Palaeogeography, Palaeoclimatology, Palaeoecology, 78 (1990): 55–69
Elsevier Science Publishers B.V., Amsterdam — Printed in The Netherlands

The Zambezi River: tectonism, climatic change and drainage evolution

CHRIS NUGENT

Department of Geography, University of Zimbabwe, P.O. Box MP 167, Mount Pleasant, Harare (Zimbabwe)

(Received October 17, 1988; revised and accepted September 5, 1989)

Abstract

Nugent, C., 1990. The Zambezi River: tectonism, climatic change and drainage evolution. Palaeogeogr., Palaeoclimatol., Palaeoecol., 78: 55–69.

The longitudinal profile of the Zambezi River forms two concave-upwards sections, with their boundary at the Victoria Falls. This form has been ascribed to the process of pediplanation and the Victoria Falls identified as one of several knick points that have traversed the river since the breakup of Gondwanaland. An alternative model explains the river's long profile by suggesting that the Upper and Middle Zambezi evolved as entirely separate river systems, which only joined together in comparatively recent times.

The alluvial sequence of the Middle Zambezi is described and interpreted in terms of the latter hypothesis. The river capture event caused a change in grade and is marked by a deposit believed to record a cataclysmic flood. Capture is inferred to have resulted from overtopping of the lake that formed the end sink of the proto-Upper Zambezi, rather than from headward erosion of the proto-Middle Zambezi. This event is dated, from archaeological evidence, to the peak of the last interglacial, at the end of the Middle Pleistocene. Subsequent drainage diversions between the Middle Zambezi and the Kalahari are interpreted as the product of rifting of the Chobe Graben and aggradation of the Chobe Swamps.

River capture by overtopping implies high rainfall over Central Southern Africa at the peak of the last interglacial, which is contrary to predictions that Africa's rain belts then lay north of their modern mean positions. This anomaly is resolved by postulating a southern polar warming episode at that time and agrees with suggestions that the last interglacial was marked by large scale ablation of the West Antarctic Ice Sheet.

Introduction

The Zambezi is the largest river in Southern Africa, flowing some 2575 km from northwestern Zambia to the Indian Ocean northeast of Beira, Mozambique. The catchment drains an area of 1.32×10^6 km^2 (Williams, 1987), including parts of eight countries (Fig.1). From its source, the Zambezi flows into Eastern Angola, swelling rapidly through this high rainfall area (Mwinilunga averages about 1400 mm a^{-1} and Balovale about 1100 mm a^{-1}). The headwater tract crosses numerous rapids, as the river descends to its plateau tract, downstream of the Chavuma Cataract (Wellington, 1955).

The Zambezi's plateau tract crosses exceptionally flat terrain, divided into three "terraces" (Verboom and Brunt, 1970). The lowest, *Bulozi* terrace fringes most of the drainage channels and is flooded annually, during and shortly after the summer rains (November to March). The greatest expanse of Bulozi bounds the Zambezi and forms the Barotse Plain (Peters, 1960). Downstream of Gonya Falls the river's course steepens, crossing several rapids before entering the Chobe Swamps at Katima Molilo.

0031-0182/90/$03.50

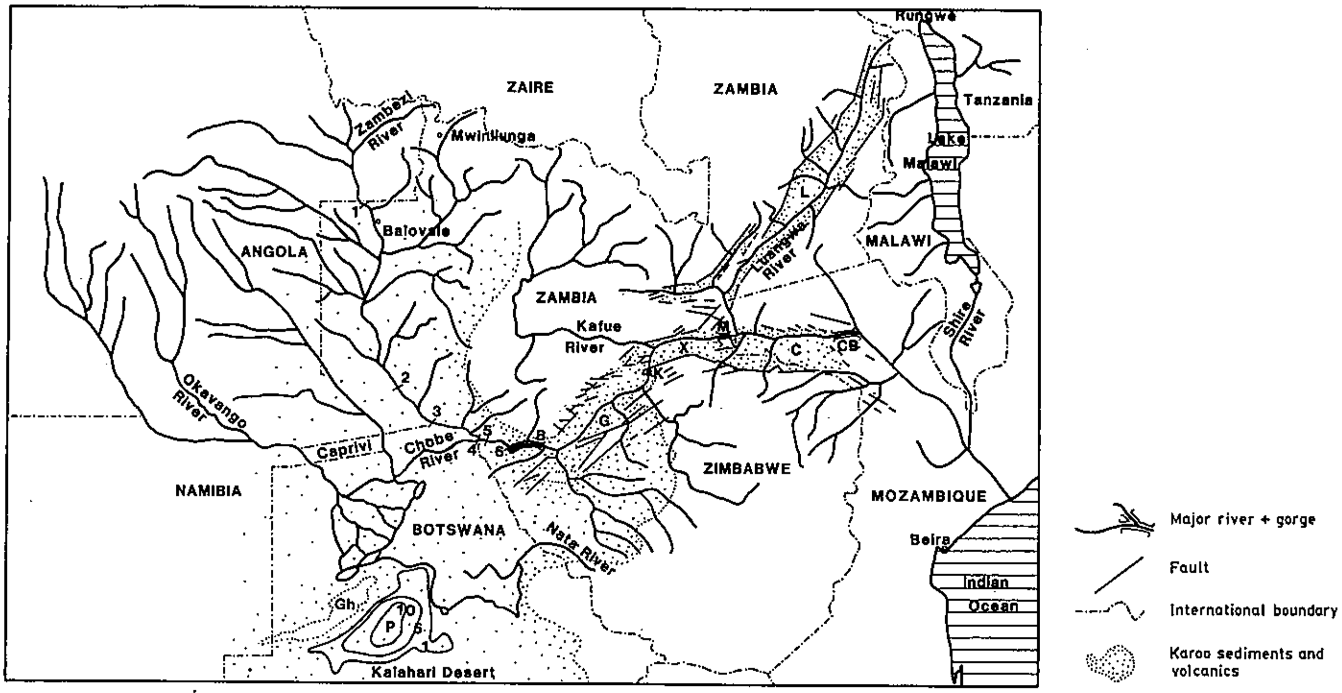

Fig.1. The catchment of the Zambezi and associated internal drainage network and their relationship to Karoo sedimentary basins.

Note: The distribution of Karoo in the west of the map is tentative, as it is mainly overlain by Cenozoic cover. The structure and Karoo of the Shire Rift and Zambezi Graben in the east of the map have been omitted. Fracture pattern mainly simplified after Vail (1967).

Rift structures: *P*=Passarge basin (depth to basement, km). *Gh*=Ghanzi Ridge. *G*=Middle Zambezi basin (Gwembe Trough). *C*=Lower Zambezi basin (Chicoa Trough). *L*=Luangwa basin. *X*=Mana Pools basin.
Gorges: *B*=Batoka Gorge. *K*=Kariba Gorge. *M*=Mupata Gorge. *CB*=Cabora Bassa Gorge.
Rapids and falls: *1*=Chavuma. *2*=Gonya. *3*=Katima Molilo. *4*=Mambova. *5*=Katombora. *6*=Victoria Falls.

The Zambezi's course steepens at the Mambova Rapids and again at the Katombora Rapids, where the river leaves the swamps, passing through a narrow gap in basalt hills. Downstream of Katombora, the Zambezi flows rapidly over the last part of its plateau tract before dropping some 100 m at the Victoria Falls. Downstream of the Falls the river continues to descend steeply within Batoka Gorge, loosing a further 300 m of height in 150 km, before entering the Gwembe Trough and Lake Kariba. The Zambezi's trough tract flows through broad rift valleys, divided by gorges at Kariba, Mupata and Cabora Bassa, which marks the start of the river's final tract across the Mozambique coastal plain.

The longitudinal profile of the Zambezi does not exhibit the characteristic single concave-upwards form (Leopold et al., 1964), in which stream gradient is progressively reduced through "youthful", "mature" and "senile" stages. In general terms, the river to Cabora Bassa Gorge consists of two such profiles, the Upper and Middle Zambezi, with their boundary at the Victoria Falls. This morphology has been explained in two different ways.

(1) In terms of the theory that the process of pediplanation has played a dominant role in the geomorphological evolution of Southern Africa (e.g. Dixey, 1942, 1955; King, 1949, 1962; Lister, 1976, 1987). The Victoria Falls is then seen as a knickpoint between erosion surfaces, one of several that have migrated up the Zambezi River since continental breakup, in Late Mesozoic times.

(2) In terms of the theory that the Upper and Middle Zambezi previously formed two separate rivers, with the Middle Zambezi draining to the east coast and the Upper Zambezi draining (possibly via the fossil channels of Verboom,

1974) to an end sink in the Kalahari (e.g. De Swart and Bennet, 1974; Thomas, 1984a). This theory postulates that the two rivers evolved separately, becoming graded to their respective base levels. The fact that two concave-upwards longitudinal profiles are so well preserved, suggests that they joined together in comparatively recent times. The theory assumes that the Middle Zambezi River underwent major rejuvenation only once, associated with continental breakup.

The Zambezi's alluvial sucession downstream of the Victoria Falls is described and found to be consistent with the latter hypothesis. The nature, chonology and possible causes of river capture are suggested and implications for present stability and future trends are examined.

Regional rifts and rifting

The plateau and trough tracts of the Zambezi River traverse continental rift structures, which were active during Karoo and Tertiary to modern times. Karoo (Permian to Jurassic) basins are developed along a SW–NE trending axis between the Central Kalahari and the Rungwe Volcanic Field, at the northern tip of Lake Malawi (Fig.1). Karoo sediments within the Middle Zambezi and Luangwa Basins have been largely exhumed to form the Gwembe and Luangwa Troughs. In contrast, the Karoo sequence of the Passarge Basin, bounding Proterozoic basement on the Ghanzi Ridge, is buried beneath Cretaceous to Holocene sediments of the Kalahari System (Tankard et al, 1982) and the basin's existence is inferred from geophysical data (Jones and Key, 1978). This lineation is believed to have undergone extensional development during Karoo times (Orpen et al, 1990), through the process of "passive" rifting (Sengor and Burke, 1978; Keen, 1985). Downstream of Kariba Gorge the Zambezi River swings to the east, crossing the Mana Pools Basin to Mupata Gorge. Beyond this, the river flows over Karoo and Cretaceous sediments of the E–W trending Lower Zambezi Basin. This rift is believed to have developed transcurrently, contemporaneous with the bounding Karoo graben (Orpen et al., 1990).

More recent rift structures underlie parts of the Zambezi's plateau tract and much of the Kalahari of Botswana. Rifts were first identified beneath the Okavango and Chobe Swamps by Du Toit (1926), during his investigation into the feasibility of diverting the Zambezi to flood the ancient lake basins of the Kalahari. The basins (Fig.2) have since been shown to be seismically active (Reeves, 1972; Hutchins et al., 1976; Scholz et al., 1976). The Okavango and Chobe and the smaller basins of Ngami and Mababe are fault bounded on their southeastern margins (Fig.2). The sedimentary infill of the Mababe thickens towards the southeast (Shaw, 1985). Detailed surveying and observations by Du Toit (1926) show the surface of the Chobe Swamps to dip gently in that direction. The structure and topography of these basins are thus typical of graben produced by passive rifting (McKenzie, 1978; Wernicke and Burchfiel, 1982; Gibbs, 1984), with listric dislocation zones bounding their southeastern margins.

The crustal structure beneath the Makgadikgadi Pans suggests that this basin also has a tectonic origin (Baillieul, 1979) and fault scarps have been identified on the southern and eastern margins of the pans complex (MacGregor, 1930; Cooke, 1980). The regional record of dispersed earthquakes (Reeves, 1972) and the subdued topography over several discrete depocentres (Cooke, 1979) suggest that the Makgadikgadi basin is now in its sag phase, developing predominantly by thermal contraction (Watts et al, 1982; Cochran, 1983).

Climatic change

Sediments and landforms of the Kalahari record a complex Cenozoic evolution, during which conditions have been both wetter and dryer than at present. Unconsolidated Kalahari Sands, which are of predominantly aeolian origin (Bond and Fernandes, 1974; Thomas, 1987), underlie some 2.5×10^6 km^2 from the Orange River to beyond the Zaire

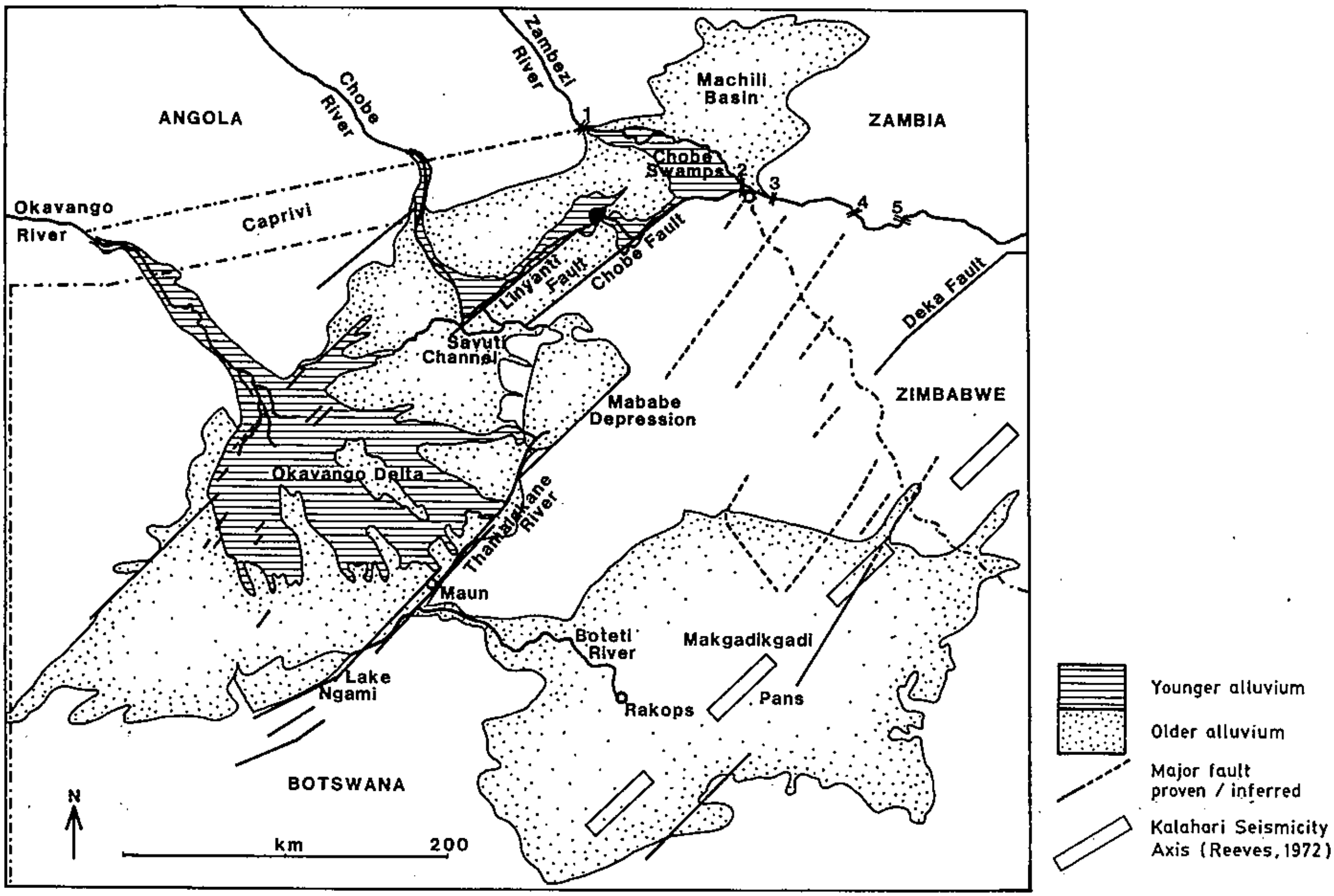

Fig.2. The distribution of alluvial and lacustrine sediments in Central Southern Africa, the basins of Greater palaeo-lake Makgadikgadi. Mainly after Mallick et al. (1981) and Shaw and Thomas (1988).
Major rapids and waterfalls: *1*=Katima Molilo. *2*=Mambova. *3*=Katombora. *4*=Victoria Falls. *5*=Chimamba.

River (Cooke, 1964), implying widespread aridity since at least the end-Miocene (Lancaster, 1984). The surface of much of the sand is formed into longitudinal dunes (Lancaster, 1980, 1981), whose alignment and distribution suggest more than one period of aridity, reaching some 2400 km north of the modern limit of aeolian activity (Thomas, 1984b). Late Glacial conditions some 1.5–2 times wetter than at present are inferred from small pans in the Kalahari (Lancaster, 1978, 1979) and cave deposits to the west of the Okavango Swamps (Cooke, 1975) also show evidence of moist conditions at that time. The levels of palaeolakes over the Makgadikgadi Pans have been used to infer past climatic conditions (e.g. Street and Grove, 1976, 1979; Heine, 1982). Unfortunately, these lake levels would have been influenced by conditions elsewhere in the catchment (Cooke, 1984) and by tectonism (see below), so constitute an unreliable indicator of local palaeo-climate.

Although sediments and landforms indicate wide departures from the present pattern of precipitation in Southern Africa, the dearth of well dated sites of unequivocal climatic significance has prevented the establishment of a reliable chronology of change. In a recent review paper Shaw and Cooke (1986) establish a general chronology over the past 20,000 years. Far from reflecting the northern hemisphere Late Glacial and Holocene sequence, of generally dry conditions becoming wetter at around 12,500 yr B.P. (Street-Perrot et al., 1985), the Southern African record shows evidence of wet conditions during the Late Glacial, becoming dryer when the northern hemisphere tropics became wetter. This suggests that Pleistocene rainfall in Africa was influenced more by latitudinal shifts of the rain belts than by changes in their activity.

The mean positions of the inter-tropical rain belts, the equatorial trough (ET) and monsoonal belt, are governed by the relative strength of the atmospheric circulations in each hemisphere, resulting from the thermal gradients between high and low latitudes (Nicholson and Flohn, 1980). By modelling these gradients, Harrison et al. (1984) confirm that the ET can be expected to have moved southwards during glacial events, when ice sheets accreted over large areas of the northern hemisphere land masses. Under conditions warmer than at present, during the Holocene climatic optimum, the ET is expected to have been located somewhat north of its modern mean annual position (6°N). This prediction accords well with the record from Northeast Africa, where vegetation zones then lay some 4° north of their modern latitudes (Ritchie and Haynes, 1987).

Drainage evolution

The rifted basins of the Kalahari are overlain by alluvial and lacustrine sediments extending over some 118,000 km^2 (Fig.2). A series of broadly concentric strandlines within the Makgadikgadi partially enclose former lake stands at 945, 920, and 912 m, with localised features at around 908 and 904 m (Cooke, 1980). Similar features around the Ngami and Mababe Basins lie at 940–945 m (contiguous with Lake Makgadikgadi), 936 m (overflowing into Lake Makgadikgadi) and at lower levels (Shaw, 1985).

At its highest (945 m) stage, the lake would inundate approximately half the area of alluvial and lacustrine sediments (60,000 km^2; Grey and Cooke, 1977). The inflow necessary to sustain a lake at that level over the Makgadikgadi (34,000 km^2) under present climatic conditions is calculated by Grove (1969) as 50 km^3 of water annually. By extension, the greater lake area that inundated all the basins to 945 m (Greater palaeolake Makgadikgadi) would have required an annual inflow of some 88 km^3. It should be emphasised that this estimate is very approximate, assuming that the lake area has not changed (e.g. by rifting, isostatic adjustment or alluviation), ignoring water loss by seepage and assuming evaporation under a climate similar to that of the present. Modern discharges of the main rivers entering the palaeo-lake and mean precipitation are listed in Table I. Under the current climatic regime, inflow to the basins would be almost sufficient to maintain Greater lake Makgadikgadi at its highest stand. Even the lowest of the fossil shore lines has not been attained in historical times, presumably because the discharge of the Zambezi and Chobe Rivers and associated local rainfall have drained out of the basins.

That the Upper Zambezi once supplied this lake is demonstrated first by the fact that the river passes through the Chobe Swamps, part of the greater palaeo-lake (Lake Caprivi; Shaw and Thomas, 1988). Secondly by the water budget (Table I), which shows that other sources of inflow would have to more than double, presumably over a prolonged period, to sustain the lake at its maximum stage. Under present topographic conditions, a lake could inundate the Makgadikgadi and Okavango basins to about 936 m, controlled by an outflow along the base of the Chobe fault scarp (Shaw

TABLE I

The mean annual inflow to Greater palaeo-lake Makgadikgadi under present climatic conditions

Input (source)	Annual inflow (km^3)
Okavango River[1]	11
Chobe (Mashi) River[2]	3
Zambezi River[3]	42
Precipitation over lake[4]	27
Total	83

[1] Wilson and Dincer (1976).

[2] Shaw (1985).

[3] Ministry of Water Development discharge records from Victoria Falls, hydrological years 1924/25 to 1986/87.

[4] Lake area of 60,000 km^2 with mean annual precipitation of 450 mm.

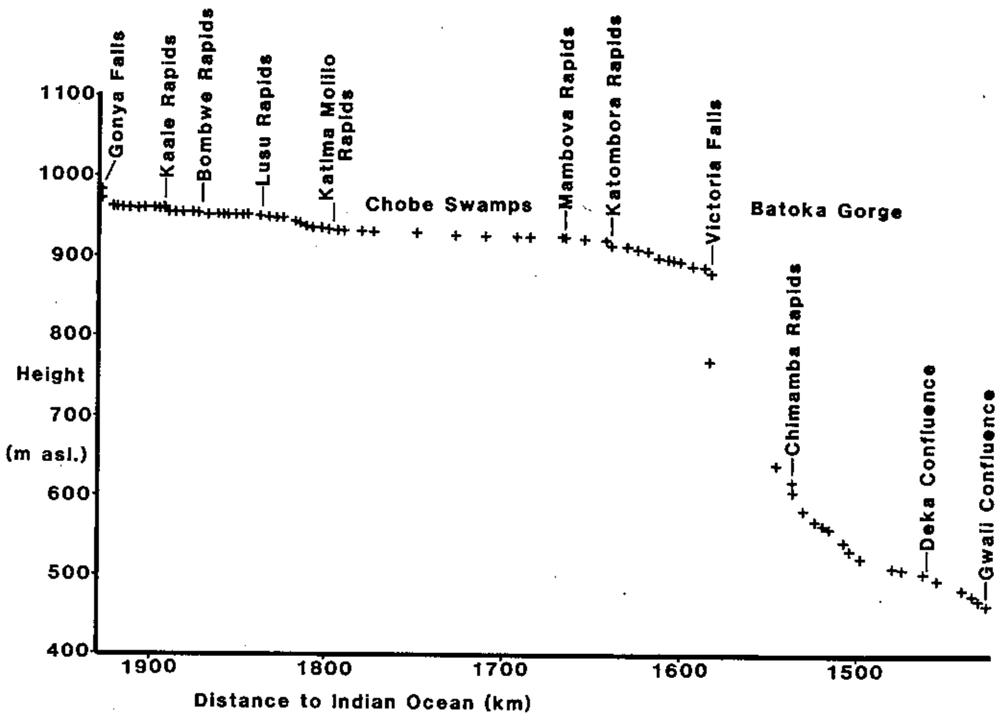

Fig.3. The longitudinal profile of the Zambezi River, Gonya Falls to Batoka Gorge.
Note: Heights and distances extracted from regional mapping, after Du Toit (1926) and Harrison and Coppinger (1962). Downstream of Victoria Falls, data compiled from spot heights on 1:50,000 topographic maps published by Surveyor General, Harare. These latter data are subject to error, since spot heights may not be measured to water level, which varies considerably within the gorge.

and Thomas, 1988). The Chobe and Machili Basins could not support a lake at all, as they drain into the Zambezi. The downstream limit of ancient alluvium (Fig.2) coincides with a major steepening of the gradient of the Zambezi (Fig.3) at the Katombora Rapids. At this point, the Zambezi passes through a narrow gap in basalt hills (Fig.4), lying 3 km north of a saddle at between 940 and 960 m. It is proposed that, prior to the capture of the Upper Zambezi, the Katombora Gap was a similar saddle feature at or slightly above 945 m a.s.l.

Landscape evolution

Interpretation of river capture on the Zambezi is inextricably linked to our understanding of landscape development within the catchment. A distinctive feature of Southern African scenery are the many surfaces, developed at different heights on a diverse range of rock types. Surfaces are separated by scarps, which often do not appear to coincide with any structural or lithological boundary (King, 1963; Lister, 1987). The peneplanation model of Davis (1922) envisages simultaneous landscape denudation over large areas, making it difficult to account for multiple surfaces at different levels.

The need to explain Southern African planation surfaces led to the development of models of landscape denudation by pediplanation (summarised by Partridge and Maud, 1987). Pediplanation invokes parallel retreat of slopes, away from a newly rejuvenated coast or drainage line. The higher, older surface is gradually destroyed as a pediment and finally a pediplain is developed at the base of the retreating scarp. Multiple surfaces may thus be generated by repeated changes in base level. It has been suggested that drainage rejuvination occurred repeatedly since the breakup of Gondwanaland, during episodes of tectonic warping of Southern Africa (King, 1955). Paral-

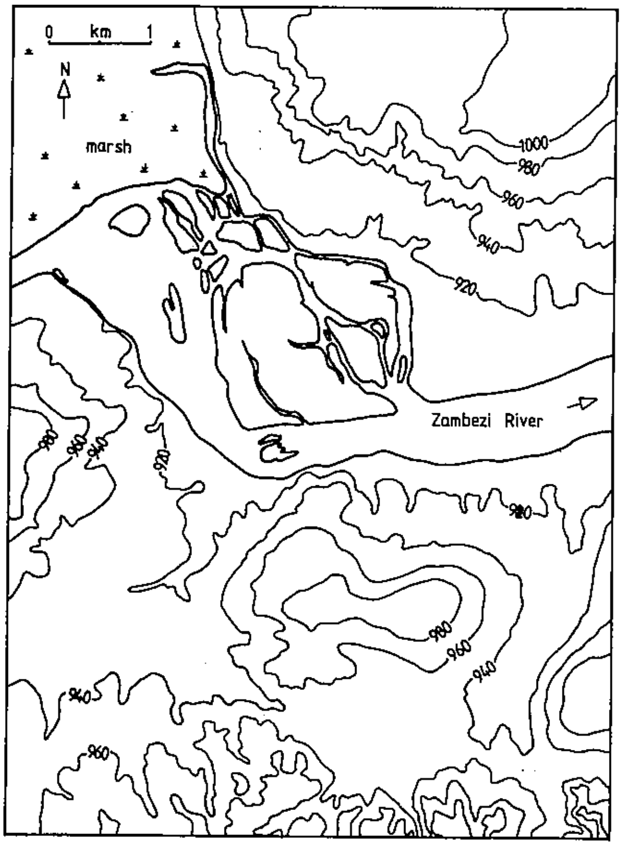

Fig.4. The Katombora Gap. This is believed to mark the point where Palaeo-lake Greater Makgadikgadi overtopped at the 945 m level, flowing across the basalt plateau and into the proto-Middle Zambezi River. Note the saddle about 3 km south of the river at between 940 m and 960 m a.s.l. The marsh in the northwest continues upstream and connects with the Chobe Swamps.

lel retreat is demonstrated on slopes in Natal in which a hard layer of dolerite overlies softer Karoo sediments (Fair, 1947; 1948). The model is extended to all slopes in Natal and the planation surfaces dated in terms of coastal deposits (Fair and King, 1954).

The pediplanation model readily explains the regional morphology above and below the Victoria Falls. The surface above the Falls may be seen as the older, being gradually eroded and (eventually) replaced by pediments, which are developing at a lower level downstream. River capture by headwards erosion should then be seen as the inevitable consequence of knickpoint retreat and may well have occurred during the erosion cycle prior to that now represented by the Victoria Falls. Extension of the pediplanation model from coastal Natal to the African interior presents certain difficulties. It assumes that a knickpoint initiated at the coast can retain its character and actually increase its height after retreating almost 1600 km upstream, into lithologies which lack the structural control considered by Fair (1947) to be an important pre-requisit of parallel slope retreat in Natal.

An alternative hypothesis for the development of the Victoria Falls does not assume that processes of pediplanation have operated and landscape denudation may have occurred in a manner more akin to the Davisian model. The major drop in height at the Falls is seen as the consequence of the joining of two rivers which were previously graded to different base levels. If parallel slope retreat is not considered to have been important, then river capture is not predicted as a consequence of landscape denudation and must be explained in some other way. The joining of the Upper and Middle Zambezi to form the modern river may constitute a test, which may help to distinguish between these two models of geomorphic evolution.

Evidence from downstream

The record of river capture on the Zambezi is preserved in the alluvial sequence downstream as a change in grade. Most of the channel within the Gwembe Trough is now inundated by Lake Kariba and there is no systematic geological mapping covering the alluvial belt (although excellent 1:25,000 topographic mapping is available). The alluvial sequence and associated Stone Age archaeology are recorded by Bond and Clark (1954), from three sites along the river. They describe gravel terraces rising to 30, 40 and 55 m above the modern (pre-Kariba Dam) channel.

Downstream of Kariba Gorge, at Nyamuomba (Fig.5), gravel covers part of a terrace surface at a little over 30 m above the river. The terrace is dissected and discontinuous but well marked on the Zimbabwe bank as far as Chirundu, downstream of which, such high

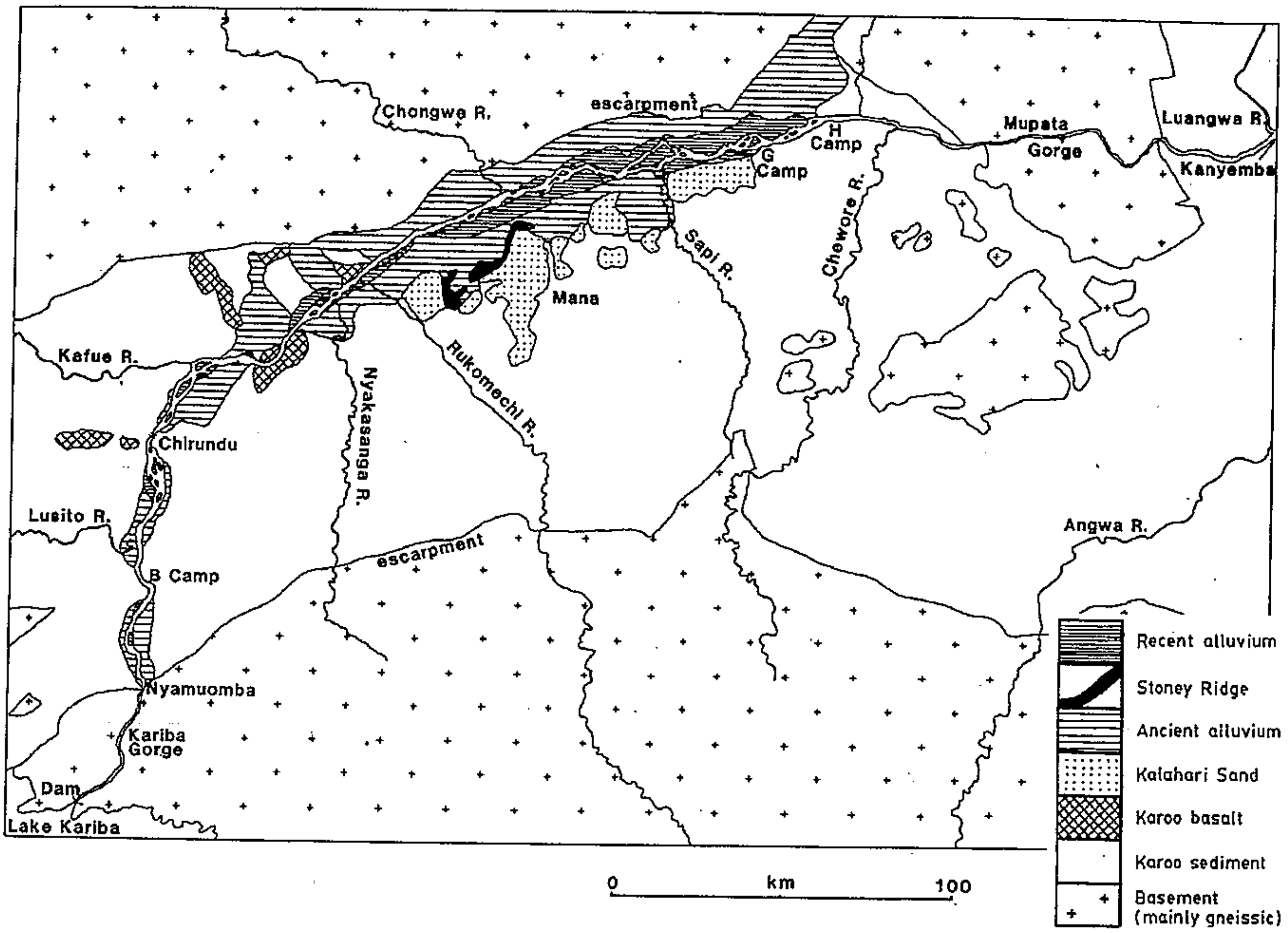

Fig.5. Geology of the Mana Pools basin.

terraces have not been recognised. Alluvium may be identified where it is cut by tributary streams but the surface is sodic or highly sodic (Bennett et al, 1986), extensively eroded by overland and gully flow and colonised by *Colophospermum mopane* or devoid of woody vegetation (Muller and Pope, 1982).

Where it is exposed in cliffs on tributary streams, this alluvium consists of partially consolidated, sandy fining upwards cycles, commonly separated by thin pebble beds spaced about 0.5–1.5 m vertically. The pebble beds contain abundant clasts of jaspilite or banded ironstone, derived from greenstone belts on the bounding cratons (Stagman, 1978). This fixes the provenance of the alluvium to the area upstream of Kariba Gorge, since banded ironstone is not present within the gneissic rocks that make up the Zambezi Escarpment. Other prominant lithologies within the pebble beds include Karoo basalt, rare agates (originally derived from the basalt) and fossil wood of *Dadoxylon* affinity. These fix the age of the alluvium to the post-Karoo period, since the bulk of agate-bearing basalt within the catchment was extruded at the end of Karoo times (Bond, 1967, 1975; Visser, 1984) and *Dadoxylon* fossil wood is developed within the Upper Karoo Pebbly Arkose (Bond, 1973). The alluvium described above is interpreted as a post-Karoo aggradational sequence. It is referred to as "ancient" alluvium, to distinguish it from "recent" deposits. It is inferred, from the grade of the river in recent times, to pre-date capture of the upper catchment.

The Zambezi River downstream of Kariba is controlled in several places by rock which outcrops in the banks and presumably also in the bed (Nugent, 1983). These geological control points are believed to have maintained grade by effectively limiting downcutting along intervening alluvial stretches. The result has been that recent deposits of the river do not extend more than about 5 m above modern water level (no floodgates open at Kariba Dam), representing maximum flood levels under modern grade conditions. The Middle Zambezi thus contrasts with other African rivers (e.g. Helgren, 1978; Butzer, 1980), which responded to Late Pleistocene and Holocene climatic change by alternately aggrading and degrading their beds to produce terraces.

Grade is controlled by the relative supply of water and sediment to the river channel (Lane, 1955). The ancient aggradational sequence indicates that the supply of sediment was then greater than the supply of water could remove from the system. At capture, the supply of water was massively increased (the upper catchment contributed seven times the flow of the lower catchment at Kariba between October 1947 and September 1958; Nugent, 1986). The supply of sediment, however, increased hardly at all. The swamp tracts of the upper catchment form such efficient sediment traps that an estimated 0.3 g of sediment per liter of water pass over the Victoria Falls at high flow (Central African Council, 1951), very much less than is transported by the sediment laden rivers of Lake Kariba's lower catchment. In terms of the relations of Lane (1955), capture altered the discharge of water and sediment in such a way as to promote degradation. It is inferred from the height of recent deposits that the river degraded to bedrock following capture and has not aggraded since, by more than a few tens of centimeters (Nugent, 1986).

The nature of capture

As the river is believed to have undergone aggradation prior to capture and degradation since, it follows that any deposit remaining from the time of capture will lie at the topographically highest points in the alluvial sequence. At Mana, ancient alluvium rises to some 50 m above modern river level, where its southern margin bounds and underlies deep acidic sands (Bennett et al., 1986) inferred to be aeolian (Kalahari) sand. A part of this contact is marked by a sinuous ridge, the Stoney Ridge (Fig.5).

Stoney Ridge is composed of generally coarse grained (typically grit to cobble grade), poorly sorted, fairly well rounded clasts, arranged chaotically as a clast supported conglomerate. Bedding has not been detected, although faces exposed in road stone workings near the eastern end of the deposit and a prospectors pit near its centre show poorly defined imbrication. The clasts of the Stoney Ridge deposit, like the underlying pebble beds, include abundant jaspilites. A major difference is that agates within the Stoney Ridge are also abundant, implying a change in provenance in which the area of the catchment underlain by basalt had been significantly increased.

The generally fairly well rounded, large grain size of the Stoney Ridge deposit suggests a high energy alluvial source. The very poor sorting and lack of systematic bedding structures suggest rapid deposition. The Stoney Ridge is interpreted as a cataclysmic flood deposit, whose topographically high location suggests that the flood coincided with capture of the upper catchment. The clast composition, with an increased abundance of agates, is consistent with this interpretation. Agates in the Stoney Ridge are believed to have been derived largely from channel lag and bounding regolith from the basalt plateau upstream of the proto-Victoria Falls (Nugent, 1990) and the proto-Batoka Gorge downstream.

The Stoney Ridge flood deposit suggests that river capture on the Zambezi resulted from overtopping of Greater palaeo-lake Makgadikgadi. Numerous African lakes are inferred to have overtopped during the Late Pleistocene and Holocene (e.g. Grove et al., 1975; Adamson and Williams, 1980; Livingstone, 1980; Durand, 1982; Hastenrath and Kutzbach, 1983; Grove, 1986). It is considered that whereas overtopping is likely to have caused the Stoney Ridge flood, capture by headwards erosion is not, due to the physical properties of the near-surface layer at the watershed at that time.

If headward erosion were the cause of capture then the head of the proto-Middle Zambezi River would have been cutting vigorously into the underlying Karoo basalt, causing its channel to be lined by relatively hard and competent rock. The maximum rate of post-capture degradation at such a connecting channel may be tentatively estimated from the rate of retrogression of the Victoria Falls, calculated as several centimeters or a few tens of centimeters per year (Derricourt, 1976; Nugent, 1990). If capture had resulted from overtopping of the palaeo-lake then no pre-existing channel existed at the watershed, which may be assumed to have been underlain by a superficial layer of weathered regolith. Such loose material would have been rapidly eroded, to produce a cataclysmic flood lasting in the order of several weeks (e.g. Malde, 1968). Such a flood would not be possible if the Middle Zambezi were eroding rapidly headwards and the watershed, at the time of capture, was cut in bedrock.

The chronology of capture

The change in grade that accompanied capture of the Zambezi's upper catchment may be used to date the event, at least in terms of the associated Stone Age stratigraphy. Bond and Clark (1954) describe the distribution of Stone Age artifacts on and within the alluvial terraces of the Gwembe Trough. The higher terraces contain rare artifacts of Early Stone Age cultures, which have been rolled and therefore re-deposited. The surfaces of these terraces are littered with unrolled tools assigned to the *Sangoan* culture, regarded as intermediate between those of the Early and Middle Stone Age (Clark, 1982; Volman, 1984). Some of the lower terraces support artifacts assigned to the first Middle Stone Age cultures (proto-Stillbay; Bond and Clark, 1954). Artefacts are described from within an alluvial sequence (Alluvium II) rising about 30 m above the river. The youngest alluvium, the flood plain of the pre-Kariba channel, underlies only later-Middle and Late Stone Age sites.

It is inferred from this distribution of Stone Age remains that the major degradation on the Zambezi occurred during or immediately following *Sangoan* times and was in progress as the first cultures of the Middle Stone Age began to evolve. Since this degradation is believed to have resulted from and immediately followed river capture, capture can be dated to latest *Sangoan* or earliest Middle Stone Age times. The distribution of Stone Age artifacts downstream of Kariba (Phillipson, 1976; 1977) is consistent with this chronology. The oldest two sites, recording a *Sangoan* occupation, are located at or above the top of the ancient alluvial sequence. A third site (Kasoka), some ten or twelve meters above the river, may record a somewhat later *Sangoan* industry. Middle and Late Stone Age sites are also found on these interfluves and at lower levels over bedrock at Chirundu.

The early part of the Middle Stone Age succession is preserved in wave-cut caves on the South African coast, at Klasies River Mouth (Singer and Wymer, 1982). The earliest cave infill has yielded a single artifact, which is classified as *Sangoan* and is thought to have been contemporaneous with the peak of the last interglacial, at about 125,000 yr B.P. The succeeding Middle Stone Age culture (MSA I) is interpreted (Butzer, 1978; Hendey and Volman, 1986) as contemporaneous with the following marine regression. Archaeological evidence thus links the capture of the Upper Zambezi with high sea level, suggesting that river capture occurred at or near the peak of the last interglacial, the boundary between the Middle and Late Pleistocene.

A possible cause of capture

Assuming that the capture of the proto-Upper Zambezi by the proto-Middle Zambezi resulted from overtopping of Greater palaeo-lake Makgadikgadi, due to an increase in precipitation within its catchment, the last interglacial would seem to have been a most improbable time. According to the relations of Nicholson and Flohn (1980) and Harrison et al., (1984), the inter-tropical rain belts of Africa can be expected to have then lain well north of the equator and probably north of their modern mean positions. High rainfall over North Africa at that time is confirmed by the presence of organic-rich "sapropel" horizons in Nile deltaic sediments (Rossignol-Strick, 1983) and large lakes in the Central Sahara, some time before the interglacial's peak (Gaven et al., 1981).

To move the rain belts southwards, over the catchment of the Upper Zambezi and other southward-draining rivers, would have required either a northern polar cooling or a southern polar warming event. Since this period coincided with high sea levels, associated with global warming and contraction of ice sheets, the latter possibility is suggested. Last-interglacial temperatures over East Antartica are inferred to have been some 2°C warmer than at present (Jousel et al., 1987). Since northern polar temperatures are also inferred to have been of this order, East Antarctic warming alone is not expected to have initiated the global changes required to move the rain belts from the Central Sahara to the catchment of the Upper Zambezi. It has been suggested (Mercer, 1973, 1978; Aharon et al, 1980) that high sea level at that time resulted from ablation of the ice sheet over West Antarctica, which may now be unstable (Weertman, 1974) and disintegrating (Hughes, 1973). It is considered that, since the inferred timing and mechanism of capture support the concept of last-interglacial southern polar warming, major ablation of West Antarctic ice seems to be the most feasible cause.

Late Pleistocene and Holocene drainage evolution

The Makgadikgadi, Okavango and Chobe basins have all held lakes at various times since the last interglacial, which fluctuated in extent and alternated with dry intervals. Bond and Summers (1954) describe a Middle Stone Age site on an ancient Makgadikgadi shoreline at 945 m, which may have been contemporaneous (Cooke, C.K., 1969) with a lake stand at that level between about 40,000 yr B.P. and 35,000 yr B.P. (Cooke, 1984). Late Stone Age sites bordering a former 912 m lake (Helgren, 1984) are consistent with lake stands at that level during the Holocene (Cooke and Verstappen, 1984). These Holocene lakes are suggested (Shaw, 1985) to have been supplied by lakes at 936 m in the Ngami and Mababe depressions and the Southern Okavango (Lake Thamalakane) and the Caprivi Strip (Lake Caprivi; Shaw and Thomas, 1988). These lakes would have required a substantial inflow and imply that the capture of the Upper Zambezi may not have been completed by the end of the last interglacial.

A mechanism whereby repeated switching of drainage could occur after the initial diversion of the river, is suggested by the behaviour of the Savuti River (Fig.2) during the historical period. The first record of the Savuti, by Oswell and Livingstone in 1849, describes the river as flowing strongly into a small lake in the Mababe (Oswell, 1900, p. 240). The Chobe did not spill into the Savuti at all from 1884 (Stigand, 1923) until 1957 (Grove, 1969), since which time it has flowed intermittently (Shaw, 1984). The Savuti offtake lies astride the Linyanti Fault, which is one of the boundary faults of the Chobe graben. It follows that a period of rifting would cause a fault scarp to develop across the offtake, diverting the flow along the Chobe River to the northeast. Conversely, during a period of relative tectonic quiescence, aggradation within the Chobe Swamps would raise water level at the offtake and re-instigate flow along the steeper channel of the Savuti.

A similar situation may have existed at the Mambova Rapids, which lie on an extension of the Chobe Fault (Fig.2). Shaw and Thomas (1988) describe former Lake Caprivi, which was ponded behind that barrier and may have drained, via the base of the Chobe scarp, into Lake Thamalakane. The effect of movement on this extension of the Chobe fault is thus inferred to have worked in the opposite sense to that on the Linyanti Fault, as rifting diverted drainage into the Kalahari and tectonic quiescence and aggradation re-established the link with the Zambezi. It is proposed that the route taken by waters of the Upper Zambezi and Chobe Rivers since the cutting of the Katombora Gap has been governed in a complex way by the rate of rifting, the rate of aggradation (largely controlled by climate) and switching of movement between the two boundary faults on the southeastern margin of the Chobe Graben.

Conclusions and implications

The alluvial stratigraphy of the Zambezi's trough tract is consistent with the hypothesis that the Upper and Middle Zambezi developed as separate rivers, at least until the end of the Middle Pleistocene. This model characterises the Cenozoic evolution of Southern Africa as comprising gradual denudation and exhumation of coastward — draining Karoo basins (troughs) and post-Karoo sedimentation within the internal basins of the Kalahari. This (gradual development) model removes the need to invoke repeated intervals of tectonic warping and drainage rejuvenation to explain the Zambezi's longitudinal profile, constituting a significantly simpler theory of the geomorphic evolution of Southern Africa.

Recognition of the Stoney Ridge as a cataclysmic flood deposit implies that river capture on the Zambezi was initiated by overtopping. This, in turn, suggests that rainfall within the catchment of Greater palaeo-lake Makgadikgadi was increased at the peak of the last interglacial, implying a southern polar warming event at that time. Such changes would be expected to follow major ablation of Antarctic ice and recession of the ice sheet's margin. This record from Southern Africa is thus in agreement with the suggestion (Mercer, 1973, 1978) that the last interglacial was marked by ablation of the West Antarctic ice sheet. This interpretation is pertinent in terms of current predictions (e.g. Joyce, 1988) that global temperatures could increase by up to 3°C over the next century, suggesting that it is Antarctic ice that will be most affected. In addition to the obvious implications for land near sea level worldwide, the effects on inter-tropical climate may be anticipated. Southerly deflection of the mean positions of the inter-tropical rain belts can be expected, resulting in further desiccation of the Sahel region and rainfall increases over tropical Africa south of the equator.

Of regional concern are the implications of Late Pleistocene and Holocene switching of the flow of the Upper Zambezi, between rifted basins of the Kalahari and the Gwembe Trough. If such drainage changes did result, as suggested, from the interplay between rifting and aggradation of the Chobe Swamps, then it follows that the process could be repeated. Although the rate of rifting of the Chobe Graben may not be predicted with any certainty, an increase in sediment discharge (due to increased agricultural utilisation of the catchment) may be anticipated. In view of the extreme economic importance of the possible re-direction of Upper Zambezi waters, to hydroelectric power production and irrigation potential, efforts should be made to understand the cause and process of past drainage changes more fully, in order to be able to predict and possibly mitigate changes in the future.

Acknowledgements

I am grateful to the Zimbabwe Hunters' Association, who financed this work with a postgraduate scholarship and the University of Zimbabwe Research Board, who supplied research funding.

References

Adamson, D. and Williams, F., 1980. Structural geology, tectonics and the control of drainage in the Nile basin. In: M. A. J. Williams and H. Faure (Editors), The Sahara and the Nile. Balkema, Rotterdam, pp. 225–252.

Aharon, P., Chappell, J. and Compton, W., 1980. Stable isotope and sea level data from New Guinea supports Antarctic ice-surge theory of ice ages. Nature, 283: 649–651.

Baillieul, T. A., 1979. Makgadikgadi Pans Complex of Central Botswana. Geol. Soc. Am. Bull., 90: 133–136.

Bennett, J., Anderson, I. P. and Brinn, P., 1986. The soils of Mana Pools National Park. Chem. Soil Res. Inst., Harare.

Bond, G., 1967. A review of the Karroo sedimentation and lithology in Southern Rhodesia. In: 1st IUGS Symp. Gondwana Stratigr., pp. 173–195.

Bond, G., 1973. The palaeontology of Rhodesia. Rhod. Geol. Surv. Bull., 70.

Bond, G., 1975. The Geology and formation of the Victoria Falls. In: D. W. Phillipson, Mosi-oa-Tunya: A handbook to the Victoria Falls Region. Longmans, Zimbabwe, pp. 19–27.

Bond, G. and Clark, J. D., 1954. The Quaternary sequence in the Middle Zambezi Valley. South Afr. Archaeol. Bull., 9: 115–130.

Bond, G. and Fernandes, T. R. C., 1974. Scanning electron microscopy applied to quartz grains from Kalahari type sands. Trans. Geol. Soc. South Afr., 77: 191–199.

Bond, G. and Summers, R., 1954. A late Stillbay hunting-camp site on the Nata River, Bechuanaland Proctorate. South Afr. Archaeol. Bull., 9: 89–95.

Butzer, K. W., 1978. Sediment stratigraphy of the Middle Stone Age sequence at Klasies River mouth, Tsitsikama coast, South Africa. South Afr. Archaeol. Bull., 33: 141–151.

Butzer, K. W., 1980. Pleistocene history of the Nile Valley in Egypt and Lower Nubia. In: M. A. J. Williams and H. Faure, The Sahara and the Nile. Balkema, Rotterdam, pp. 253–280.

Central African Council, 1951. Report on the Kariba Gorge and Kafue River Hydro-Electric projects. Inter-Territorial Hydro-Electric Power Comm. Rep. M.W.D. files, Harare.

Clark. J. D., 1982. The cultures of the Middle Palaeolithic/Middle Stone Age. In: J. D. Clark (Editor), The Cambridge History of Africa. C.U.P., Cambridge, pp. 248–340.

Cochran, J. R., 1983. Effects of finite rifting times on the development of sedimentary basins. Earth Planet. Sci. Lett., 66: 289–302.

Cooke, C. K., 1969. Radiocarbon dates for the Rhodesian Stone Age. Rhod. Prehist., 3: 6–8.

Cooke, H. B. S., 1964. The Pleistocene environment in Southern Africa. In: D. H. S. Davies (Editor), Ecological Studies in Southern Africa. Junk, The Hague, pp. 1–23.

Cooke, H. J., 1975. The palaeoclimatic significance of caves and adjacent landforms in Western Ngamiland, Botswana. Geogr. J., 141: 430–444.

Cooke, H. J., 1979. The origin of the Makgadikgadi Pans. Botswana Notes Rec. 11: 37–42.

Cooke, H. J., 1980. Landform evolution in the context of climatic change and neo-tectonism in the middle Kalahari of north-central Botswana. Trans. Inst. Br. Geogr., 5: 80–90.

Cooke, H. J., 1984. The evidence from Northern Botswana of Late Quaternary climatic change. In: J. C. Vogel (Editor), Late Cainozoic palaeoclimates of the Southern Hemisphere, Proc. SASQUA Symp., Swaziland, pp. 265–278.

Cooke, H. J. and Verstappen, H. T., 1984. The landforms of the western Makgadikgadi basin in Northern Botswana. With a consideration of the chronology of the evolution of lake palaeo-Makgadikgadi. Z. Geomorphol., 28: 1–19.

Davis, W. M., 1922. Peneplains and the geographical cycle. Bull. Geol. Soc. Am., 33: 587–598.

Derricourt, R. M., 1976. Retrogression rate of the Victoria Falls and the Batoka Gorge. Nature, 264: 23–25.

De Swart, A. M. J. and Bennet, G., 1974. Structural and physiographic development of Natal since the Late Jurassic. Trans. Geol. Soc. South Afr., 77: 309–322.

Dixey, F., 1942. Erosion cycles in Central and Southern Africa. Trans. Geol. Soc. South Afr., 45: 151–181.

Dixey, F., 1955. Some aspects of the geomorphology of central and southern Africa. In: Alex L. du Toit Memorial Lecture, 4. Geol. Soc. South Afr. 58 pp.

Durand, A., 1982. Oscillations of Lake Chad over the past 50,000 years: new data and new hypothesis. Palaeogeogr., Palaeoclimatol., Palaeoecol., 39: 37–53.

Du Toit, A. L., 1926. Report on the Kalahari reconnaissance of 1925. Dep. Irrigation, Pretoria, 69 pp.

Fair, T. J. D., 1947. Slope form and development in the interior of Natal, South Africa. Trans. Proc;. Geol. Soc. South Afr., 50: 105–118.

Fair, T. J. D., 1948. Slope form and development in the coastal hinterland of Natal. Trans. Proc. Geol. Soc. South Afr., 51: 37–52.

Fair, T. J. D. and King, L. C., 1954, Erosional land surfaces in the eastern marginal areas of South Africa. Trans. Proc. Geol. Soc. South Afr., 57: 19–26.

Gaven, C., Hillaire-Marcel, C. and Petit-Maire, N., 1981. A Pleistocene lacustrine episode in southeastern Libya. Nature, 290: 131–133.

Gibbs, A. D., 1984. Structural evolution of extensional basin margins. J. Geol. Soc. Lond., 141: 609–620.

Grey, D. R. C. and Cooke, H. J., 1977. Some problems in the Quarternary evolution of the landforms of northern Botswana. Catena, 4: 123–133.

Grove, A. T., 1969. Landforms and climatic change in the Kalahari and Ngamiland. Geogr. J., 135: 191–212.

Grove, A. T., 1986. Geomorphology of the African Rift System. In: L. E. Frostick, R. W. Renaut, I. Reid and J. J. Tiercelin (Editors), Sedimentation in the African Rifts. Blackwell, pp. 9–16.

Grove, A. T., Street, F. A. and Goudie, A. S., 1975. Former lake levels and climatic change in the rift valleys of Ethiopia. Geogr. J., 141: 177–202.

Harrison and Coppinger, 1962. Navigational study of the Zambezi River (Barotseland). Fed. Dep. Surv. Rhodesia and Nyasaland.

Harrison, S. P., Metcalfe, S. E., Street-Perrott, F. A., Pittock, A. B., Roberts, C. N. and Salinger, M. J., 1983. A climatic model of the last Glacial/Interglacial transition based on palaeotemperature and palaeohydrological evidence. In: J. C. Vogel (Editor), Late Cainozoic Palaeoclimates of the Southern Hemisphere, Proc. SASQUA Symp., Swaziland, pp. 21–34.

Hastenrath, S. and Kutzbach, J. E., 1983. Paleoclimtic estimates from water and energy budgets in East African lakes. Quat. Res., 19: 141–153.

Heine, K., 1982. The main stages of the Lake Quaternary evolution of the Kalahari region, Southern Africa. Palaeoecol. Afr., 15: 53–76.

Helgren, D. M., 1978. Environmental stratigraphy of the relict alluvia and terraces along the Lower Vaal River, South Africa. Palaeoecol. Afr., 10: 163–170.

Helgren, D. M., 1984. Historical geomorphology and geoarchaeology in the southwestern Makgadikgadi basin Botswana. Ann. Assoc. Am. Geogr., 74: 298–307.

Hendey, Q. B. and Volman, T. P., 1986. Last interglacial sea levels and coastal caves in the Cape Province, South Africa. Quat. Res., 25: 189–198.

Hughes, T., 1973. Is the West Antarctic Ice Sheet distintegrating? J. Geophys. Res., 78: 7884–7910.

Hutchins, D. G., Hutton, S. M. and Jones, C. R., 1976. The geology of the Okavango Delta. In: Symp. Okavango Delta and its Future Utilisation, Gaborone, 1976, Botswana Society, Gaborone, pp. 13–19.

Jones, C. R. and Key, R. M., 1978. Botswana's contribution to the international Geodynamics project: a terminal report. Botswana Notes Rec., 11: 43–54.

Jouzel, J., Lorius, C., Petit, J. R., Genthon, C., Barkov, N. I., Kotlyakov, V. M. and Petrov, V. M., 1987. Vostok ice core: a continuous isotopic temperature record over the last climatic cycle (160,000 years). Nature, 329: 403–408.

Joyce, C., 1988. Global warming could wipe out wildlife. New Sci., 1598: 29.

Keen, C. E., 1985. The dynamics of rifting: Deformation of the lithosphere by active and passive driving forces. Geophys. J. R. Astron. Soc., 80: 95–120.

King, L. C., 1949. On the ages of African landsurfaces. Q. J. Geol. Soc. Lond., 416: 439–453.

King, L. C., 1955. Pediplanation and isostacy: An example from South Africa. Q. J. Geol. Soc. Lond., 111: 353–359.

King, L. C., 1962. The Morphology of the Earth. Oliver and Boyd, London, 2nd ed., 726 pp.

King, L. C., 1963. South African Scenery. Oliver and Boyd, Edinburgh, 3rd ed.

Lancaster, I. N., 1978. The pans of the Southern Kalahari, Botswana. Geogr. J., 144: 81–98.

Lancaster, I. N., 1979. Evidence for a widespread Pleistocene humid phase in the Kalahari. Nature, 279: 145–146.

Lancaster, I. N., 1980. Dune systems and palaeoenvironments in Southern Africa. Palaeontol. Afr., 23: 185–189.

Lancaster, I. N., 1981. Palaeoenvironmental implications of fixed dune systems in Southern Africa. Palaeogeogr., Palaeoclimatol., Palaeoecol., 33: 327–346.

Lancaster, I. N., 1984. Aridity in Southern Africa: Age, origins and expression in landforms and sediments. In: J. C. Vogel (Editor), Late Cainozoic Palaeoclimates of the Southern Hemisphere, Proc. SASQUA Symp., Swaziland, pp. 433–444.

Lane, E. W., 1955. The importance of fluvial morphology in hydraulic engineering. Proc. Am. Soc. Civil Eng., 81: 1–17.

Leopold, L. B., Wolman, M. G. and Miller, J. P., 1964. Fluvial Processes in Geomorphology, Freeman, San Francisco, 522 pp.

Lister, L. A., 1976. The erosion surfaces of Rhodesia. Thesis, Univ. Zimbabwe, Harare, 218 pp.

Lister, L. A., 1987. The erosion surfaces of Zimbabwe. Zimbabwe Geol. Surv. Bull., 90: 163 pp.

Livingstone, D. A., 1980. Environmental changes in the Nile headwaters. In: M. A. J. Williams and H. Faure (Editors), The Sahara and the Nile. Balkema, Rotterdam, pp. 339–359.

MacGregor, A. M., 1930. Geological notes on a circuit of the Great Makarikari salt pan, Bechuanaland Protectorate. Trans. Proc. Geol. Soc. South Afr., 33: 89–102.

McKenzie, D., 1978. Some remarks on the development of sedimentary basins. Earth Planet. Sci. Lett., 40: 25–32.

Malde, H. E., 1968. The catastrophic Late Pleistocene Bonneville flood in the Snake River Plain, Idaho. U.S. Geol. Surv. Prof. Pap., 536, 52 pp.

Mallick, D. I. J., Habgood, F. and Skinner, A. C., 1981. A geological interpretation of Landsat imagery and air photography of Botswana. H.M.S.O. Overseas Geol. Miner. Resour., 56, 35 pp.

Mercer, J. H., 1973. Cainozoic temperature trends in the southern hemisphere: Antarctic and Andean glacial evidence. Palaeoecol. Afr., 8: 85–114.

Mercer, J. H., 1978. West Antarctic ice sheet and CO_2 greenhouse effect: A threat of disaster. Nature, 271: 321–325.

Muller, T. and Pope, G. V., 1982. Vegetation of the Valley. In: R. F. Du Toit, A Preliminary Assessment of the Environmental Implications of the Proposed Mupata and Batoka hydroelectric Schemes (Zambezi River, Zimbabwe). Nat. Resour. Board, Harare, pp. 53–67.

Nicholson, S. E. and Flohn, H., 1980. African environmental and climatic changes and the general atmospheric circulation in Late Pleistocene and Holocene. Clim. Change, 2: 313–348.

Nugent, C., 1983. Channel changes of the middle Zambezi. Zimbabwe Sci. News, 17: 127–129.

Nugent, C., 1986. Historical changes in the behaviour of the Zambezi River at Nyamuomba. Zimbabwe Sci. News, 20: 121–131.

Nugent, C., 1990. Development of the Victoria Falls. In: D. W. Phillipson (Editor), Mosi-oa-Tunya, A Handbook for the Victoria Falls. Longman, Zimbabwe.

Orpen, J. L., Swain, C. J., Nugent, C. and Zhou, P. P., 1990. Wrench-fault and half-graben tectonics in the development of the Palaezoic Zambezi Karoo basins of Zimbabwe — the "Gwembe" and "Chicoa" basins and regional implications. J. Afr. Earth Sci., in press.

Oswell, W. C., 1900. William Cotton Oswell, hunter and explorer. Heinemann, London (2 vols).

Partridge, T. C. and Maud, R. R., 1987. Geomorphic evolution of Southern Africa since the Mesozoic. South Afr. J. Geol., 90: 179–208.

Peters, D. U., 1960. Land usage in Barotseland. Rhodes — Livingstone Commun., 9, 60 pp.

Phillipson, L., 1976. Mode Three artefact occurrences at Bagasa in the Middle Zambezi Valley. South Afr. Archaeol. Bull., 31: 112–126.

Phillipson, L., 1977. Stone Age sites near Chirundu in the Middle Zambezi Valley. South Afr. Archaeol. Bull., 32: 28–62.

Reeves, C. V., 1972. Rifting in the Kalahari? Nature, 237: 95–96.

Ritchie, J. C. and Haynes, C. V., 1987. Holocene vegetation zonation in the Eastern Sahara. Nature, 330: 645–647.

Rossignol-Strick, M., 1983. African monsoon, an immediate climate response to orbital insolation. Nature, 304: 46–49.

Scholz, C. H., Koczynski, T. A. and Hutchins, D. G., 1976. Evidence for incipient rifting in Southern Africa. Geophys. J. R. Astron. Soc., 44: 135–144.

Sengor, A. M. C. and Burke, K., 1978. Relative timing of rifting and volcanism on earth and its tectonic implications. Geophys. Res. Lett., 5: 419–421.

Shaw, P. A., 1984. A historical note on the outflows of the Okavango delta System. Botswana Notes Rec., 16: 127–130.

Shaw, P. A., 1985. Late Quaternary landforms and environmental changes in northwest Botswana: The evidence of Lake Ngami and the Mababe Depression. Trans. Inst. Br. Geogr., 10: 333–346.

Shaw, P. A. and Cooke, H. J., 1986. Geomorphic evidence for the Late Quaternary palaeoclimates of the Middle Kalahari of Northern Botswana. Catena, 13: 349–359.

Shaw, P. A. and Thomas, D. G. S., 1988. Lake Caprivi: A Late Quaternary link between the Zambezi and Middle Kalahari drainage systems. Z. Geomorphol., in press.

Singer, R. and Wymer, J., 1982. The Middle Stone Age at Klasies River Mouth in South Africa. Univ. Chicago Press, 234 pp.

Stagman, J. G., 1978. An outline of the geology of Rhodesia. Rhod. Geol. Surv. Bull., 80, 126 pp.

Stigand, A. G., 1923. Ngamiland. Geogr. J., 42: 401–419.

Street, F. A. and Grove, A. T., 1976. Environmental and climatic implications of Late Quaternary lake level fluctuations in Africa. Nature, 261: 385–390.

Street, F. A. and Grove, A. T., 1979. Global maps of lake-level fluctuations since 30,000 B.P. Quat. Res., 12: 83—118.

Street-Perrott, F. A., Roberts, N. and Metcalfe, S., 1985. Geomorphic implications of Late Quaternary hydrological and climatic changes in the northern hemisphere tropics. In: I. Douglas and T. Spencer (Editors), Environmental Change and Tropical Geomorphology. Allen and Unwin, London, pp. 165–183.

Tankard, A. J., Jackson, M. P. A., Eriksson, K. A., Hobday, D. K., Hunter, D. R. and Minter, W. E. L., 1982. Crustal Evolution of Southern Africa. Springer, New York, N.Y. 523 pp.

Thomas, D. G. S., 1984a. Geomorphic evolution and river channel orientation in northwest Zimbabwe. Proc Geogr. Assoc. Zimbabwe, 15: 12–22.

Thomas, D. G. S., 1984b. Ancient ergs of the former arid zones of Zimbabwe, Zambia and Angola. Trans. Inst. Br Geogr., 9: 75–88.

Thomas, D. G. S., 1987. Discrimination of depositional environments using sedimentary characteristics in the Mega Kalahari, Central Southern Africa. In: L. Frostick and I. Reid, (Editors), Desert sediments: Ancient and Modern. Geol. Soc. Spec. Publ., 35: 293–306.

Vail, J. R., 1967. The southern extension of the East African rift system and related igneous activity. Geol. Rundsch., 57: 601–614.

Verboom, W. C., 1974. The Barotse loose sands of Western Province, Zambia. Zambian Geogr. Assoc. Mag., 27: 13–17.

Verboom, W. C. and Brunt, M. A., 1970. An Ecological Survey of Western Province, Zambia, with Special Reference to the Fodder resources, Vol. 1, The Environment. Land Resour. Stud. Dir. Overseas Surv., 8, 95 pp.

Visser, J. N. J., 1984. A review of the Stormberg Group and Drakensberg Volcanics in Southern Africa. Palaeontol. Afr., 25: 5–27.

Volman, T. P., 1984. Early prehistory of Southern Africa. In: R. G. Klein (Editor), Southern African Prehistory and Palaeoenvironments. Balkema, Rotterdam, pp. 169–220.

Watts, A. B., Karner, G. D. and Steckler, M. S., 1982. Lithospheric flexure and the evolution of sedimentary basins. Philos. Trans. R. Soc. Lond., A305: 249–281.

Weertman, J., 1974. Stability of the junction of an ice sheet and an ice shelf. J. Glaciol., 13: 3–11.

Wellington, J. H., 1955. Southern Africa — A Geographical Study. Vol. 1, Physical Geography. Cambridge Univ. Press, 528 pp.

Wernicke, B. and Burchfiel, B. C., 1982. Modes of extensional tectonics. J. Struct. Geol., 4: 105–115.

Williams, G. J., 1987. Pulse of the Zambezi. Geogr. Mag., 59: 608–613.

Wilson, B. H. and Dincer, T., 1976. An introduction to the hydrology and hydrogeography of the Okavango delta. In: Symp. Okavango Delta and its Future Utilisation, Gaborone, 1976. Botswana Society, Gaborone.