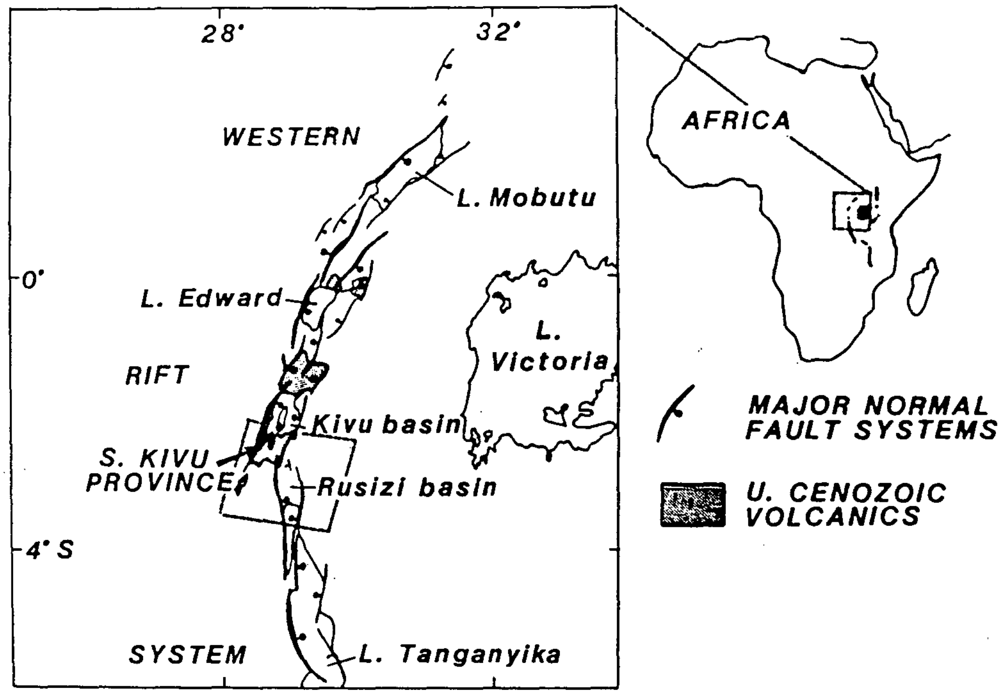

Fig. 1. Location of the Kivu and Rusizi basins and the South Kivu volcanic province within the Western rift, the western branch of the East African rift system. Inset indicates location of study region with respect to African continent, with lakes shaded. Box encloses Landsat 5 coverage used in this study.

Fernandez-Alonso, 1981; Kampunzu et al., 1986; Auchapt et al., 1987; Pasteels et al., 1989]. However, the structural relationship between volcanic centers and fault systems bounding basins was not addressed in these studies. The relative timing of volcanism, uplift, and subsidence within individual basins provides constraints on the development of the along-axis segmentation and proposed propagation of rifting in the Western rift [e.g., Crossley and Crow, 1980; Ebinger et al., 1984]. Likewise, the orientation of faults linking basins along the length of the rift valley (transfer faults) constrains the amount and direction of extension.

The primary objectives of this study were to examine the three-dimensional geometry of border fault systems bounding Western rift basins and transfer faults linking individual half-grabens. Remote sensing and detailed field studies conducted within the East Kivu, West Kivu, and Rusizi basins and along the uplifted flanks of these basins and existing geophysical data from beneath Lake Kivu and the Rusizi basin at the northern end of Lake Tanganyika provide information on structural patterns and the internal geometry of basins [Wong and Von Herzen, 1974; Patterson, 1983; D. Stone, personal communication, 1987]. Because several researchers have suggested that the geometry of late Cenozoic border faults and transfer faults are controlled by preexisting shear zones within Precambrian basement [e.g., McConnell, 1972; Rosendahl et al., 1986; Daly et al., 1988], the relationship between late Cenozoic faults and Precambrian structures is examined. These studies were also designed to investigate a proposed along-axis propagation of volcanism and rifting and to determine the spatial and temporal relationship between eruptive volcanic centers and faults bordering and linking basins. Within the Kivu and Rusizi basins, previously dated volcanic units, core data from sedimentary basins, and a new ^{40}K-^{40}Ar age determination provide kinematic constraints on the timing of crustal movements within extensional basins and along the uplifted flanks of the rift valley.

BACKGROUND

Discontinuous normal faults bounding the Western rift valley follow a roughly north-south trend along the western side of the broad East African plateau. The 150-200 km-wide zones of uplift flanking both sides of the rift valley rise over 1 km above the surrounding topography of the 1.3-km-high East African plateau and over 4 km above sea level. Faults bounding the Kivu and Rusizi basins have developed within Proterozoic orogenic belts linking the Archaean cratons of central and eastern Africa (Figure 2). The Rusizian system (~2100 Ma) is characterized by northwest trending mylonites and shear zones [Villeneuve, 1980], and the Burundian system northeast of Lake Tanganyika is characterized by northeast striking folds and shear zones [Theunissen, 1988, a, b]. A comparison of Figures 1 and 2 suggests that the general location of the Western rift valley may follow inherited zones of weakness within the continental lithosphere and avoid the Archaean cratons.

Seismic reflection data from Lakes Malawi and Tanganyika reveal that depth to basement is greatest at the foot of escarpments near the central part of ~100 km-long border fault segments, producing spoon-shaped basins. However, directions of asymmetry and amounts of subsidence vary from basin to basin [Crossley and Crow, 1980; Ebinger et al., 1984; Rosendahl et al., 1986; Morley, 1988]. Earthquake focal mechanisms from the Western rift system indicate that normal faults bounding basins dip 40°-60° at depths of 17-29

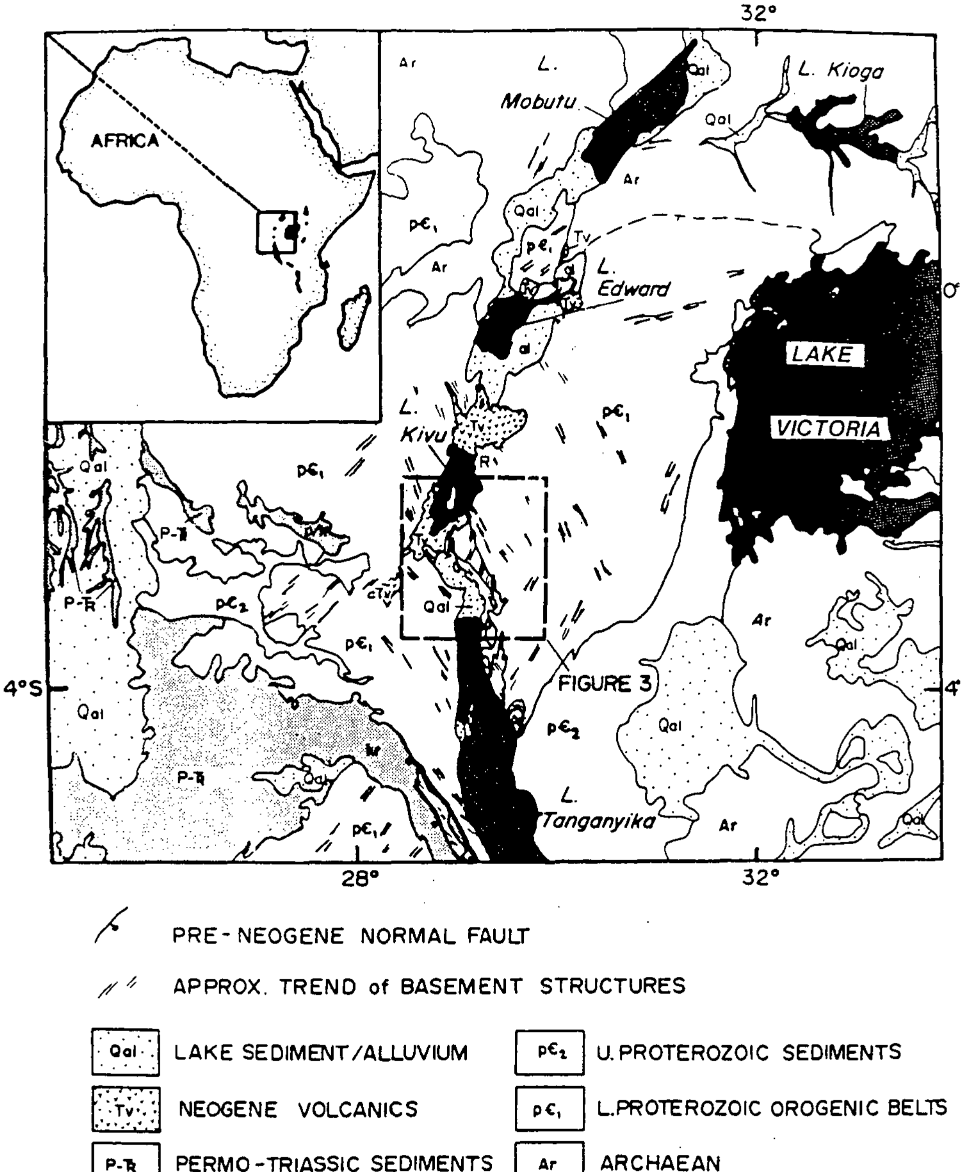

Fig 2. Summary of regional geology showing the general orientation of folds, mylonites, and metamorphic foliations within Precambrian basement (light lines). Inset shows location of study region in African continent. Dashed box encloses region shown in Figure 3. Geologic information from Cahen and Snelling [1984]; Lavreau et al. [1981], Lepersonne [1977], Theunissen [1989a, b]. Note poor correlation between basement structures and late Cenozoic faults shown in Figure 1.

km beneath rift basins [Shudofsky, 1985; Zana and Hamaguchi, 1978].

On the basis of the results of seismic refraction experiments, estimates of crustal thickness beneath the Kivu and the northern Tanganyika rift are approximately 30 km, or slightly less than the 35- to 41-km-thick crust found beneath the largely unfaulted region between the Western and Kenya rift valleys [Rykounov et al., 1972; Bram and Schmeling, 1975; Hebert and Langston, 1985]. However, crustal thickness beneath individual rift basins may be less than 30 km because these seismic refraction profiles average crustal structure over distances of 200-800 km [e.g., Bram and Schmeling, 1975]. Estimates of crustal thinning within other Western rift basins based on reconstructions of fault geometries range from 3-11%, or less than 4 km, assuming an approximately east-west extension direction [Ebinger et al., 1987; Morley, 1988]. Chorowicz [1983] interpreted a NW-SE extension direction and suggested that a northwest striking transform fault produced the rift valley segmentation between Lakes Kivu and Tanganyika.

Initial volcanism in the South Kivu volcanic province between the Kivu and Rusizi basins commenced at approximately 10 Ma, or 10-15 m.y. after initial volcanic activity within the central and eastern East African plateau

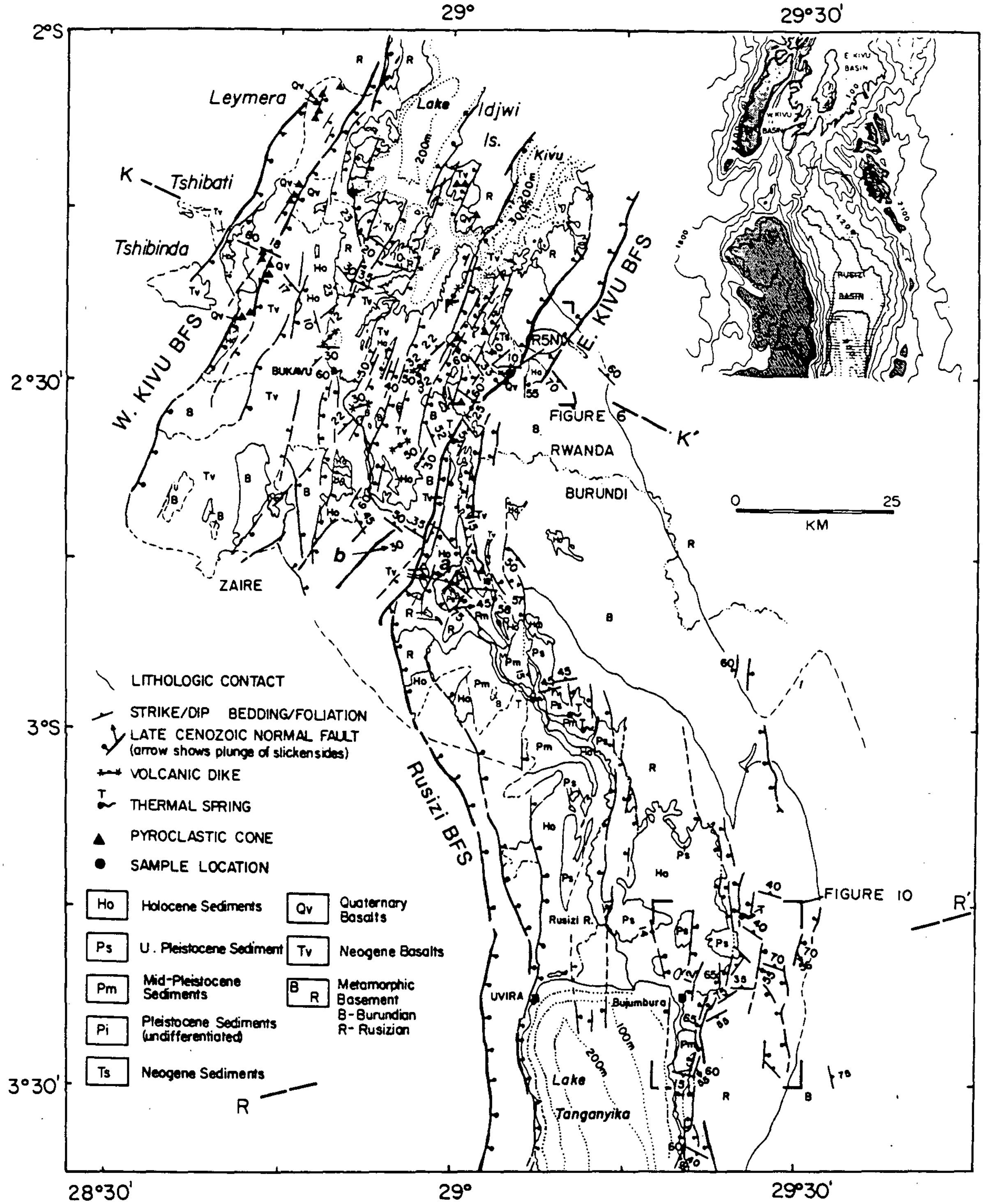

Fig. 3. Structural interpretation of Landsat image from South Kivu province calibrated by detailed field studies. Inset shows elevations (above sea level) within study region. Lithologic units mapped by Boutakoff [1939], Lepersonne [1977], De Paepe and Fernandez-Alonso [1981], Ilunga [1984], and Theunissen [1986; 1989 a, b]; faults and contacts dashed where uncertain. K-K', R-R' indicate locations of cross sections shown in Figures 8 and 11, respectively. Boxes enclose regions shown in more detail in Figures 6 and 10; a, b are locations referred to in text.

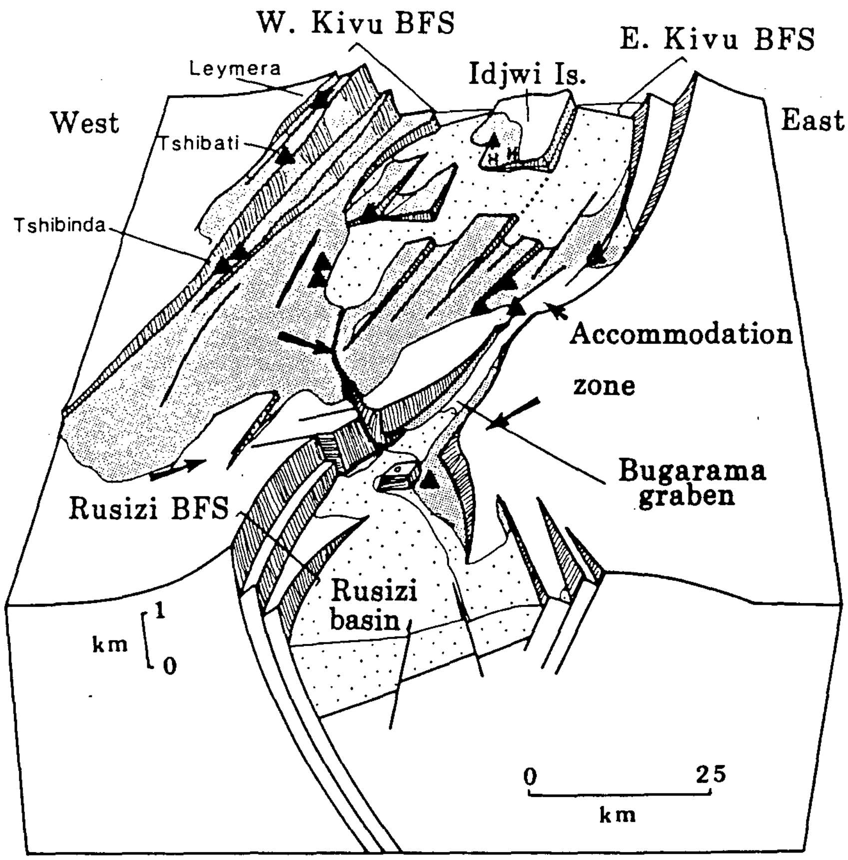

Fig. 4. Three-dimensional model of spoon-shaped rift basins and accommodation zone between basins (scale approximate). Lacustrine sediments and alluvium indicated by dotted pattern; upper Cenozoic volcanics indicated by stippled pattern; volcanic centers indicated by triangles; hyaloclastites indicated by H; BFS is Border Fault Segment. Arrows point to ~30-km-wide, ~20-km-long accommodation zone. Border fault segments are regionally curvilinear in plan view, and ~100-km-long basins are spoon-shaped. Volcanic centers are located within basins near the accommodation zone and along the tips of border fault segments.

region [e.g., Baker, 1986; Pasteels et al., 1989] (Figure 1). Pasteels et al. report three principal stages of volcanic activity: a period of tholeiitic volcanism 10-6 m.y., a period of alkalic volcanism 8-4 m.y., and a period of tholeiitic and alkalic volcanism beginning at 2 Ma and continuing to historic time. Distinct compositional variations between these flow suites indicate that they were derived from different parent magmas and/or different source depths [e.g., Bellon and Pouclet, 1980; Kampunzu et al., 1986; Auchapt et al., 1987]. The Rusizi River, which serves as the hydrographic connection between Lake Kivu at an elevation of 1420 m and Lake Tanganyika at an elevation of 774 m, developed during Holocene time when Lake Kivu was dammed in the north by volcanic flows from the Virunga province [Hecky and Degens, 1973].

SOUTH KIVU VOLCANIC PROVINCE

Throughout this study, border fault segments refer to approximately 100-km-long sections of fault systems that are characterized by large throws and uplifted rift flanks (high topographic relief), and that bound half-graben [e.g., Ebinger et al., 1987]. Transfer faults link discrete extensional basins in regions referred to as accommodation zones [e.g., Burchfiel and Stewart, 1966; Harding and Lowell, 1979; Bally, 1982; Gibbs, 1984; Bosworth, 1985; Rosendahl et al., 1986]. From north to south, border fault segments are delineated along the northwestern margin of Lake Kivu (West Kivu basin), along the southeastern shores of Lake Kivu (East Kivu basin), and along the western margin of the Rusizi basin (Figure 3). The Rusizi basin is separated from the Kivu basins by a fault-bounded topographic high that transects the rift valley within this accommodation zone (Figure 4).

BASIN STRATIGRAPHIES

Individual basalt flows are of limited areal extent and eruptive centers are difficult to locate due to pervasive alteration and vegetative cover. Local basin stratigraphies described below are based on petrographical correlations with the three volcanic stages and dated sedimentary horizons [Degens et al., 1973; Stoffers and Hecky, 1978; Bellon and Pouclet, 1980; Pasteels et al., 1989]. In addition, a picrite interbedded with lake sediments was dated by standard ^{40}K-^{40}Ar analyses at the Berkeley Geochronology Center.

Within the East Kivu basin, highly altered tholeiitic basalts

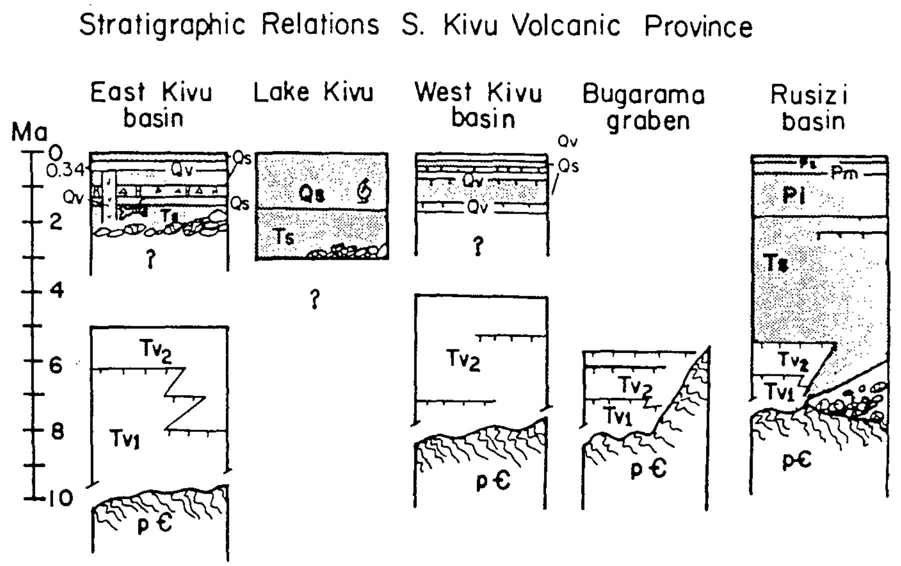

Fig. 5. Summary of stratigraphic relations within the East Kivu, West Kivu, and Rusizi basins and the Bugarama graben. Vertical scale is units of time; thickness of units not drawn to scale. Units Tv_1 (tholeiitic basalts), Tv_2 (alkalic basalts), and Qv (olivine basalts) based on range of ages determined in ^{40}K-^{40}Ar analyses made by Pasteels et al. [1989]. Age of olivine basalt unit Qv in Gisakura terrace (0.342 ± 0.1 Ma) from ^{40}K-^{40}Ar analyses of sample R5N (Figures 3 and 6). Hyaloclastites indicated by ΔH Δ. Sedimentary units (stippled pattern) beneath Lake Kivu from Stoffers and Hecky [1978] and above basalts in Rusizi basin from Ilunga [1984], and D. Stone (personal communication, 1987).

(Tv_1) similar to samples dated at 10-6 Ma overlie metamorphic basement at the base of the East Kivu border fault segment (Figures 3 and 5). Within the southeastern and central part of the East Kivu basin, these tholeiitic basalts are overlain by alkalic basalts, basanites, and hawaiites (Tv_2) petrographically correlated with flows dated at 8.0-5.4 Ma (De Paepe and Fernandez-Alonso, [1981], Pasteels et al., [1989] and this study). Within both sequences (Tv_1, Tv_2) basalt flows are generally less than 5 m thick, and depositional contacts are conformable or erosional. Predominantly varved biogenic sediments interbedded with pyrite and siderite-rich sandstone and siltstone layers (Ts) directly overlie metamorphic basement and basalts beneath Lake Kivu [Degens et al., 1973; Stoffers and Hecky, 1978].

Along the series of subparallel faults forming the East Kivu escarpment a previously unreported lacustrine sedimentary sequence containing Neogene mammal bones was found (Gisakura terrace; Figures 3 and 6). Along the western margin of the Gisakura terrace, basalt tubes filled by reworked Precambrian sandstones are interbedded with rounded cobbles interpreted as an ancient lake shoreline (Ts; Figure 5). These volcaniclastic units are overlain by a <1-m-thick sequence of well-sorted siltstones (Ts, Qs), and two approximately 5-m-thick basalt flows (Qv). The upper picrite flow (Qv) was dated at 0.342 ± 0.10 Ma (B. Drake, personal communication, 1987). The Gisakura sequence thickens toward Lake Kivu where metalliferous sediments are interbedded with hyaloclastites and thin olivine basalt flows, probably the basinal equivalent of olivine basalts (Qv) found along its western margin. On the basis of compositional similarity to sediments described by Stoffers and Hecky [1978], the coarsening upward Gisakura terrace sequence is correlated with upper Pleistocene sequences beneath Lake Kivu.

A thin (<100 m) sequence of predominantly alkalic basalts (Tv_2) overlies metamorphic basement within the West Kivu basin (Figure 5). These flows are correlated with alkalic basalts dated at 8.0-4.1 Ma by Pasteels et al. [1989] that cover the southwestern West Kivu basin (Figure 3). Thickness variations of locally occurring units suggest that possible centers for fissure eruptions are north of Bukavu and along faults bounding peninsulas along the western shores of Lake Kivu (Figure 3). Beneath Lake Kivu, volcanic sequences are overlain by up to 500 m of primarily biogenic sediments [Degens et al., 1973]. Flows (Qv) dated at 1.9, 0.350, and 0.175 Ma and derived from uneroded cinder cones located along the West Kivu escarpment flowed down the escarpment and fill topographic depressions in upper Miocene to lower Pliocene basalts (Tv_2) found at the base of the escarpment [Bellon and Pouclet, 1980; Pasteels et al., 1989] (Figures 3 and 5).

In the northern Rusizi basin tholeiitic (Tv_1) and alkalic basalts (Tv_2) petrographically correlated with flows dated at 7.6 and 6.5 Ma, respectively, directly overlie metamorphic basement (Figure 5; De Paepe and Fernandez-Alonso, [1981], Tack and De Paepe, [1983], and this study). At least two flows occur within the Rusizi basin and they are separated by a paleosol (Figure 5). Some basalt layers in the northern Rusizi basin show signs of sublacustrine alteration, but there is no evidence a lake existed at the time of eruption (Tack and DePaepe, [1983], and this study). Plio-Pleistocene gravels and clastic sediments lacking volcanic clasts are found along the inner Rusizi fault system and cover much of the plain from Lake Tanganyika north to the Bugarama graben (Figures 3 and 5). Upper Pleistocene alluvial cones and talus (Ps) extend onto the valley floor over lacustrine sediments along the faulted eastern and western margins of the basins (Figures 3 and 5).

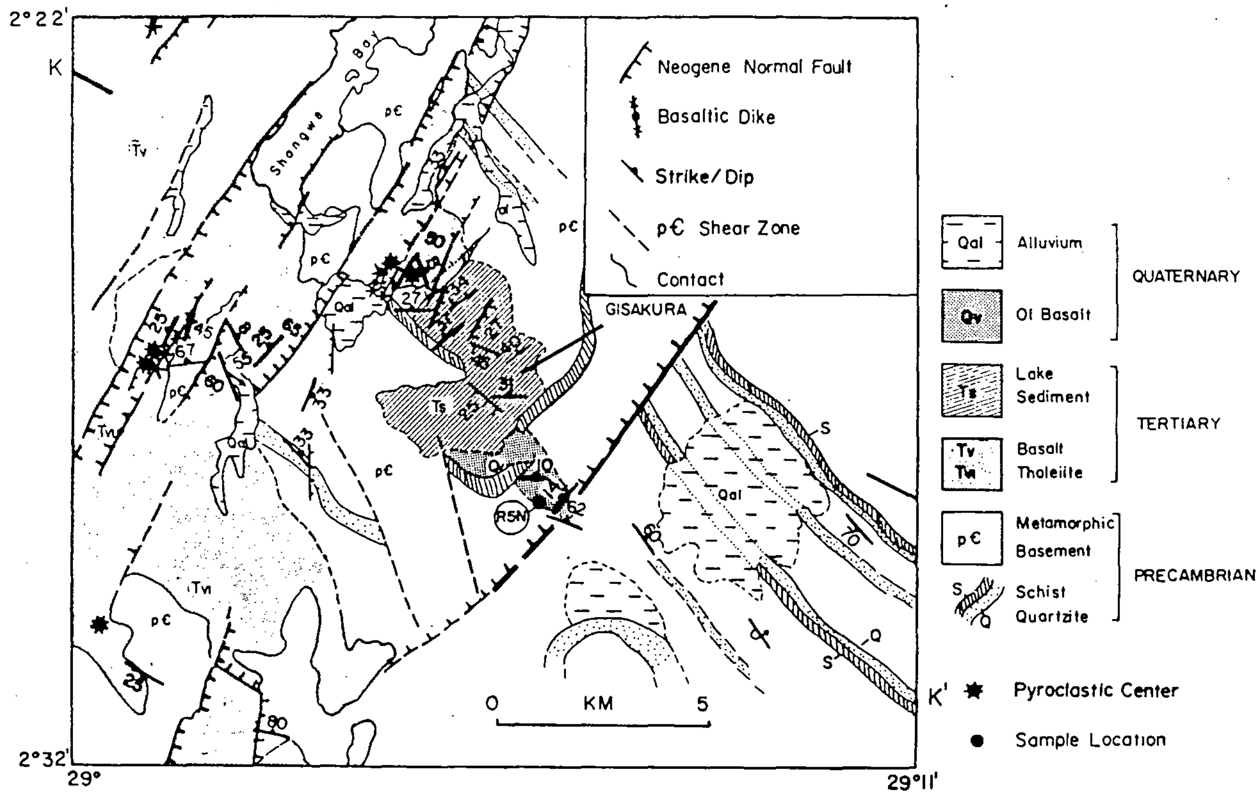

Fig. 6. Detailed map of East Kivu border faults showing uplifted terrace of lacustrine sediments (Gisakura) and crosscutting relation between late Cenozoic and NW-striking Precambrian folds. Basement structures from K. Theunissen (personal communication, 1987).

Metamorphic basement within the narrow Bugarama graben is overlain by tholeiitic (Tv_1) and alkalic basalts (Tv_2) and a thin cover of alluvium (Figure 5). Tufa and alluvium (Qs) within the Bugarama graben are separated by calcrete horizons, indicating a depositional environment characteristic of the hotter, drier climate found at the lower elevation of the Rusizi basin [e.g., Ilunga, 1984]. Peaty horizons within the hydrothermal deposits contain upper Pleistocene mammal bones and gastropods typical of Lake Tanganyika assemblages (A. Cohen, personal communication, 1986; P. Williamson, personal communication, 1987).

STRUCTURAL ANALYSES

The geometry of border fault systems, accommodation zones, and the timing of faulting with respect to volcanism and sedimentation are illustrated through examples of structural and stratigraphic patterns within the Kivu and Rusizi basins (Figures 3 and 4). Color Landsat 5 Multi-Spectral Scanner (MSS) imagery were used as a primary data base to examine fault patterns within, between, and along the uplifted flanks of these basins, as well as to compile existing data (see the appendix). Within the South Kivu volcanic province, aerial photographs were used to establish continuity of structures and to extend field observations throughout the region covered by Landsat imagery. Although exposure is poor in the Kivu region, faults bounding basins are readily apparent in Landsat imagery due to the youthfulness of faulting and lack of thick volcanic cover found in the Kenya rift system to the east [e.g., King, 1978; Baker, 1986]. Multichannel and single channel seismic reflection data from the Rusizi basin and existing seismic reflection and core data from beneath Lake Kivu constrain subsurface structures [Degens et al., 1973; Wong and Von Herzen, 1974; Patterson, 1983; D. Stone, personal communication, 1987]. The uplifted flanks of the northern and central parts of the remote Western rift system are thickly vegetated; hence these imagery were used to identify areas for detailed field studies conducted during 1986.

Where faults and lineaments had not been noted previously, the following criteria were used to differentiate late Cenozoic faults from older structures: displacement of Miocene-Recent volcanic and sedimentary units, amount of topographic relief, appearance of fault scarps, occurrence of hot springs, and horizontal offsets of prerift basement faults and geologic contacts. The scale of topographic maps and aerial photographs used was 1:50,000.

East Kivu Basin

The East Kivu border fault segment bounds the eastern side of Lake Kivu, and Idjwi Island marks the western extent of the East Kivu basin (Figures 3 and 4). To the north along the eastern margin of the Virunga volcanic province, the East Kivu border fault segment splays out in a diffuse zone of normal faults with throws less than 50 m. A series of

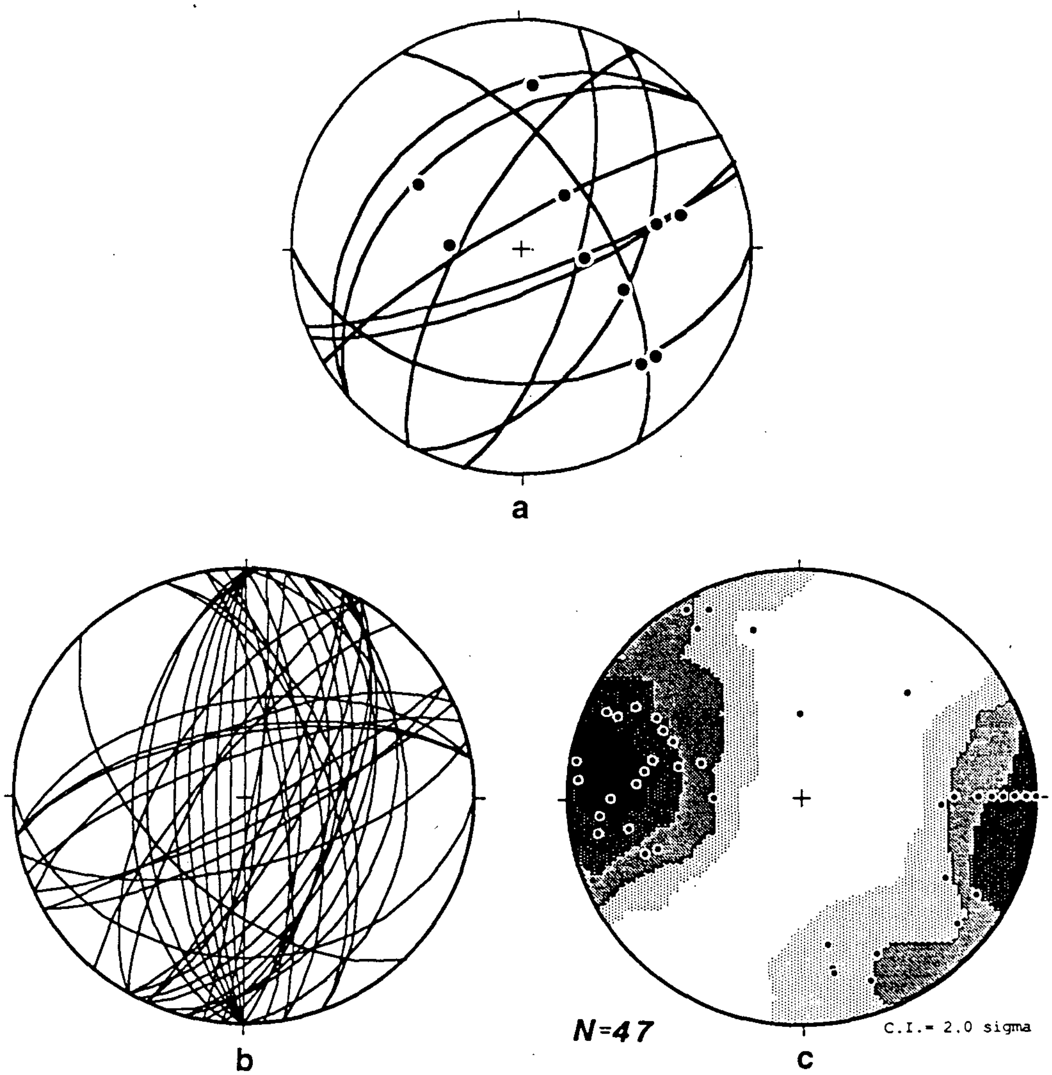

Fig. 7. (a) Equal-area lower hemisphere projection of slip vectors and planes of slickensides within the ~30-km-wide, ~20-km-long Kivu-Rusizi accommodation zone (Figures 3 and 4). (b) Equal-area lower hemisphere projection of fault planes within accommodation zone showing two fault sets striking N-S and ENE. (c) Contour of poles to fault planes shown in Figure 7b.

subparallel high-angle normal faults rise in a stair-step pattern to the elevation of the uplifted flank (~2500 m). Along the East Kivu escarpment, the Gisakura sedimentary terrace is elevated at least 600 m above correlative sequences beneath Lake Kivu (Figure 6). Low-angle growth faults, drag folds, debris flows, and bedding parallel sills in the uplifted lake sediments attest to synsedimentary volcanism and deformation (Figure 6). Near the base of the escarpment, metamorphic basement is in fault contact with a small pyroclastic cone on the hanging wall (Figure 6).

The central and southern sections of the East Kivu basin are characterized by 5- to 10-km-wide tilted blocks that generally strike NNE and parallel the East Kivu border fault segment (Figures 3 and 6). Wong and Von Herzen [1974] report the thickest sediments (~500 m) at the base of the central East Kivu border fault segment. Outside the lake, steeply dipping faults (>45°) cut mid-Miocene tholeiitic and alkalic basalt flows, and depositional contacts between units often dip 10°-25° (Figures 3 and 6). However, no systematic decrease in dip to the top of the section was found, indicating that fault block rotations postdate late Miocene time. Erosion has removed the thin cover of basalts from the upthrown footwall of many of these blocks, but marshes along the hanging walls of normal faults attest to recent faulting (Figures 3 and 6). In the southern part of the East Kivu basin near the accommodation zone the orientation of slickensides along N60°E and less commonly N10°W fault planes indicate that the most recent offsets have a component of oblique-slip movement (Figures 3 and 7).

The East Kivu basin is bounded to the west by Idjwi Island (Figure 3). In the geometrical arrangement of the West and East Kivu border fault segments the Idjwi Island horst separates an eastward tilted basin from a westward tilted basin (Figure 4). Seismic reflection data from Lake Kivu indicate

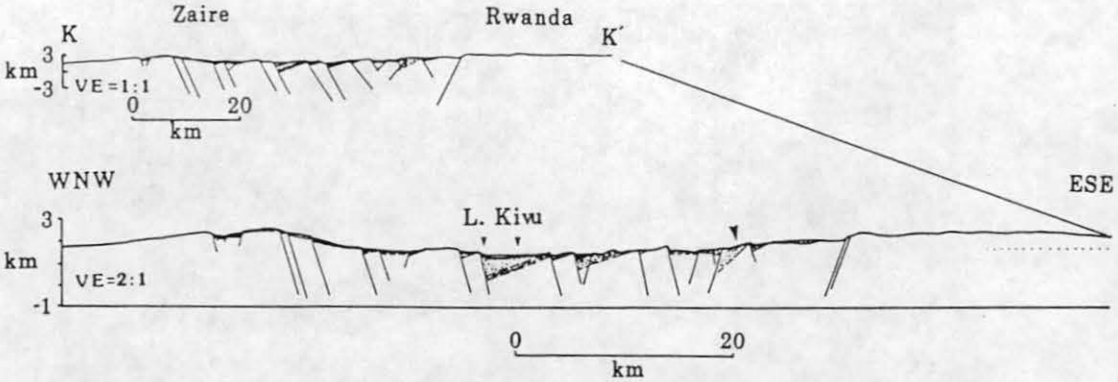

Fig. 8. Cross section of the West and East Kivu basins along profile K-K' north of the accommodation zone and south of the Idjwi Island horst (Figure 3). Neogene sediments indicated by stipple pattern; Miocene-Quaternary volcanics by random lines; metamorphic basement unshaded; bold arrow points to Gisakura terrace (Figure 6); region between small arrows beneath Lake Kivu. Depth to detachment estimated at 20-30 km, although estimate affected by excess topography on eastern side of profile due to 1300-km-wide East African plateau (dotted line). The bottom is a vertically exaggerated section (2:1) drawn to illustrate stratigraphic relations.

that the horst extends north of Idjwi Island to the northern end of Lake Kivu [Wong and Von Herzen, 1974]. Wong and Von Herzen report that hydrothermal vents occur along the length of Idjwi Island, and earthquake epicenters are located along both margins of the horst. Thus this horst serves as a hinge for active subsidence in basins bounded by border fault segments on opposite sides of the rift valley (Figures 3 and 4).

Known eruptive volcanic centers coincide with border faults and intrabasinal faults within the East Kivu basin (eg., Figures 3, 4 and 5). For example, fissures and dikes striking N10°E and N30°E parallel the East Kivu border fault segment (Figure 6). Recent road cuts expose a complex of pyroclastic cones along an east dipping normal fault that is along strike and possibly continuous with east dipping faults on the eastern side of the Idjwi Island horst (e.g., Figures 3 and 4). DePaepe and Fernandez-Alonso [1981] suggest that centers for the second-stage alkali basalts (Tv_2) were located near the accommodation zone, where explosive volcanic centers and fissures are located along the margins of tilted blocks (Figures 3 and 4). Upper Miocene tholeiites (Tv_1) and hyaloclastites interbedded with Holocene diatomites indicate repeated episodes of volcanic activity occurred along faults bounding Idjwi Island [Guibert, 1977b].

West Kivu Basin

The western side of Lake Kivu is bounded by a steep escarpment that borders a basin filled with up to 500 m of sediments and basalts derived from the Virunga province to the north [Wong and Von Herzen, 1974]. The West Kivu escarpment comprises a series of 10- to 20-km-long, high-angle normal faults within a 10- to 15-km-wide zone. Earthquake hypocenters within the Kivu region cluster along the West Kivu escarpment [Wohlenberg, 1968; Zana and Hamaguchi, 1978]. This fault zone extends north into the Virunga basin where normal faults displace Quaternary basalt flows (Pouclet, [1977], De Mulder and Pasteels, [1986], and this study) and splays at its southern end (Figure 3). Throws along the West Kivu border fault segment increase from its northern and southern tips toward the central West Kivu basin where the adjacent rift flanks reach an elevation of 3 km above sea level (inset, Figure 3). Pleistocene flows and intervening ash layers (Qv) have been faulted and rotated to the northwest along the escarpment, and the footwalls of normal faults often are covered by marshes, attesting to recent faulting (e.g., Boutakoff, [1939], Meyer and Burette, [1957], and this study).

Fault patterns within the southern part of the West Kivu basin near the accommodation zone are characterized by approximately 5-km-wide tilted blocks bounded by normal faults oriented N10°-20°E or less often, normal faults oriented north-south to N10°W that dip to the northeast (Figures 3 and 7). Southeastward thickening sedimentary wedges cover three approximately 5-km-wide tilted fault blocks beneath Lake Kivu [Degens et al., 1973]. Tilted fault blocks marked by hot springs form peninsulas that project into Lake Kivu, and fault scarps bounding these blocks are triangular (Figure 3). The remote region at the southwest end of the West Kivu basin is bounded by approximately east-west striking faults and northward-dipping monoclines on its southwestern margin, based on interpretations of Landsat imagery and topographic relief (Figure 3).

Within the West Kivu basin, eruptive centers and dikes identified in the field coincide with normal faults striking north-south and oblique-slip faults striking ENE (Figure 3). Along the southern West Kivu border fault segment, Quaternary basalt flows from cinder cones aligned parallel to the escarpment flowed down the escarpment (Meyer and Burette, [1957], Guibert, [1977a], Bellon and Pouclet, [1980], and this study), indicating that the escarpment existed prior to latest Pliocene time. Along the uplifted rift flank, flows directed away from the axis of the rift valley (northwestward) follow present-day topography, attesting to the youthfulness of faulting and volcanism in this area.

The cross section of the South Kivu volcanic province shown in Figure 8 is based on observations made along a new road crossing the East Kivu basin that provides excellent exposure and on a transect across the West Kivu border fault segment in Zaïre (Figure 3). No subsurface structural information is available from the southern part of Lake Kivu, and depth to metamorphic basement beneath Lake Kivu is interpreted from an extrapolation of dips measured on the footwall side of rotated fault blocks (Figure 3). This

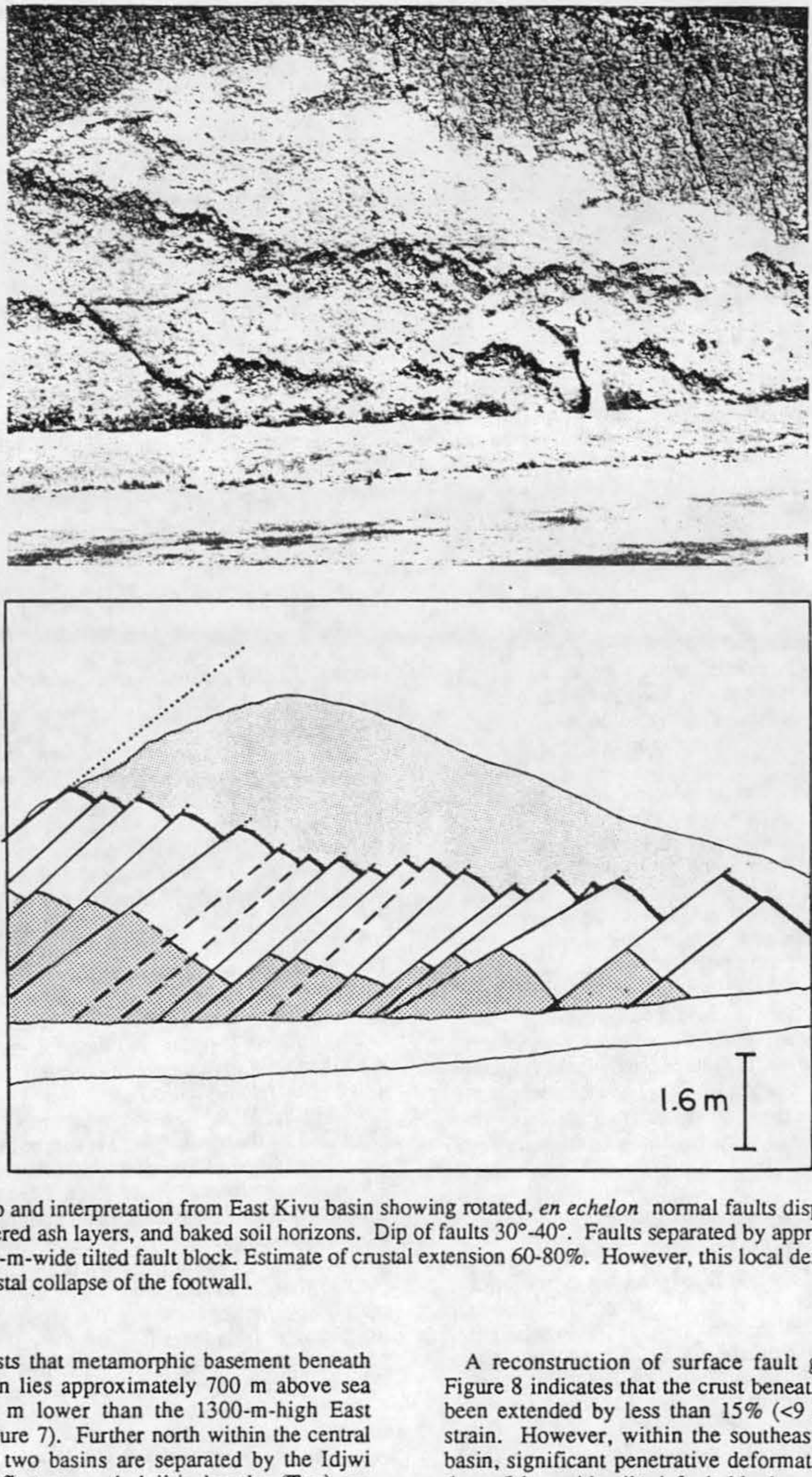

Fig. 9. Photo and interpretation from East Kivu basin showing rotated, *en echelon* normal faults displacing basalts (shaded), altered ash layers, and baked soil horizons. Dip of faults 30°-40°. Faults separated by approximately 1 m within the 10-m-wide tilted fault block. Estimate of crustal extension 60-80%. However, this local deformation may be due to crestal collapse of the footwall.

extrapolation suggests that metamorphic basement beneath the West Kivu basin lies approximately 700 m above sea level, or over 600 m lower than the 1300-m-high East African plateau (Figure 7). Further north within the central part of the rift, the two basins are separated by the Idjwi Island horst. The first-stage tholeiitic basalts (Tv_1) are restricted to the East Kivu basin and Idjwi Island, whereas eruptive centers for the second-stage alkalic basalts are found within both basins (Figure 5). Tv_1 and Tv_2 are probably less than 100 m thick, based on an extrapolation of dips of contacts between volcanic units.

A reconstruction of surface fault geometries shown in Figure 8 indicates that the crust beneath the Kivu region has been extended by less than 15% (<9 km), assuming plane strain. However, within the southeastern part of the Kivu basin, significant penetrative deformation internal to one of these 5-km-wide tilted fault blocks was found (Figure 9). Geometrical relations between normal faults that dip 30°-40° and that displace a Miocene basalt flow dipping 30°-35° indicate that crustal extension locally is of the order of 60-80% [e.g., Wernicke and Burchfiel, 1982]. This outcrop-scale extensional deformation within fault blocks would be

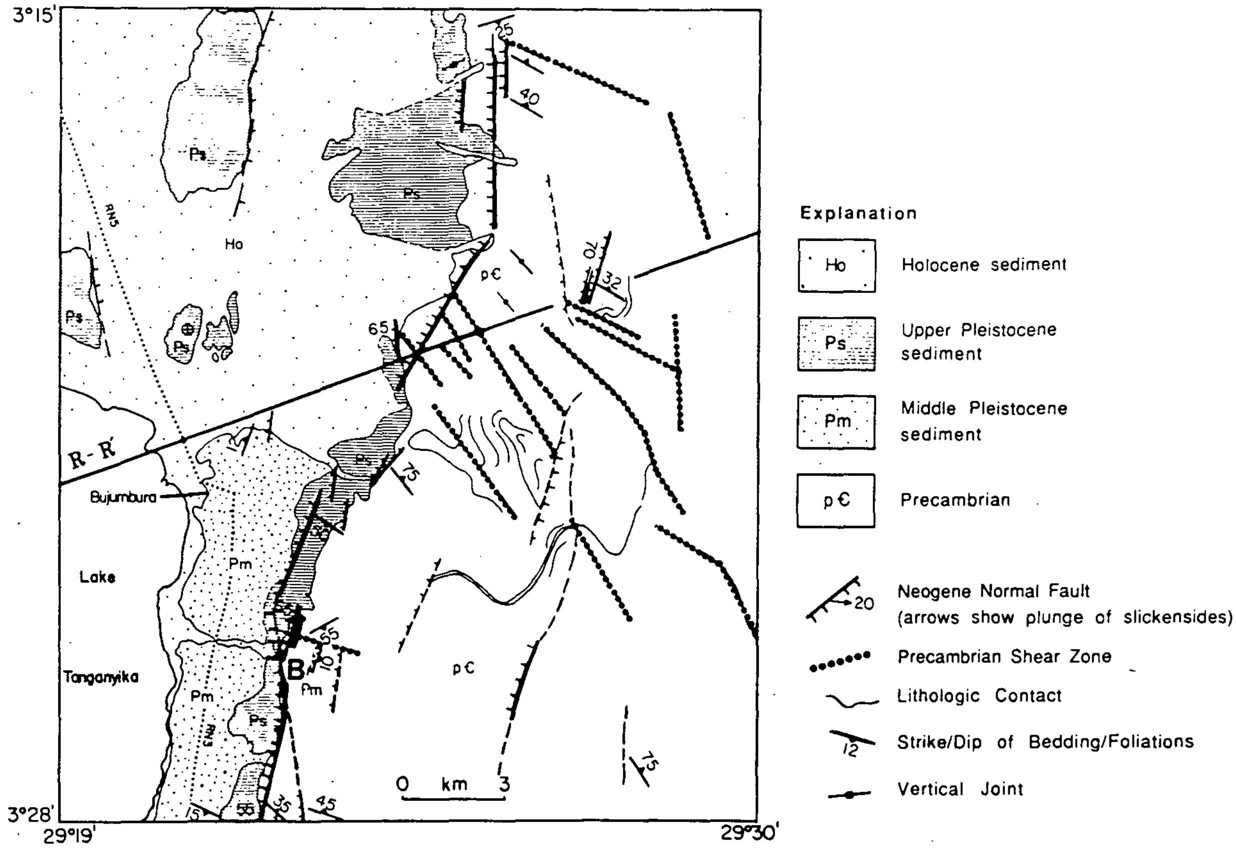

Fig. 10. Relationship between Precambrian structures and Neogene faults along the eastern margin of R-R' across the Rusizi basin. Basement structures from Theunissen [1989a,b] and this study. Note predominantly N40°W strike of Precambrian faults. Uplifted terrace of lacustrine sediments indicated by letter B. Dotted lines (RN5, RN3) are Burundi highways.

irresolvable on seismic reflection records, and if common, could bias estimates of crustal extension based on reconstructions of fault geometries apparent in seismic records to smaller values. However, these low-angle normal faults may be part of a local slide or crestal collapse affecting the shallow subsurface and not involving metamorphic basement. Therefore crustal extension within the Kivu rift valley probably falls at the lower end of the range between these two estimates (15-80%), in agreement with less than 25% extension interpreted from seismic refraction studies in the Kivu region [e.g., Bram and Schmeling, 1975]. Depth to detachment [e.g., Gibbs, 1983] is estimated at 20-30 km, although the much longer wavelength topographic relief of the East African plateau centered to the east of the rift may affect this estimate.

Rusizi Basin

Northeast striking normal faults bounding the west side of the Rusizi basin curve southward to follow a nearly north-south strike along the central Rusizi basin (Figures 3 and 4). As along the Kivu escarpments, the Rusizi border fault segment consists of 3-4 subparallel normal faults within a 10- to 15-km-wide zone. The innermost scarp is the least dissected suggesting progressive collapse of the hanging wall [e.g., Gibbs, 1984]. Seismic reflection data from the central part of the Rusizi basin indicate that faults synthetic to the western side of the basin remain steep (>45°) to depths of at least 5 km, measured from the top of the uplifted escarpment [Patterson, 1983]. Seismicity within the Rusizi basin during the period 1958-1965 clusters along the central Rusizi border fault segment, and epicenters of aftershocks from a 1960 earthquake fall throughout the depth range 5 to 30 km [Wohlenberg, 1968; Zana and Hamaguchi, 1978]. The northeastern side of the Rusizi basin south of the accommodation zone is bounded by ~10-km-long, en echelon normal faults or relay ramps [e.g., King, 1978; Larsen, 1988]. These faults dip 60°-75° to the west, and slickenside striations generally are subvertical within narrow (<5 m wide) brecciated fault zones (Figure 3). An eastward tilted wedge of lacustrine sediments along the eastern, or monoclinal, margin of the Rusizi basin forms a terrace similar to the uplifted Gisakura terrace along the East Kivu border fault. These sediments may be correlative with compositionally similar sequences (Pm) found at elevations 500 m below the terrace (Figure 10).

Sedimentary accumulations are greatest near the central part of the Rusizi border fault segment and decrease toward the tips, producing a spoon-shaped basin tilted to the west [Patterson, 1983; D. Stone, personal communication, 1987].

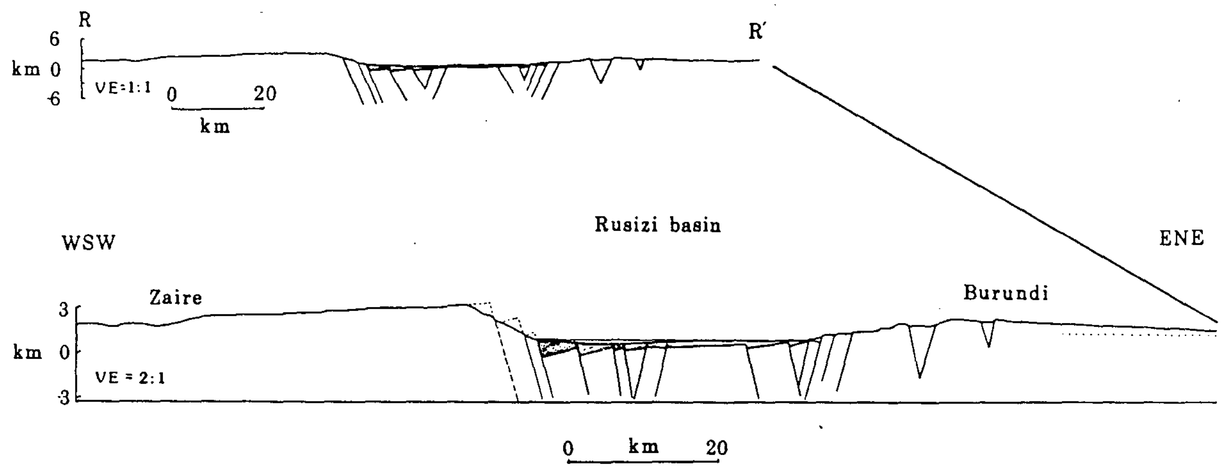

Fig. 11. Rusizi cross-section R-R' (Figure 3) illustrating monocline in rollover position on eastern side of basin [after Gibbs, 1984]. Neogene sediments indicated by stipple pattern; metamorphic basement unshaded. Structural patterns beneath Lake Tanganyika interpreted from Figure 8, [Patterson, 1983], dashed lines indicate probable pre-erosion geometry. Depth to detachment estimated at 20-30 km [Morley, 1988], although estimate affected by excess topography on eastern side due to 1300 km-wide East African plateau (dotted line). The bottom is a vertically exaggerated section (2:1) drawn to illustrate stratigraphic relations.

The floor of the northern Rusizi basin has been faulted into a series of north-south striking tilted blocks often marked by hot springs (Figure 3). Faults bounding these rotated blocks cut Miocene basalts (Tv_1, Tv_2) and overlying sediments (Figure 3, 4 and 11). High-amplitude magnetic anomalies (100 γ) 10 km south of the southernmost exposure of basalts are comparable to anomalies measured where basalts crop out, indicating that these late Cenozoic sediments overlie or are interbedded with upper Miocene basalts (D. Stone, personal communication, 1987). Surface sediments within the central Rusizi graben (Rusizi River) contain volcanic clasts derived from flows covering the accommodation zone between the two basins [Tack and DePaepe, 1983; Ilunga, 1984]. Thus faults bounding the inner graben are younger than the 11 Ka Rusizi River (Figure 3). Seismic reflection data from the Rusizi plain (Amoco Production Company) and beneath Lake Tanganyika reveal over 1.5-km throw along many faults, although most faults have little surface expression [Chorowicz and Thouin, 1985; Patterson, 1983; D. Stone, personal communication, 1987] (Figure 11).

A cross-sectional profile of the central Rusizi basin shown in Figure 11 illustrates the half-graben morphology characteristic of Western rift basins, with the western flank of the Rusizi basin at a higher elevation (3275 m) than the eastern flank of the basin (2600 m). Regionally, the eastern side of the Rusizi basin is an uplifted monocline in the rollover position [e.g., Gibbs, 1984]. If acoustic basement beneath the Rusizi basin is interpreted as metamorphic basement, the floor of the Rusizi basin lies over 400 m below sea level, or nearly 1200 m lower than metamorphic basement beneath the Kivu basins. Eastward dipping faults are found near the eastern side of Lake Tanganyika, and faults penetrate to all stratigraphic levels in sedimentary sequences beneath Lake Tanganyika (Figure 11). Estimates of crustal extension from a reconstruction of surface fault geometries mapped along the margins of the rift and faults that displace acoustic basement in seismic reflection profiles beneath Lake Tanganyika are less than 15% (<10 km), assuming that the regional extension direction is ENE, or subparallel to transfer faults described below. On the basis of multichannel seismic reflection profiles from the Rusizi basin, Morley [1988] estimates a depth to detachment for normal faults at 20-30 km or near the base of the crust. Thus the deep depth to detachment, the rollover geometry of this 65-km-wide basin, and earthquake epicenters at depths of 5-30 km indicate that high-angle border faults fracture the crust, consistent with recent interpretations of gravity data from the Western rift [Ebinger et al., 1989].

Accommodation Zone

Upper Miocene basalt flows cover the accommodation zone between the Kivu and Rusizi extensional basins, a region where transfer faults link basins bounded by border faults on opposite sides of the rift valley (Figures 3 and 4). The eastward shift of the rift valley axis between the Kivu and Rusizi basins is accompanied by a 650-m drop from the elevation of Lake Kivu to the average elevation of the Rusizi plain (Figure 3; inset). Within this accommodation zone, the narrow Bugarama graben is bounded to the east by seismically active normal faults [e.g., Wohlenberg, 1968] that mark the southern tip of the East Kivu border fault segment and to the west by north-south striking normal faults (Figure 3). Altered alkalic basalts dip 15°E along the eastern fault system bounding the Bugarama graben (Figure 3). Fault zones are less than 5 m wide and often mineralized, and slickenside striations are subvertical along fault planes (Figure 3). Undissected triangular fault scarps, hot springs, and sag ponds at the base of the western fault system are evidence for recent activity (Figure 3). Prior to faulting, thin basalt flows in the Bugarama graben were contiguous with Tv_1 and Tv_2 found 500 m higher in the Kivu basins, although much of the basalt cover has been eroded along the uplifted footwall of the western fault system (DePaepe and Fernandez-Alonso, [1981], and this study).

Oblique-slip faults striking ENE crosscut the rift valley between the southern Bugarama graben and the northern Rusizi basin (Figure 3). The floor of the Bugarama graben ramps up to the north, but no faults are apparent in the field or in aerial photographs at its northern end. Numerous ENE-striking faults occur within the northernmost part of the Rusizi basin, and segments of the youthful Rusizi River follow this northeast trend (Figures 3 and 7). For example, an approximately 2-km-wide rhomb-shaped fault block is bounded on two sides by normal faults that strike N45°E and on the other two sides by faults that strike N10°E, or subparallel to the Rusizi border fault segment (location a, Figure 3). The plunge of slickenside striations along faults striking N45°E is 45°; these N45°E-striking faults are interpreted as oblique-slip faults probably with a left lateral offset (Figure 3 and 7). Isolated exposures of lacustrine sediments and tholeiitic basalts atop this horst were probably once contiguous with a 7.6 m.y. tholeiite flow (Tv_1) now exposed adjacent to the horst but found at an elevation 300 m lower [Tack and DePaepe, 1983] (Figure 3). The basalts at the lower elevation (footwall of adjacent tilted fault block) flowed to the east, but now are tilted down to the west, indicating fault block rotations postdate late Miocene time (Tack and DePaepe, [1983], and this study). Within the northern Rusizi basin near the accommodation zone, synthetic and antithetic faults are more numerous than within the central basin, and faults are difficult to trace between multichannel seismic reflection profiles shot by Amoco Production Company that are separated by only 2-4 km (D. Stone, personal communication, 1987) (Figure 3).

ENE striking faults on the western flank of the Rusizi basin and the southern end of the Bugarama graben correspond to narrow, steep-walled valleys characterized by brecciated and calcified fault zones (Figure 3). These faults have no apparent throw, but left-lateral movement is indicated by slickenside striations plunging 30° to N45°W in one of these fault zones (location b, Figure 3). Subparallel faults striking N40°-60°E mapped by Villeneuve [1980] also are interpreted as strike-slip faults, or oblique-slip faults where scarps are visible in aerial photos (Figures 3 and 7).

On the basis of kinematic and geometric constraints within the 40-km-wide, 30-km-long accommodation zone, N40°-60°E-striking oblique-slip transfer faults accommodate differential translational movements between the Kivu and Rusizi extensional segments (Figures 3 and 12). Throws along oblique-slip transfer faults and north-south striking normal faults accommodate the elevation difference between the Kivu and Rusizi basins. The en echelon fault segments bounding the eastern margin of the northern Rusizi basin form a series of relay ramps linking East Kivu border faults with the eastern Rusizi basin via the Bugarama graben. The geometry of normal faults and en echelon, right-stepping oblique-slip transfer faults support an approximately east-west extension direction [e.g., Fairhead and Girdler, 1971; Shudofsky, 1985; Morley, 1988]. The absence of late Cenozoic faults striking northwest argues against tectonic models in which offset along a northwest striking transform fault produces the observed rift valley segmentation [e.g., Chorowicz and Mukonki, 1980; Chorowicz, 1983].

Relationship Between Neogene and Precambrian Structures

In general, the attitude of Precambrian shear zones and metamorphic foliations, and the attitudes of late Cenozoic faults bounding the Kivu and Rusizi basins are poorly correlated (Figure 3). For example, high-angle normal faults of the East Kivu border fault segment striking N20°E crosscut Proterozoic sediments folded along N30°-45°W striking axes (Figure 6). However, Neogene uplift and erosion enhances lithologic contacts between basement structures in the Landsat imagery (Figure 6). Within the West Kivu basin, lineaments trending N30°-50°W interpreted by Boutakoff [1939] as rift-related structures correspond to unstriated joints in metamorphic basement flooring the Kivu valley and are probably unrelated to late Cenozoic strains. Prerift faults and shear zones in Precambrian basement along the eastern margin of the Rusizi basin generally are oriented N20°-N50°W or nearly orthogonal to late Cenozoic normal faults (Figure 10). Elsewhere in the Rusizi basin, some late Cenozoic faults are subparallel to high-angle north-south striking Precambrian faults, but these Precambrian shear zones show little evidence for reactivation (K. Theunissen, personal communication, 1986 and this study).

A poor correlation between late Cenozoic faults and Precambrian structures also is found within the accommodation zone between the two basins. Lower Proterozoic gneisses and phyllites exposed along the length of the 500-m-deep Rusizi River gorge have a dominant N30°-50°W foliation that is nearly orthogonal to late Cenozoic strike-slip and oblique-slip faults linking these basins (Figure 3). Northwest striking lineaments apparent in Landsat imagery interpreted by Chorowicz and Mukonki [1980] as late Cenozoic strike-slip faults generally correspond to lithologic contacts within this prerift fold belt [Villeneuve, 1980] (Figure 3).

TIMING OF CRUSTAL MOVEMENTS

A structural control on volcanic occurrences is evident throughout the development of the Kivu and Rusizi basins, as eruptive centers for mid-Miocene to Quaternary flows coincide with Late Cenozoic faults. Recognizable centers for late Miocene-Pliocene volcanic eruptions are along the margins of tilted fault blocks in the Kivu and Rusizi basins near the accommodation zone and along the southern tips of the East and West Kivu border fault segments (Figures 3 and 4). Pleistocene-Holocene volcanic centers also coincide with the southern East and West Kivu border fault segments and faults bounding Idjwi Island, indicating a repetitious and interactive relation between volcanism and faulting. . Similarly, observations within other volcanically active rift zones and in mechanical models of dike injection indicate that magmatic processes can lead to slip along existing faults and the initiation of new faults [e.g., Rubin and Pollard, 1988].

Kinematic constraints described above and summarized below indicate that the East Kivu basin formed prior to the development of the West Kivu and Rusizi border fault systems and that the accommodation zone between the Kivu and Rusizi basins developed as faults bounding these three basins propagated to the north and south.

1. The first-stage tholeiitic volcanism (Tv_1; 10-6 Ma) is limited to the East Kivu basin and along the southern part of Idjwi Island, which suggests that the East Kivu border fault segment formed prior to the development of the West Kivu border fault segment (Figure 12). The spatial occurrence of Tv_1, the lack of significant thickness variations, and directional flow indicators suggest that the tholeiitic basalts were erupted onto a fairly level surface (De Paepe and Fernandez-Alonso, [1981], Pasteels et al., [1989], and this study).

2. Eruptive centers for the second-stage alkalic volcanism (8.0-4.1 Ma; Figure 5) within both basins indicate that faulting within the West Kivu region began during latest Miocene or early Pliocene time. Basalts of the third stage

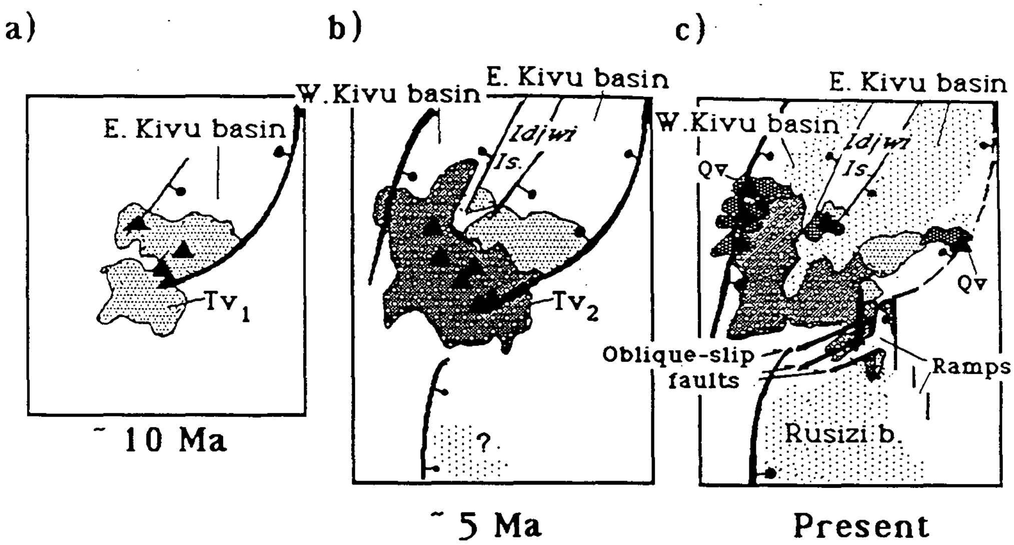

Fig. 12. Schematic diagram illustrating development of the Kivu and Rusizi basins and accommodation zone. Interpretations constrained by structural and stratigraphic data described in text. Triangles, volcanic centers; shaded pattern, volcanics; stippled pattern, late Cenozoic sediments. (a) In mid-Miocene time tholeiitic basalts (Tv_1) erupt onto a surface of low relief from centers near the East Kivu border fault. (b) During late Miocene time the curvilinear East Kivu border fault segment propagates to the north and south. Between 8 and 6 Ma, alkalic basalts (Tv_2) erupt from centers within the East and West Kivu basins, and the northwestern margin of the Rusizi basin is faulted between 6 and 4 Ma. By ~3 Ma, clastic sediments are deposited within all three basins. (c) During Pleistocene time Rusizi border faults propagate to the north, and the East Kivu and Rusizi border fault segments link via oblique-slip transfer faults. Transfer faults also accommodate elevation differences between basins and their uplifted flanks. Rapid subsidence and volcanism occurs along the West Kivu border fault segment in Holocene time.

(<1.9 Ma), erupted from cones along the West Kivu border fault system, flowed down an existing escarpment; thus faulting and subsidence occurred between 8 and 1.9 Ma in the West Kivu basin.

3. Several observations indicate that subsidence within the West Kivu basin postdates mid-Miocene to early Pliocene magmatism within the South Kivu province. First, alkalic basalts (Tv_1) directly overlie metamorphic basement along the eroded footwall of normal faults within the West Kivu basin, and no sediments have been found beneath or intercalated with basalts. Second, conformable contacts and minor thickness variations between Miocene-lower Pliocene basalt flows indicate that extensional faulting occurred near the end or after the second stage of volcanic activity (8-4 Ma). Third, Wong and Von Herzen [1974] suggest that the northern part of the West Kivu basin formed approximately 3 Ma, based on an extrapolation of present-day biogenic sedimentation rates to the 500 m of sediments found in the northern and eastern basins of Lake Kivu. Fourth, Degens et al. [1973] note an absence of volcanogenic sediments in cores from beneath the southwestern arm of Lake Kivu and suggest that subsidence in southern Lake Kivu commenced after volcanic activity (~ 4 Ma).

4. Faults that cut conformable upper Miocene-lower Pliocene flows (Tv_2) within the northern Rusizi basin and the Bugarama graben, and the absence of upper Pliocene-Holocene flows within the Rusizi basin, indicate that the northern Rusizi basin formed after 6 Ma and before ~2 Ma. Although few fauna are found within exploratory drill data, more than 1500 m of Pliocene or younger sediments have accumulated within the Rusizi basin, and these sequences onlap or are interbedded with upper Miocene basalts at the northern end of the basin (D. Stone, personal communication, 1987). The narrow Bugarama graben may have opened after initial subsidence within the Rusizi basin, as Precambrian basement and Tv_2 are overlain by less than 50 m of alluvium.

5. Uplift along the flanks of rift basins occurred in Late Pleistocene time, based on the geometry of terraces of Pleistocene lacustrine sediments. Narrow terraces along the East Kivu border fault segment are elevated 300-600 m above compositionally similar, and probably correlative, sequences within lake basins. Lacustrine sediments are not found at the same or lower elevations along the shores of Lake Kivu, so that the Gisakura terrace probably represents local uplift of the rift flanks, rather than rapid subsidence or a much higher paleolake level. Sequences are tilted to the northwest and thicken away from border faults or in the opposite sense expected if they had been deposited at their present elevation in a narrow lake along the hanging wall of a border fault (Figure 8). Thus the Gisakura terrace was uplifted at a minimum rate of 0.2 cm/yr, assuming that these lacustrine sediments and basalts (Qv) were contiguous with sequences beneath Lake Kivu. This Pleistocene uplift along the flanks of the rift effectively narrows the zone of subsidence with time. Uplifted terraces of lacustrine sediments also are found along the eastern (monoclinal) side of the Rusizi basin and along other Western rift border faults [Pouclet, 1977].

6. Several observations indicate that the West Kivu border fault segment has been more active than the East Kivu border fault segment during Holocene time. Faults bounding the western side of the Kivu basin appear more youthful than faults bounding the eastern side, the western scarp has been

more active seismically, and Pleistocene to historic age cinder cones are aligned along the southwestern West Kivu escarpment [Wohlenberg, 1968; Guibert, 1977a]. Using a Holocene lake level highstand as a horizontal datum, the West Kivu border fault segment subsided at a rate of 5 cm/yr during the past 6 Ka as Idjwi Island was uplifted at a minimum rate of 0.8 cm/yr: beach deposits dated at 6 Ka were found at 300 m water depth in the West Kivu basin; hyaloclastites on Idjwi Island interbedded with shallow water diatomites dated at 10-12 Ka now are elevated approximately 80 m above the lake level high stand [Degens et al., 1973; Guibert, 1977b].

Summarizing these constraints on the timing of crustal movements, tholeiitic volcanism and faulting along the East Kivu border fault segment began in mid-Miocene time (Figure 12a). Alkalic volcanism within the West Kivu and East Kivu basins began in late Miocene time and continued to early Pliocene time (12b). Faults bounding the northwestern Rusizi basin formed during early Pliocene time (12b) and propagated northward during Plio-Pleistocene time (12c). By Pleistocene time, ENE striking transfer faults and north-south striking faults bounding the Bugarama graben linked the eastward dipping East Kivu border fault segment and the westward dipping Rusizi border fault segment (12c). The West Kivu border fault segment has served as the master fault for crustal extension during Quaternary time when tholeiitic and alkalic basalts were erupted along both margins of the Kivu rift (12c).

SUMMARY OF OBSERVATIONS AND INTERPRETATIONS

Observations

Summarizing observations of border fault and accommodation zone geometries within the Kivu and Rusizi basins and illustrated in Figure 4 are the following:

1. High-angle (45°-75°) normal faults separated by 1-5 km occur in a 10- to 15-km-wide zone along one side of asymmetric basins. These planar border fault systems with large throws (1-6 km) may penetrate the crust. The opposite side of basins regionally has the form of a monocline or a ramp. The geometrical arrangement of dip-slip and oblique-slip faults along the length of basins produces discrete segments of the border fault system that are curvilinear in plan view. Throws are greatest near the central part of basins at the base of the escarpment and decrease toward the tips of ~100-km-long basins. Thus sedimentary basins are spoon-shaped, although magnitudes of subsidence vary significantly from basin to basin.

2. Where border fault segments overlap along the length of the rift, dip-slip faults striking north-south and oblique-slip faults striking ENE accommodate differential throws and horizontal offsets between extensional basins. Oblique-slip faults and relay ramps also accommodate regional variations in topographic relief related to the broad domal uplift of the East African plateau.

3. Volcanic centers within the South Kivu volcanic province occur within the comparatively high-strain accommodation zone and between en echelon border faults at the tips of border fault segments. However, a progressive shift in eruptive centers to the center of rift basins [e.g., Pasteels et al., 1989] is not found, as Quaternary eruptive centers are located along the margins of Kivu basins and along faults bounding the Idjwi Island horst.

4. Three structural relations consistently observed within the Kivu and Rusizi regions indicate that zones of crustal thinning are limited to rift basins bounded by approximately 100-km-long border fault segments. First, few faults occur on the uplifted flanks outside inward facing normal faults bounding the rift valley. Second, oblique slip transfer faults do not appear to extend outside the rift valley bounded by inner-facing border faults. Third, the absence of eruptive centers along the rift flanks reflects a lack of faulting along the uplifted margins of basins. Although initial flows that preceded major fault movements covered a wider region than the present-day rift basins, eruptive centers are restricted to rift basins within the inward facing normal faults bordering basins.

5. The Western rift system generally follows older orogenic belts and avoids the Archaean cratons. At the length scale of border fault segments, however, the orientations of Miocene-Recent faults generally extend across older contacts between tectonic units or crosscut structures in metamorphic basement. Where the orientations of prerift faults and metamorphic fabrics trend approximately north-south, or subparallel to late Cenozoic faults, there is little evidence for reactivation of older structures enhanced by differential erosion along the uplifted rift flanks.

Interpretations

1. Differences between basins in the age of initial volcanism, faulting, subsidence, uplift, and composition of magmatic products suggest that border fault segments propagated in a roughly north-south direction. The location of Miocene and Quaternary volcanic centers along Quaternary faults paralleling border faults indicates that the general orientations of border fault segments undergo little change between successive episodes of rifting. Thus the along-axis propagation and crosscutting geometry of transfer faults contributes to the along-axis segmentation of the Western rift valley.

2. The atypical full-graben morphology of the Kivu basin may be caused by border fault propagation. Although the East Kivu border fault segment developed before the West Kivu border fault segment, the West Kivu border fault segment has served as the master fault during Quaternary time. This interpretation alleviates the complicated detachment fault geometries at depth beneath the rift valley required by simultaneously active structures on both sides of the rift valley [e.g., Mohr, 1987; Morley, 1988].

3. Uplift of the rift flanks after initial subsidence, as well as progressive collapse of the hanging wall, effectively narrows rift basins.

Spatial and temporal patterns of crustal movements observed within the Kivu and Rusizi basins also are predicted by recent thermomechanical models of extensional basin formation. In one simple shear model, extension along planar faults leads to an asymmetric pattern of uplift along the flanks of the rift, with the zone of maximum uplift located along the hanging wall above the intersection of the shear zone with the surface [Buck et al., 1988]. Models of rifting processes also predict a progressive narrowing of the zone of maximum subsidence with uplift along the rift flanks [Moretti and Froidevaux, 1986; Buck et al., 1988].

Transfer faults linking high-angle, planar border faults bounding Western rift basins contrast with transfer faults accommodating largely horizontal offsets between low-angle, shallow detachments within the Basin and Range province [e.g., Burchfiel and Stewart, 1966; Gibbs, 1984], with these differences perhaps due to the thermal structure of the lithosphere [e.g., Smith and Bruhn, 1984]. However, patterns of along-axis rift valley segmentation and alternating basin asymmetries similar to those observed in the South Kivu province are found in other continental rift systems that have developed atop plateaux (e.g., Baikal rift; Logatchev et al., [1983]) as well as rifts that have formed at or near sea level (e.g., Gulf of Suez; Chenet and Letouzey, [1983]). Thus observations within the youthful Kivu and Rusizi basins provide kinematic constraints on the along-axis segmentation of a wide class of continental rifts.

APPENDIX

The Landsat 5 Multi-Spectral Scanner (MSS) image used in this study covers a 170 km by 185 km region (E-50870-07370). Pixel to pixel resolution is 32 m, quality of data is excellent, and cloud cover was less than 10%. Standard radiometric and geometric corrections were made at the EROS processing center, and images are displayed using a space oblique Mercator projection that preserves length and angular relations. False color composite images corresponding to Scene 1 were generated from reflectance data in bands 2 (0.6-0.7 μm), 3 (0.7-0.8 μm), and 4 (0.8-1.1 μm), displayed as blue, green, and red, respectively.

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