30th Alex L. du Toit Memorial Lecture

THE OKAVANGO DELTA AND ITS PLACE IN THE GEOMORPHOLOGICAL EVOLUTION OF SOUTHERN AFRICA

T.S. McCARTHY

Island vegetation shows a distinct zonation: the outer fringes are characterized by evergreen trees, followed inwards by deciduous trees, ivory palms, shrubs and grass and finally barren soil. The swamp surrounding the island is densely vegetated with sedges and grasses.

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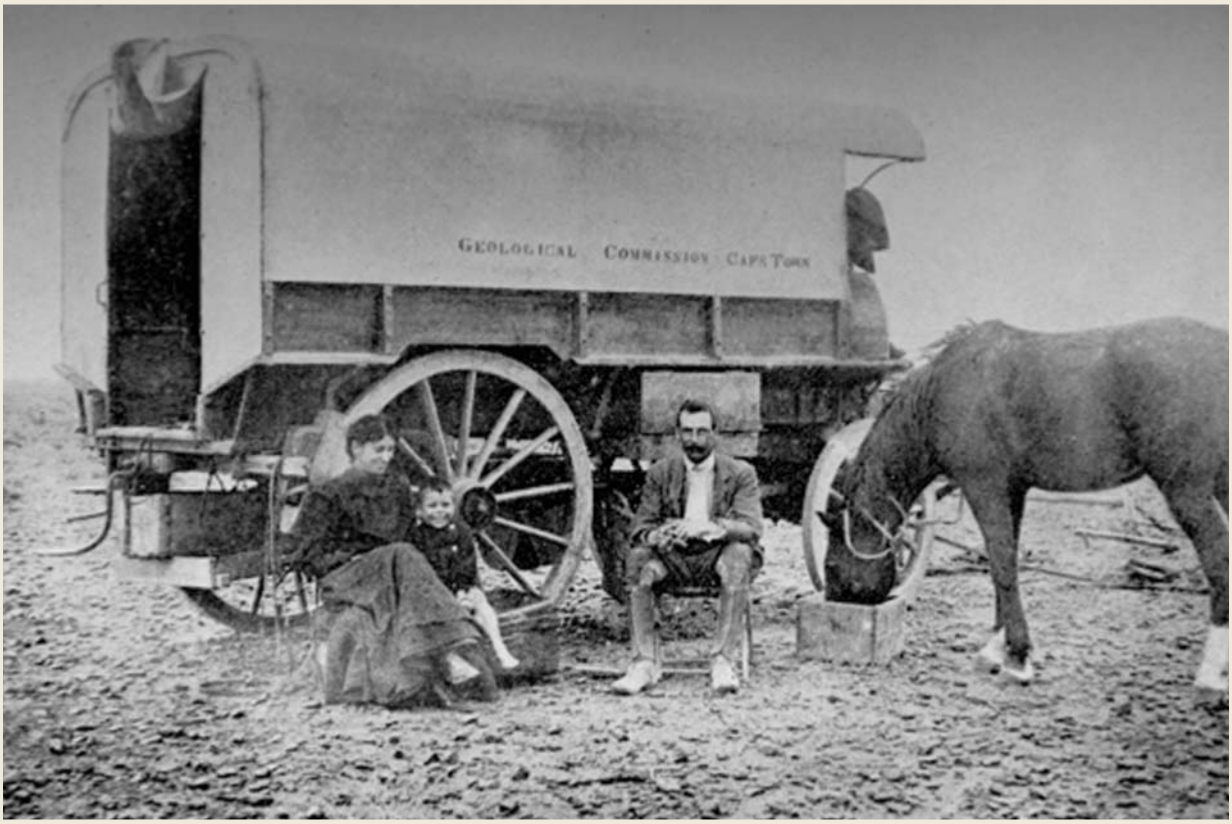

Alex du Toit in the field with his first wife and only son in the early 1900s.

Picture by kind courtesy M.C.J. de Wit, E.O. Köstlin and R.S. Liddle *Photo: Gevers*

THE OKAVANGO DELTA AND ITS PLACE IN THE GEOMORPHOLOGICAL EVOLUTION OF SOUTHERN AFRICA

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ABSTRACT

The Okavango Delta is southern Africa's largest wetland ecosystem and probably the most pristine large wetland ecosystem in the world. Alex du Toit was the first to recognize the role of faulting in the origin of the Delta, proposing that the Delta lies within a graben structure related to the East African Rift Valley system. The history of its rivers is more ancient, extending back to the breakup of Gondwana.

Two mantle plumes initiated the breakup of Gondwana: the 140 Ma Karoo plume and the 130 Ma Parana-Etendeka plume. Domical uplift and rifting associated with these plumes created two major river systems: the Okavango-Zambezi-Limpopo system and the Vaal-Orange system. The climate at the time of breakup was hot and humid and the interior experienced extensive erosion, so much so that by the Oligocene (ca. 30 Ma), the sub-continent had been planed to base level, rising only a few hundred metres above sea level and mantled by thick, leached soils (now known as the African Erosion Surface).

Warping in the continent interior created uplifted arches and depressions, most notably the Kalahari basin. Arching severed the link between the lower Limpopo and its central African headwaters (Zambezi-Okavango), and a large lake formed in the Kalahari depression (Lake Palaeo-Makgadikgadi). This lake gradually disappeared, partly due to sedimentation but mainly due to the increasing dry climate.

The East Africa Rift system commenced in the Afar about 30 Ma ago and began to propagate southwards. In southern Malawi and especially in Zambia, the path of rifting has not yet been clearly established and the region is characterized by numerous horsts and grabens. One of these grabens passes through the Okavango Delta. The formation of these grabens has profoundly affected the courses of the rivers in the region.

The Okavango River debouches into the graben forming a large alluvial fan. Lakes have periodically existed at the toe of the Okavango fan where it abuts the bounding fault scarps, but these are not permanent. Some Okavango water discharges across the bounding fault scarps and flows into the Makgadikgadi depression to the southeast.

The Okavango River catchment is largely underlain by Kalahari sand, which forms the major particulate sediment carried by the river. Consequently, sediment carried by the river is mainly fine sand, with little silt and mud. The dissolved solid concentration in the river water is low (ca. 40 mg/L) and consists mainly of silica and calcium and magnesium bicarbonates. However, the volume of water entering the Delta each year is large and hence the solutes constitute the largest component of the sediment carried into the Delta.

The Okavango River discharges onto the alluvial fan where water is carried in channels that form the major primary distributaries. Channel margins are formed by vegetal material and are permeable, leaking water which sustains permanent swamps in the upper portion of the alluvial fan. The arrival of the seasonal flood increases the rate of channel leakage, forming the seasonal swamps on the lower fan. The advance of the flood water across the seasonal swamps is slow, as much of it infiltrates the ground, taking four to five months to traverse the 250 km length of the fan.

Bedload is confined to channels and as water leaks through the channel margins, channel beds aggrade, increasing leakage, which further promotes bed aggradation. Channels eventually fail and water diverts elsewhere. Channel formation and failure results in a constant shift in the distribution of water across the Delta surface. The demise of a channel system results in desiccation of the surrounding peat, which is then destroyed by slow-burning peat fires. Nutrients and fine particulate material accumulated in the peat is released, enriching the soil.

Most of the water delivered to the Delta annually is lost to the atmosphere by evapotranspiration because of the semi-arid climate. Transpiration of groundwater by terrestrial plants is the dominant means of water loss. The high transpiration rate of trees is particularly important: trees grow on islands and their transpiration lowers the water table beneath islands so there is a net flow of water towards islands. Solutes, especially silica and carbonates of calcium and magnesium, precipitate, leaving only the very soluble sodium carbonate in solution. Its concentration rises and impacts on the vegetation, resulting in a zonation in the distribution of plant species on islands. The salinity of the groundwater ultimately rises to the point where gravity induced advection occurs, thus transferring the sodium carbonate to the deep groundwater. This process prevents the formation of surface saline brine in the Delta and surface water remains fresh. The accumulation of precipitated solutes results in expansion and this form of chemical precipitation is the major mechanism of sedimentation in the distal regions of the Delta. Islands are mainly initiated as a consequence of termite activity during dry periods. Sand ridges which form by channel bed sedimentation may also result in islands.

Constant changes in the distribution of water across the fan due to channel failure have profound effects on the ecology of the Delta: regions of swamp may revert to dry land, when rain flushes accumulated salts from the island soils; and formerly dry areas become seasonally or even permanently flooded. Such constant changes, operating on time-scales of decades to centuries, underpin the immense habitat diversity of the Okavango Delta.

Introduction

Alex du Toit's contribution to the geological understanding of southern Africa remains unsurpassed by any other individual. He was first and foremost a field geologist (Gevers, 1950), and from early in his career had the advantage of being exposed to the diverse geology of South Africa through his involvement in the regional mapping of the country. The preparation of these maps and reports required a very diverse knowledge and no doubt honed his observational skills and fostered his very broad interests. He travelled extensively in southern Africa and as a result he developed an intimate knowledge of the geology and landscape of the region, which culminated in his major work, the "Geology of South Africa" (du Toit, 1926; 1939; 1954). He read and travelled widely outside of southern Africa, which put him in a position to be able to compare the geology of different continents from first hand observations, resulting in his other great work, "Our Wandering Continents" (du Toit, 1937).

Although du Toit is perhaps best remembered for his writings on the geology of South Africa and on continental drift, he made major contributions to our understanding of many other aspects of the geological and geomorphological evolution of our region. Important amongst these were his observations and

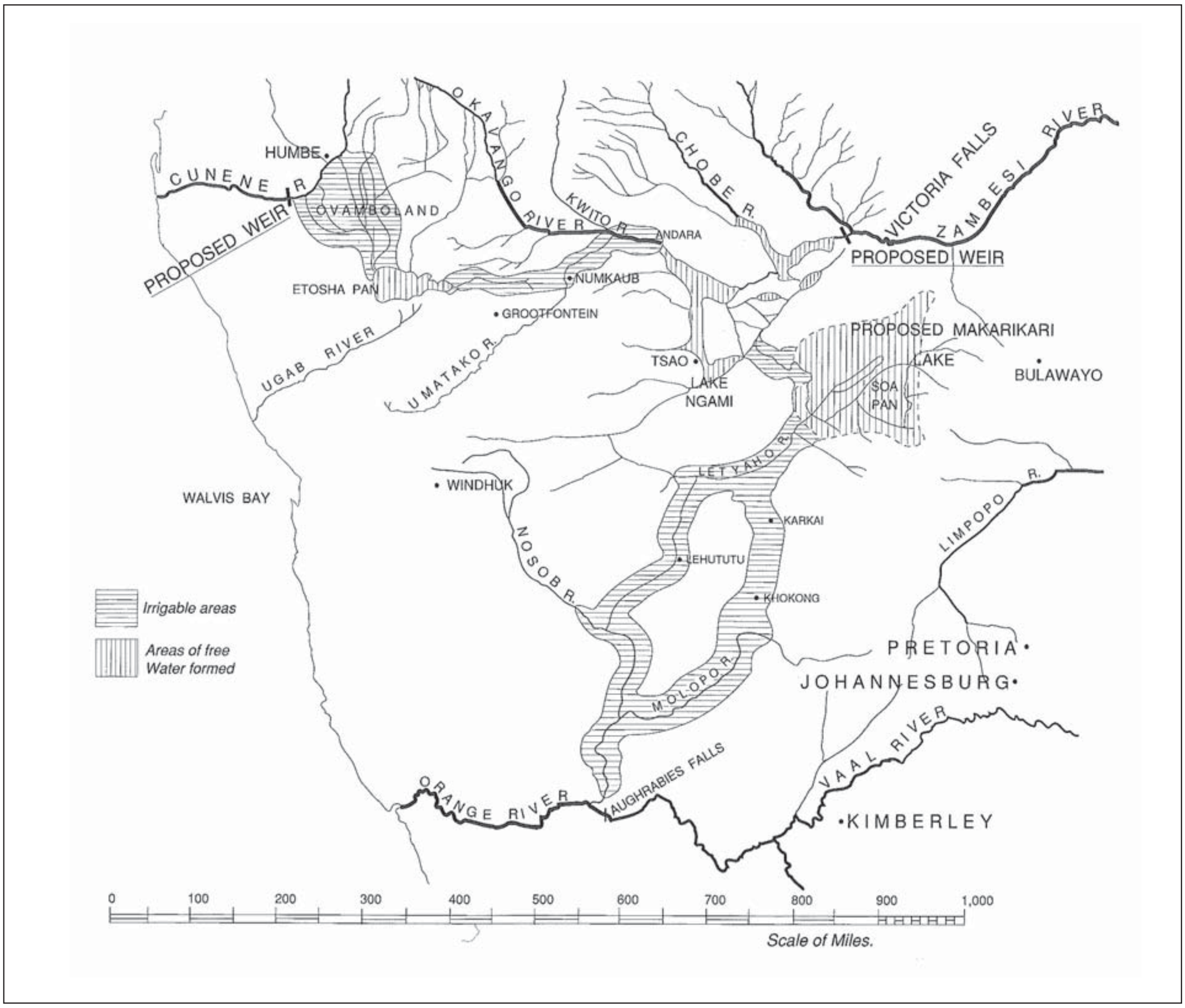

Figure 1. *Schwarz (1920) proposed that the Etosha and Makgadikgadi Basins be flooded by the construction of weirs to divert the Cunene (sic) and Okavango/Chobe (sic) Rivers into the respective basins. He postulated that evaporation from the resulting lakes would enhance rainfall over the entire Kalahari region, making extensive irrigation possible, as illustrated in his map of the scheme.*

ideas on the development of the southern African landscape, especially the Kalahari. On a more local scale, and particularly relevant to this article, he undertook the first detailed field investigation of the Okavango and Kwando Rivers and the adjacent reaches of the Zambezi and their relationship to the Makgadikgadi pans.

Du Toit's involvement in the Okavango region stemmed from a notion formulated by Professor E.H.L. Schwarz (1920) of Rhodes University. Schwarz posited that southern Africa was becoming increasingly arid because the extensive lake systems that had previously existed in the interior (Figure 1) had dried up because of the capture and diversion of the Okavango and Kwando Rivers from their former course into the Makgadikgadi pans. He proposed that the flow of these rivers to the Makgadikgadi should be restored, thus re-establishing the lake and thereby increasing local rainfall. His idea caught the attention of the media and of politicians in particular and Alex du Toit, who was then geologist for the Department of Irrigation, was dispatched to the Okavango region in 1925 to investigate the feasibility of Schwarz's proposal.

Du Toit's expedition was well resourced and his team carried out detailed topographic surveys of the region as well as aerial reconnaissance and hydrological analysis of the rivers. On the basis of the data gathered, du Toit realized that a good proportion of the discharge of the Kwando and all of that of the Okavango River was being lost to the atmosphere, without evidently influencing local climate. He concluded that the only way to recreate the lakes would be to divert not only the Okavango and Kwando Rivers but also entire discharge of the Zambezi into the pans as well, but even then, he surmised that an improvement of the climate of the region was unlikely (du Toit, 1926).

The investigation provided du Toit with deep insight into the development of the region. Observations he

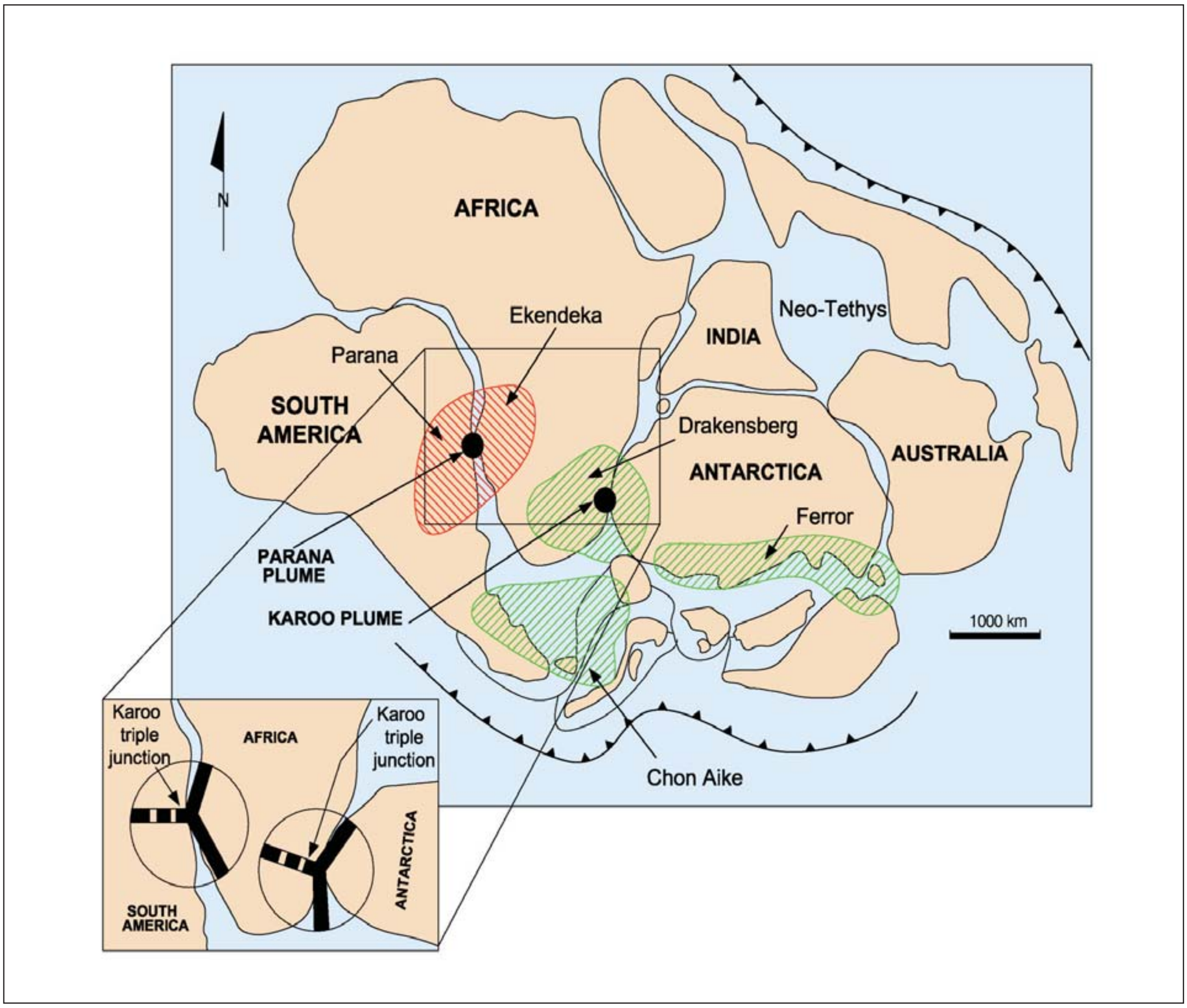

Figure 2. *Mantle plumes that were responsible for the eruption of the Drakensberg and Etendeka volcanics (Karoo and Parana plumes respectively) are believed to have been instrumental in initiating the break-up of Gondwana, as illustrated here (compiled from: White and McKenzie, 1989; Storey and Kyle, 1997; Moore and Blenkinsop, 2002).*

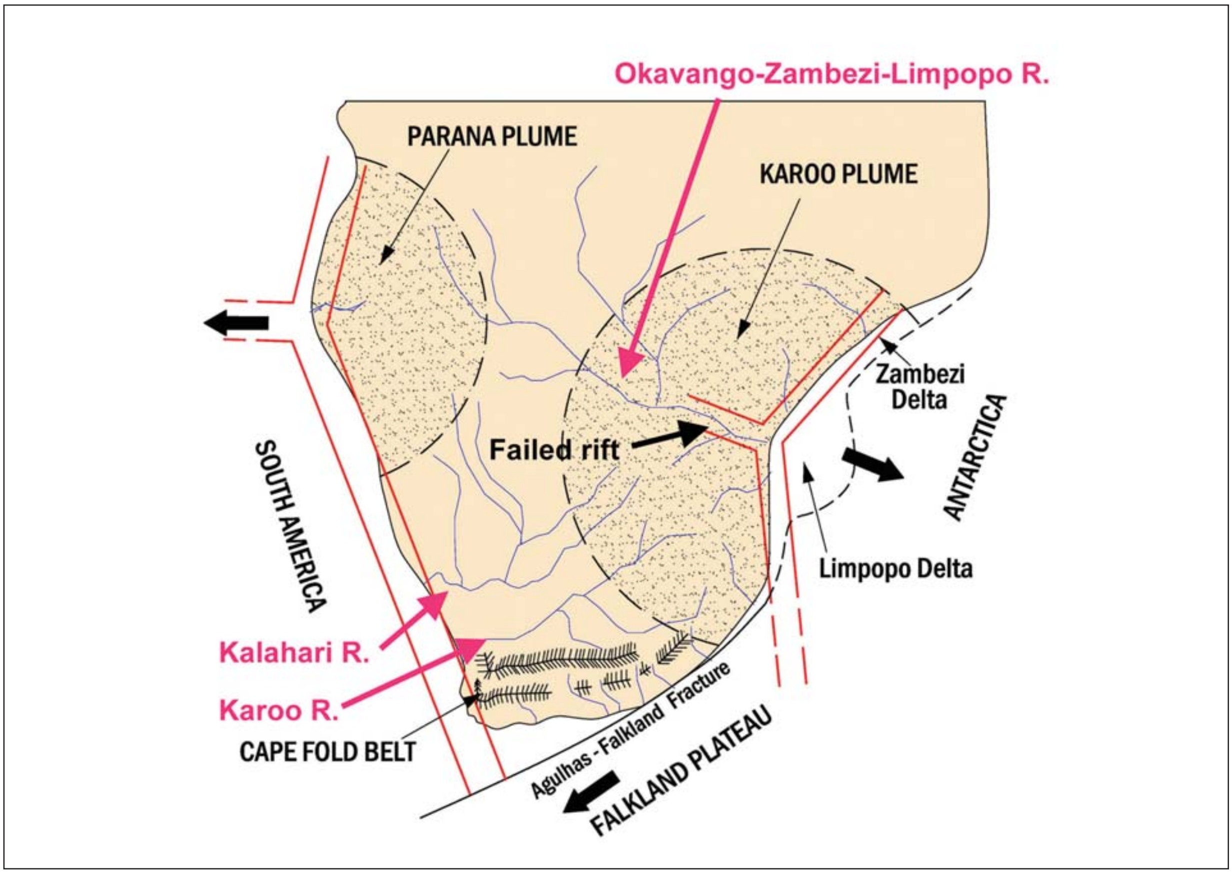

Figure 3. *The Parana and Karoo plumes are believed to have profoundly influenced the drainage patterns that developed in southern Africa after the break-up of Gondwana (Cox, 1989; Moore and Blenkinsop, 2002).*

made during his survey were integrated with his extensive knowledge of the surrounding Kalahari, and in an address to the South African Association for the Advancement of Science in 1927 he was able to weave together information from disparate disciplines into a coherent picture of the development of the landscape of the interior of southern Africa (du Toit, 1927).

In this article, I shall re-examine du Toit's views on the geomorphological development of the region in the light of the research that has taken place in the subsequent 80 or so years, and then focus on the Okavango Delta itself, an area only touched on in the most general terms by du Toit. The morphology of southern Africa and the nature of its river systems have been dictated by events surrounding the break-up of Gondwana, so it is there that my account commences, which is appropriate as it was du Toit who laid the foundations for our present understanding of how southern Africa came to be (du Toit, 1937).

The break-up of Gondwana and the development of early drainage systems.

Breakup of Gondwana commenced with the opening of the Mozambique Channel and the Indian Ocean, which were initiated by the Karoo plume (Storey and Kyle, 1997). The plume was responsible for the eruption of the Drakensberg igneous province in South Africa, the Ferrar province in Antarctica and the Chon Aike province in South America (Figure 2). Burke and Dewey (1973) identified the focus of the plume at the time of breakup as situated at the intersection of the Lebombo and Sabi monoclines, a position coinciding with a major convergence of dykes of Karoo age (Reeves, 1978, 2000; Uken and Watkeys, 1997). A second plume, situated in Angola (Figure 2), gave rise to the Etendeka volcanics of Namibia and the Parana volcanics of South America (White and McKenzie, 1989). This plume, which erupted between 137 and 127 Ma (Moore and Blenkinsop, 2002), is believed to have led to the opening of the Atlantic.

Cox (1989) examined the effects of plume-induced doming on drainage patterns in India and southern Africa. He demonstrated that the peculiar asymmetry of drainage in both India and southern Africa can be understood in terms of a plume model. Thus, the headwaters of the Vaal and Orange Rivers, which discharge into the Atlantic in the west, lie close to the eastern margin of the continent on the flanks of the Karoo plume, whilst the headwaters of the Zambezi system, which discharges into the Indian Ocean in the east, lie in the western highlands of Angola on the flanks of the Parana plume.

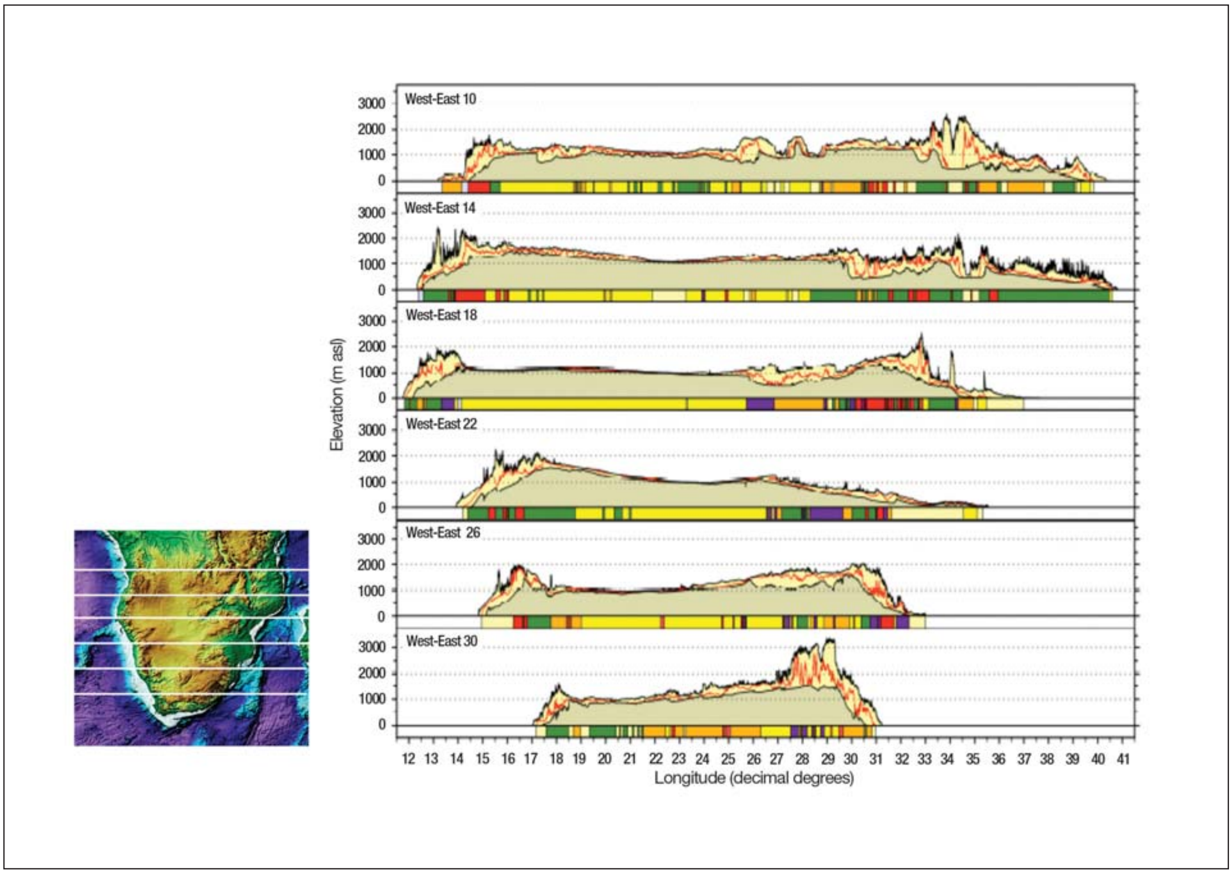

Figure 4. *Topographic cross sections at various latitudes across southern Africa. The red lines show the actual topography and the yellow envelopes show the variation in the elevation over a 2 degree swathe astride the section line (Rowberry, McCarthy and Tooth, in prep.). Inset shows the locations of the sections.*

Moore and Blenkinsop (2002) extended this model (Figure 3). Following du Toit (1933) and Thomas and Shaw (1991), they envisage that the Okavango River linked to the Limpopo, and moreover was joined by the Quando, Upper Zambezi, Kafue and Luangwa Rivers. In their view, the lower Limpopo was the primary conduit for runoff and sediment eroded from central southern Africa, which occupied a failed arm of the Karoo triple junction.

Geomorphological analysis indicates that the Orange River also had a complex history, possibly involving two river systems (de Wit, 1999), the Kalahari River, which drained the southern portion of the Parana dome and the region between the Parana and Karoo domes, and the Karoo River, which included the upper Orange and Vaal Rivers and drained the western flank of the Karoo dome (Figure 3).

Events surrounding the breakup of Gondwana thus had a profound influence on the drainage system of southern Africa. The bulk of the region was drained by just three river systems - the Okavango-Zambezi-Limpopo in the north, flowing eastwards through the failed Limpopo Rift and draining the Parana dome; the Karoo River flowing southwestwards, draining the Karoo dome and the highlands of the Cape Fold Belt in the south; and the Kalahari River draining the southern flank of the Parana dome and the region between the domes. Thermal expansion related to the rifting had probably elevated the Angolan and KwaZulu-Natal coasts, resulting in escarpments facing the newly formed sea-ways. Short, steep rivers formed on these escarpments, and included the northern Olifants River, which drained the rifted flank of the Karoo dome immediately south of the Limpopo rift. Although of steep gradient, the rate of headward erosion by these rivers has been generally slow because of their restricted catchments. The drainage history along the southern margin of the continent is less well known. Separation along this margin was by strike-slip faulting and hence without thermal uplift. It seems likely, however, that at the time of breakup, elevated topography arising from the Cape orogeny still persisted in this region. This elevated terrain created a watershed between rivers draining northwards to the Karoo River and new, short rivers that formed along the southern coast as the Falklands Plateau detached (Figure 3).

Planation of southern Africa

At the time of initiation of the break-up of Gondwana, most of what is now southern Africa was covered by

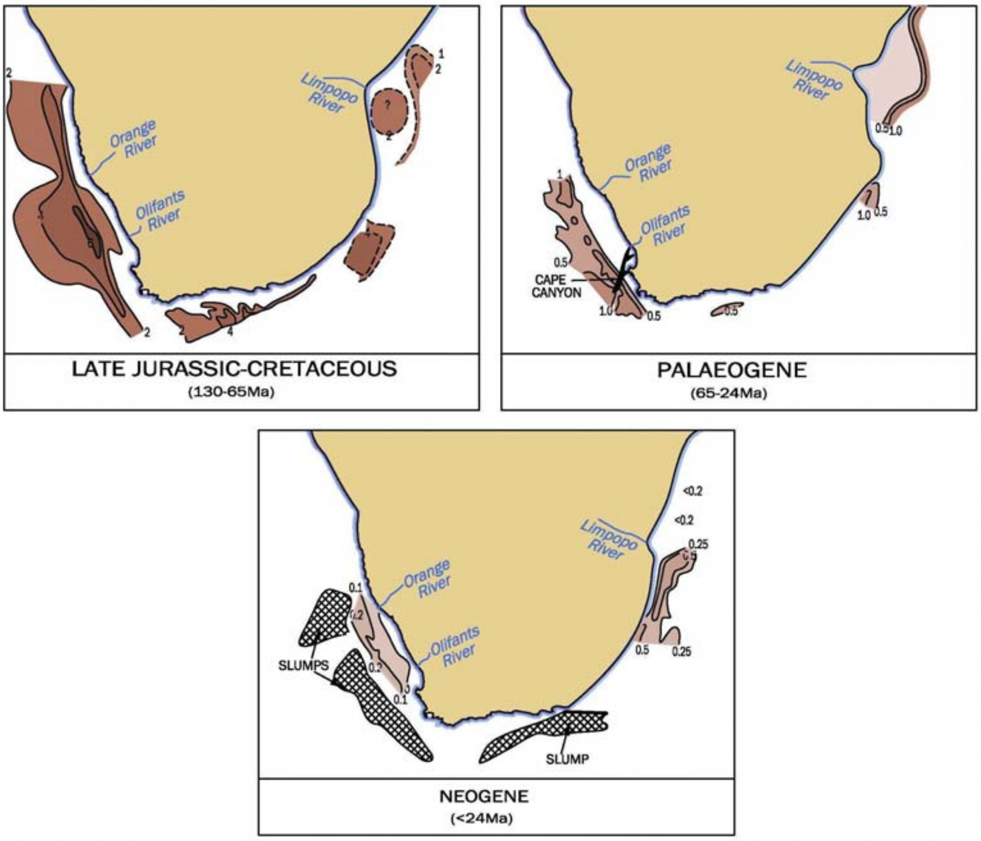

Figure 5. *Offshore sedimentation around the southern African coast (isopachs in km). Compiled from Dingle et al. (1983), Flores (1973), Moore and Larkin (2001) and McMillan (2003).*

Drakensberg lavas and their equivalents, with the probable exception of the southern Cape. Fairly substantial thicknesses of these rocks are still to be found across the region, notably along the Lebombo Mountains, in the Maluti Mountains of Lesotho, in the Springbok Flats between Warmbaths (Bela Bela) and Potgietersrus (Mokopane), in grabens astride the Limpopo River, in central Botswana (largely beneath a cover of Kalahari Group sediments) and southern Namibia. They also probably underlie Cretaceous sediments off the east and west coasts (Flores, 1973; Dingle et al, 1983; Burke and Gunnell, 2008). Areas underlain by these lavas have clearly experienced relatively little erosion since the break-up of Gondwana. However, it is unlikely that the actual land surface of Gondwana is anywhere preserved.

Du Toit (1933, 1939) noted that southern Africa had undergone a period of intense planation, followed by uplift. He inferred a late Cretaceous to Eocene age for the planation surface. The interior of southern Africa, he noted, had been worn down “into the condition of a plain of near ideal perfection” (du Toit, 1933, p 6). He noted too that this surface had experienced uplift, which he estimated to have exceeded 600 m in the interior, and parallel to the coast a hinge-line had formed where uplift was a maximum, so that the surface dipped towards the ocean around the coast and also centrally in the interior. He surmised that this phase of uplift led to intense erosion, especially around the coast, and “that striking, lengthy feature the Great Escarpment fronting the ocean was thereby formed” (du Toit, 1939, p 510).

Following the pioneering ideas of Lester King, Partridge and Maud (1987, 2000) emphasized the widespread development of one planation surface in particular, the African Surface, which they recognized both inland and below the escarpment. Where fully preserved, this surface is capped by laterite, often silcrete. In their view, this surface could be seen to have experienced two later period of incision, which they termed the post-African I and II surfaces. The African Surface has more recently been examined on a continental scale by Burke and Gunnell (2008).

Key aspects of the morphology of southern Africa are also revealed by the east-west topographic sections shown in Figure 4 (Rowberry, McCarthy and Tooth, in prep.). The interior is characterized by a gently tilted and slightly warped surface lying at an elevation of between 1000 and 1800 m amsl, which shows little relief. The smooth nature of the surface (i.e. low local relief) in the Kalahari basin is partly due to the overlying veneer of sediment, but as this sediment is generally less that 200 m thick (Haddon, 1999, 2005; Haddon and McCarthy, 2005), the underlying pre-Kalahari surface must also be more or less equally smooth. This planar surface corresponds to the African Surface of Partridge and Maud (1987) and du Toit's (1937) "plane of near ideal perfection". In the northerly cross sections in Figure 4 at it can be seen to have an overall inclination to the east whereas in the southerly sections it is inclined to the west. A few residuals rise above this surface, notably the Maluti Mountains of Lesotho, the Kamiesberge of Namaqualand, and the Asbesberge in the interior. The interior surface is terminated by the Great Escarpment where the land surface drops to a shoreward sloping, moderately dissected, coastal plain. Here too, hill tops define a sloping palaeosurface, which Partridge and Maud (1987) also considered to be the African Surface, based on the presence of laterite cappings on hill tops.

Two aspects of the African Surface are important: its low relief and its inclination. The low relief implies that it has been planed to base level. The northern, easterly sloping portion of the surface is drained, and was presumably graded by, the Zambezi-Limpopo system, which discharges into the Indian Ocean, whilst the southern portion of the surface is drained and was graded by the Vaal-Orange-Molopo-Nossob-Auob system, which drains to the Atlantic Ocean. Since both the easterly and westerly sloping surfaces are graded to base level by different river systems yet are contiguous, it follows that both surfaces must have been graded to the same base level, which must therefore have been sea level. This implies that the African Surface was graded to the oceanic base level, and hence southern Africa must at one time have lain at an elevation of just a few hundred metres above sea level – possibly analogous to the interior of Australia today.

Further insight into the planation history is provided by sediment accumulation offshore of southern Africa as shown in Figure 5. The bulk of sedimentation occurred during the Cretaceous. Lesser amounts of sediment accumulated off the southern and southeastern coast, particularly in small graben structures formed along the Agulhas-Falkland Fracture Zone. In the case of the Limpopo Delta, much of the Cretaceous sediment deposition took place under terrestrial conditions (Sena Formation) and marine sedimentation only commenced in the upper Cretaceous (Dingle et al., 1983).

The paucity of post Eocene sediments (Figure 5) indicates that planation of the interior had been completed by this time. This inference is further supported by the occurrence of Eocene marine sediments lying on the planation surface in the eastern Cape (Bathurst Formation; du Toit, 1939; Maud and Botha, 2000), although it must be emphasized that the planation surface is undoubtedly diachronous, as erosion ceased at different times in different places (Burke and Gunnell, 2008).

Apatite fission track studies (e.g. Brown et al., 2000; Brown et al., 2002; Tinkler et al., 2008) confirm extensive Cretaceous erosion across southern Africa. Inland sites provide further indications of the age of the African Surface. Crater-fill sediments are preserved in the mid-Cretaceous (92 Ma) Orapa kimberlite pipe in Botswana (Bamford, 2000) and Cretaceous age alluvial deposits have been recorded on the Ghaap Plateau at Mahura Muthla (Partridge, 1998). Crater fill sediments are also preserved in several late Cretaceous diatremes in Namaqualand (Banke) and Bushmanland (Stompoor) (Bamford, 2000), indicating minimal erosion since that time. Using crustal xenoliths in kimberlite pipes, Hanson et al. (2009) established that the Drakensberg lavas were stripped from large areas in central South Africa between 120 Ma and 85 Ma (the time of emplacement of Group II and Group I kimberlites).

The African Surface is characterized by deep kaolinization (or bauxite development in Central and North Africa; Burke and Gunnell, 2008) and is further armoured by a variety of duricrust deposits, commonly silcrete or calcrete, ferricrete and more rarely manganiferous deposits (Partridge and Maud, 1987, 2000). Duricrust formation was probably also diachronous but appears to be everywhere older than 40 Ma, with maximum development taking place around 53 Ma during the Paleocene-Eocene Thermal Maximum (Burke and Gunnell, 2008).

River systems in the interior

The bulk of the planation of the African Surface was evidently brought about by two river systems that formed in the aftermath of the breakup of Gondwana: the Orange River and its precursors in the south and the Limpopo and its precursors in the north (Figure 3). The southern and southeastern sections were drained by short rivers (Partridge and Maud, 1987).

Several models have been proposed for the evolution of the Orange River (e.g. Dingle and Hendey, 1984; Partridge and Maud, 1987; de Wit, 1999; Burke and Gunnell, 2007). Although important in the context of the beveling of the interior, the Orange is of limited relevance to the Okavango Delta and will not be discussed further.

Du Toit (1927) first proposed a connection between the Okavango and Limpopo Rivers. These notions were developed and extended by Thomas and Shaw (1991), who proposed that the upper Zambezi, Kafue and Luangwa Rivers, along with the Okavango, were originally tributaries of the Limpopo, entering via Shashi River. Moore and Larkin (2001) extended this model and proposed that this combined river system with its vast catchment was largely responsible for the construction of the Limpopo Delta, which extends from Maputo to Beira. It was this river system that was responsible for

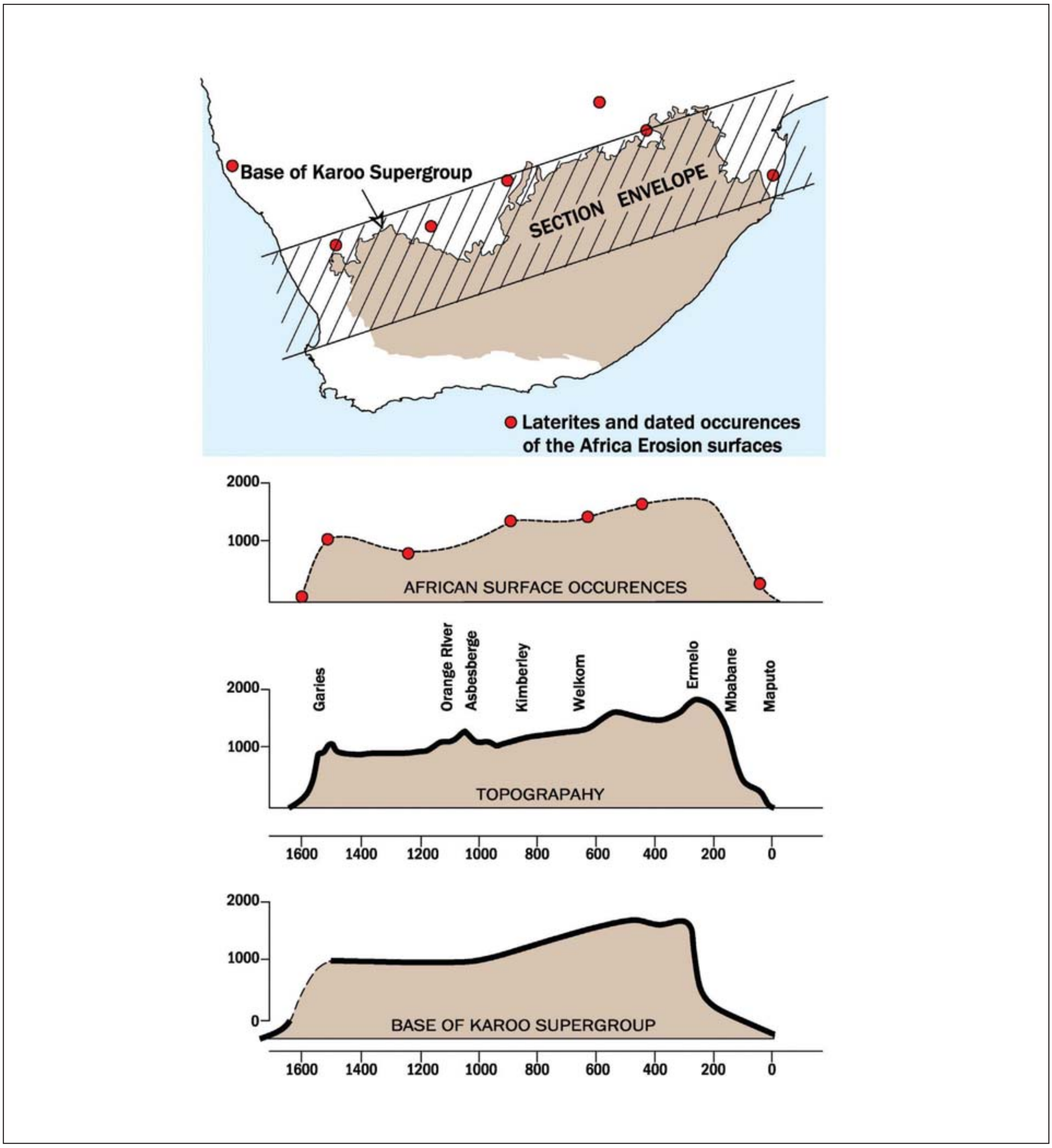

Figure 6. *Cross sections showing the elevations amsl of laterite cappings on the African Surface (top), the present-day topography (middle) and the elevation of the base of the Karoo Supergroup (bottom). The sections are encompassed in the envelope shown in the map.*

planation of the African Surface in central Botswana, Angola, western Zimbabwe and the northern parts of South Africa. By early Palaeocene times, this drainage basin had been eroded to base level and southern Africa was a vast, low lying plain.

Uplift of southern Africa

The African Surface in the interior lies at an elevation above 1000 m amsl, indicating major uplift. Insight into the manner of uplift can be gleaned from analysis of laterites and from geological considerations. Figure 6 shows sections across South Africa on which the elevations of laterite remnants on the African Surface, the present topography and the elevation of the base of the Karoo Supergroup are plotted. The Karoo Supergroup rises from below sea level on the east and west coasts to form arches that flank a broad depression in the interior – a pattern that exactly mirrors the present day land surface. Although only a few laterite remnants are preserved, their distribution, which reflects that of

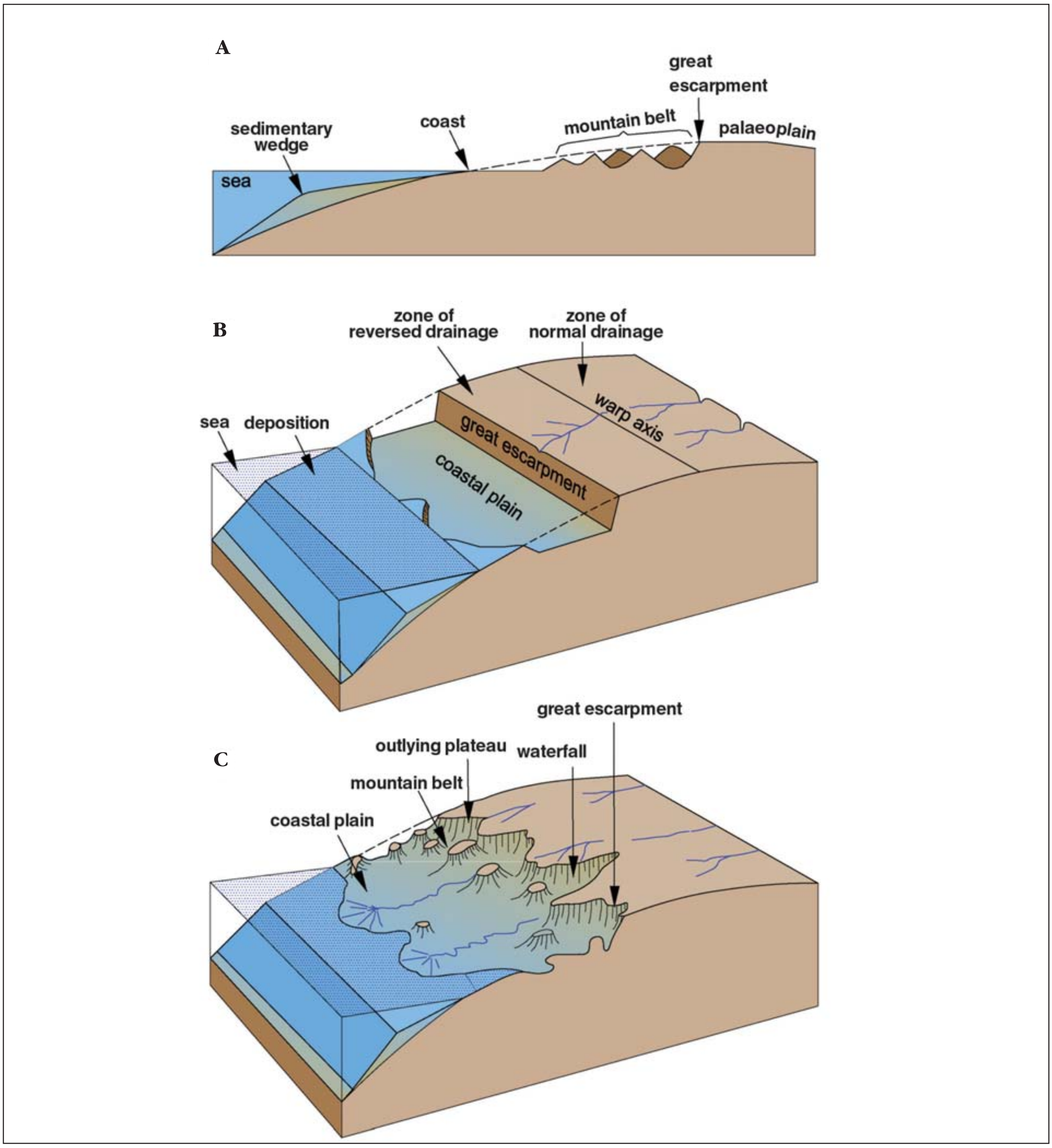

Figure 7. *(**A**) model developed by Ollier and Pain (1997) for the formation of a continental marginal escarpment following continental uplift. Diagram A shows the redistribution of eroded material to form an offshore sedimentary wedge; isostatic compensation along the margin results in uplift, resulting in the formation of an anticlinal arc and local drainage reversal. Headward erosion by the rivers produces the marginal escarpment (diagram (**B**). Diagram (**C**) illustrates the main geomorphological features of the resulting margin.*

the African Erosion Surface, also conforms to that of the base of the Karoo Supergroup and to the present day topography. The co-incidence of these three surfaces further supports the notion that their present general form is the result of post-planation uplift of southern Africa.

The implication is that the African Surface developed on the coastal regions was continuous with the African Surface in the interior which, at its final development, extended from coast to coast, rising to a maximum elevation of probably only a few hundred metres above sea level. This conclusion differs from the interpretation of Partridge and Maud (1987) who believed that the coastal and inland portions of the African Surface were separated by an escarpment dating back to the break-up of Gondwana and evolved simultaneously but separately, but concurs with the conclusions of du Toit (1939), Burke (1996) (see figure 26 of Burke, 1996) and Burke and Gunnell (2008).

Partridge (1998) and Partridge and Maud (1987, 2000) envisaged uplift of southern Africa in two stages: the first occurred in the early Miocene and the second and larger episode of uplift occurred in the Pliocene. These episodes purportedly initiated renewed erosion in the interior (the Post-Africa I and Post-African II cycles), resulting in stripping of much of the laterite surface and also causing incision of the meandering rivers on the coastal segment. In contrast, Burke (1996) and Burke and Gunnell (2008) proposed that uplift commenced at about 30 Ma and has been continuous. Roberts and White (2010) examined the uplift history of southern Africa using the longitudinal profiles of rivers. Their models indicate that uplift commenced about 30 Ma and was essentially uniform (and on-going) in the case of the Namibian highlands, but occurred in two pulses in the Angolan and Lesotho Highland regions, the first peaking at 20 to 25 Ma and the second at 5 to 10 Ma.

The overall shape of the southern African land surface, namely an elevated central plateau with a gentle depression in the middle, bordered by escarpments that are surrounded by coastal plains (Figure 4), is not unique to the subcontinent. Ollier (1985) and Ollier and Pain (1997) explored the paths that could lead to uplift and the observed morphology. One of these is particularly apposite, namely uplift of the continent to create an escarpment, followed by retreat of the escarpment due to erosion, with the eroded sediment being deposited offshore. Isostatic compensation occurs as the coastal plain widens, causing uplift of the escarpment, thereby creating a bulge and central depression and tilting the coastal margin seaward. Gilchrist and Summerfield (1994) modelled escarpment retreat and showed that a zone of uplift behind the escarpment would retreat inland as the escarpment is eroded back. A combination of these models seems to fit the southern African situation well (Figure 7).

The net effect of this sequence of events in southern Africa was the formation of a highly dissected coastal plain where erosion was focused, while far inland, erosion rates remained low. Dissection of the coastal plain left residuals capped by the palaeo-land surface, especially towards the coast, and continuing uplift exposed marine sediment lying on the palaeo-surface.

The ultimate origin of the epeirogenic uplift of southern Africa is a controversial topic, with little consensus. Nyblade and Robinson (1994), noting the

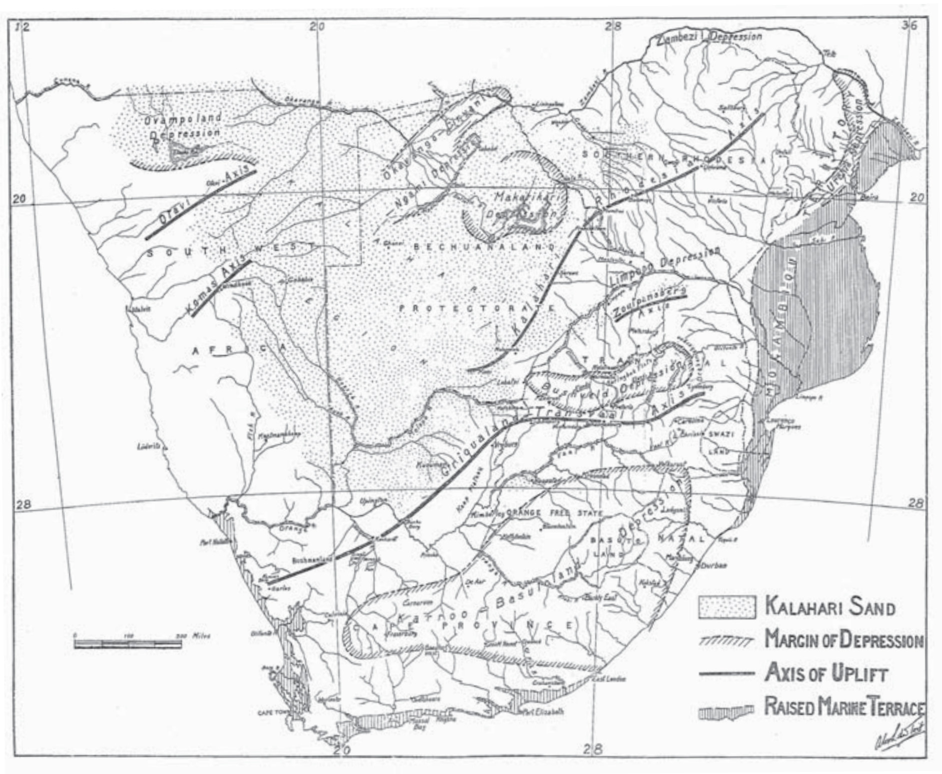

Figure 8. *Du Toit (1933) identified a basin and swell structure across southern Africa, shown here, which has had a profound influence on the geomorphological development of the region.*

uniqueness of the southern Africa plateau in the context of global hypsometry, coined the term the African Superswell for the phenomenon, and suggested it may be due to localized heating of the lithosphere. Lithgow-Bertelloni and Silver (1998) proposed that upwelling in the deep mantle was the cause. Burke (1996) ascribed the Superswell to heat accumulation arising from the fact that the African Plate came to rest about 30 Ma. However, observed drainage patterns appear to be incompatible with either notion (Moore et al., 2009). Uplift due to a deep-seated source should have caused doming with consequent symmetrical, radial drainage, whereas the actual topography in the interior is more basin-shaped. The origin of the Superswell remains an unsolved and contentions issue.

Continental flexure

Du Toit (1927; 1933; 1939) noted that the interior of southern Africa was broken into a number of smaller basins by linear arches or 'axes' of uplift, creating a basin and swell structure. The zones of uplift generally coincide with present day watersheds indicating their youthful origins (Figure 8). The most significant of these in the present context are the Griqualand-Transvaal Axis and the Kalahari-Zimbabwe Axis. These arches are asymmetrical, with the axes being due more to subsidence on their northwestern side than actual uplift along the ridge axes.

The basin and swell structure and the marginal bulge resulted in depressions in the interior in which sediment began accumulating, most notably the Kalahari Basin, and created new drainage divides.

The timing of commencement of warping is uncertain but could be constrained by the onset of sedimentation in the Kalahari Basin. The 92 Ma Orapa kimberlite pipe is overlain by crater-fill sediments, which in turn are overlain by Kalahari Group sediments, indicating that sedimentation commenced after 92 Ma. Partridge (1998) suggested that sedimentation commenced after 80 Ma, based on fluvial deposits of this age at Mahura Muthla, situated on the Transvaal-Griqualand Axis on the southern margin of the Kalahari Basin. He regarded these as having formed in a southerly flowing river and hence inferred that basining had not yet commenced at this time. However, Ward et al. (2004) have shown that this river drained to the north, implying that the Transvaal-Griqualand Axis had already formed by this time, implying that Kalahari sedimentation in central Botswana commenced between 80 and 90 Ma.

The Kalahari-Zimbabwe axis lay athwart the course of the Okavango-Zambezi-Limpopo River. Uplift along the axis evidently back-tilted a reach of the river system, resulting in the formation of a large lake or lakes in the interior. A large river could easily have dissected a path across the swell had the swell formed purely by uplift. The Kalahari Basin, as du Toit (1933) observed, formed more by subsidence, and the river was therefore unable to carve an exit. The floor of the Kalahari basin may even have been depressed below sea level, as is the case of the Lake Eyre Basin in Australia today – a product of intra-continental warping. The lakes in the Kalahari Basin may have spilled over the rim from time to time, possibly draining into the Kalahari River in the south via the Trans-Tswana River (McCarthy, 1983).

Sedimentation in the Kalahari Basin

The Kalahari Basin, in which sediments of the Kalahari Group accumulated, occupies a depression some 2700 km long and 1800 km wide with maximum depth of about 450 m. The thickest development occurs in southern and east-central Angola and north of the Okavango Delta (Haddon, 1999; 2005; Haddon and McCarthy, 2005).

The stratigraphy of the Kalahari Group is spatially extremely variable, making lateral correlations difficult. Nevertheless, there appears to be an overall pattern to the sequence. The base of the succession is commonly marked by a conglomerate that can be up to 120 m thick. It usually has a local provenance and appears to have been deposited fluvially in river valleys and less commonly as talus slope deposits. The conglomerate is often overlain by calcareous mudstone or siltstone (marls), reaching thicknesses of up to 100 m. Sandy layers are commonly intercalated with the mudstones, and occasionally gravel layers may also be present. In places, the conglomerate is absent and the mudstone lies on the pre-Kalahari basement. The mudstones generally coarsen upwards into sandstones, but the latter may also lie directly on bedrock. Gravel layers are occasionally intercalated in the sandstone. Although sedimentary structures are rare, the lower sandstones in the main appear to have been deposited under braided river conditions. The sandstones are cemented by carbonate, silica or hematite. They are locally overlain by gravel, especially in the Makgadikgadi area (the Letlhakane Stone Line of du Plessis, 1993), which appears to represent a deflationary lag deposit. The sandstone is commonly overlain by calcrete (Kalahari Limestone). These deposits are overlain by unconsolidated sand mainly of aeolian origin (the Kalahari Sand). The aeolian sand typically has a grain size of about 0.2 mm and is well sorted (Thomas and Shaw, 1991). The aeolian sands are transgressive across the underlying Kalahari sediments, and extend over a far wider area. Numerous pans are also developed across the Kalahari.

The Kalahari Group sediments reflect deposition in an environment undergoing profound climatic change. Early fluvial gravels and lacustrine deposits gave way to extensive braid-plains on which they deposited sand and gravel, possibly formed after infilling of the lakes. The overlying unconsolidated surficial aeolian sands reflect more intense aridification.

The climate of the Kalahari region

The mid-Cretaceous was characterized by exceptionally warm global climates. There were no ice caps and in general continental climates were humid, with limited areas of aridity (Burke and Gunnell, 2008). Southern

Africa was vegetated mainly by tropical, semi-deciduous forest, with temperate forest along the southern coast (Dingle et al., 1983; Bamford, 2000), dry woodland conditions across the northern Cape (Namaqualand, Bushmanland and the Ghaap Plateau, Bamford, 2000). Conditions peaked at about 53 Ma during the Paleocene-Eocene Thermal Maximum. As Africa moved northwards, conditions along the west coast became drier, and at Bogenfels in Namibia, dry to semi-arid woodland seem to have existed (Bamford, 2000). Northward migration of the continent ceased at about 30 Ma.

The Benguela Current became significantly colder, reducing the rainfall along the west coast of southern Africa. Uplift of southern Africa, which commenced at about this time, increased orographic rainfall on the east coast, thus reducing moisture levels in the interior, further exacerbating drying in the western interior. Cooling and drying continued throughout the Neogene.

The commencement of the Pleistocene was characterized by dramatic global cooling and the development of the Arctic ice cap. Sea surface temperatures off the west coast started falling between 3 and 4 Ma and by 1 Ma total temperature decrease amounted to as much as 10°C (Marlow et al., 2000). Oscillations in the ice cap volume at approximately 100 ka intervals commenced in the mid-Pleistocene (the glacial-interglacial cycles), which induced sea level fluctuations exceeding 100 m.

Anticyclonic circulation dominates air movements over southern Africa and moisture in the interior of southern Africa is derived primarily from the Indian Ocean. As the air masses move westwards from the warm ocean they rise against the eastern escarpment causing precipitation and a reduction in moisture levels, and these decline progressively in a westerly direction, resulting in a general decline in rainfall across southern Africa from east to west. The cold Benguela Current along the west coast ensures that air masses from the west contribute little to interior rainfall. Air over the ocean is cooled whilst the land surface is heated, causing strong southwesterly on-shore winds. These winds result in movement of ocean water away from the coast in a process known as Ekman transport, causing up-welling of cold water from the deeps. This further cools the Atlantic air, resulting in very dry conditions along the southwestern coast (Namib Desert).

The overall effect of these controls is the pronounced north to south and east to west rainfall gradients across the summer rainfall region of southern Africa (Figure 9). The Kalahari Basin straddles the very pronounced north-south climatic gradient of the region and in the north annual rainfall exceeds 1400 mm per annum. In contrast, the southern Kalahari has an annual rainfall below 200 mm per annum.

Rainfall in the lower rainfall regions occurs almost entirely in the summer months and is due largely to convective thunderstorms that develop in the late afternoons. Towards the equator in the north, however, seasonality becomes less pronounced, and in northern Angola, two rainfall peaks occur associated with the two yearly passes of the Inter Tropical Convergence Zone (Mendelsohn and el Obeid, 2004).

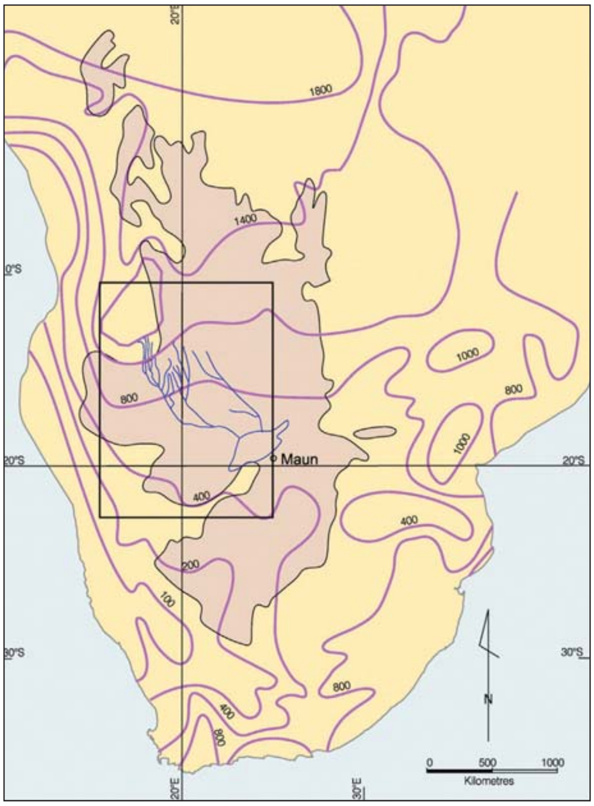

Figure 9. *Mean annual rainfall (in mm/a) over southern Africa (Thomas and Shaw, 1991). The distribution of the Kalahari Group and the Okavango-Kwando River systems are also shown.*

The high elevation of the interior of southern Africa results in generally lower temperatures than would be the case were elevation lower. The temperature extremes generally increase with decreasing rainfall due to excessive radiative heat loss in the clear skies of the lower rainfall regions, especially in winter. Humidity likewise decreases with decreasing rainfall. As a consequence, potential evapotranspiration (PET) increases with decreasing rainfall. In the drier southern portion of the region, measured evaporation is greatest in the summer months (wet season), whereas in the tropical regions in the north, evaporation is highest in winter (dry season). High evapotranspiration is an extremely important factor in creating the semi-arid conditions over much of the eastern Kalahari. In addition to decreasing total annual rainfall, the year-on-year variability of the precipitation also increases from north to south (Mendelsohn and el Obeid, 2004).

The controls on the present climate of southern Africa are characteristic of interglacial periods when generally warmer conditions prevail. Glacial cycles coincided with widespread cooling and drying, especially in southern Africa (Tyson and Partridge, 2000). The cold Benguela Current and its associated up-welling extend considerably further north at these times, reducing moisture levels in tropical Atlantic air and resulting in northward extension of the Namib Desert into the tropics (Burke and Gunnell, 2008). Throughout the Pleistocene, the climate of southern Africa has oscillated between these extremes and this has left an indelible mark on the geomorphology of southern Africa, which will be discussed in more detail below.

The Kalahari is characterized by aeolian sand that extends from the Equator in the north to the Orange River in the south, covering an area of 2.5 x 10^6 km^2. This is the largest continuous aeolian sand sheet on Earth. Linear dune systems occur as far north as 11°S, and indicate rainfall of less than 200 mm/yr in areas that today receive in excess of 1500 mm/yr. In stark contrast are the old shorelines of the lakes in the lower Okavango (Mababe, Ngami) and in the Makgadikgadi basin that indicate the former presence of large lakes, one of which was comparable to Lake Victoria in size, indicating a much wetter climate than today. These various features have been subject to intensive research over the past 50 years as they potentially provide a record of past climates not only for the Okavango but for southern Africa in general.

The lake basins have been intensively investigated, particularly by Paul Shaw and collaborators over many years (see Thomas and Shaw, 1991, and references cited therein). On a basis of their similar elevations, Shaw (1985) proposed the former existence of two major lake systems in the Okavango region: Lake Palaeo-Makgadikgadi, which stood at a lake level of 940 m to 945 m; and Lake Thamalakane at a level of 936 m. He estimated that an increase of about 160% over the present mean annual rainfall could have sustained Lake Thamalakane and an increase of 225% could also have sustained a lake in the Makgadikgadi basin at the 912 m level. He suggested that Palaeo-Lake Makgadikgadi existed about 40,000 to 35,000 years ago and Lake Thamalakane from 17,000 to 13,000 years BP and again from 2500 to 2000 years BP.

Burrough and Thomas (2008) and Burrough et al (2007, 2009a) undertook systematic luminescence dating of the main beach ridges of lakes in the lower Okavango and in the Makgadikgadi basin. Highstands occurred at 8.5 ± 0.2 kyr, 17.1 ± 1.6 kyr, 26.8 ± 1.2 kyr, 38.7 ± 1.8 kyr, 64.2 ± 2.0 kyr, 92.2 ± 1.5 kyr and 104.6 ± 3.1 kyr, and possibly at older periods (131 kyr, 211 kyr, 267 kyr and 288 kyr) (Figure 10). Burrough et al. (2009a) are equivocal on the causes of these high lake stand episodes, generally favouring climate change.

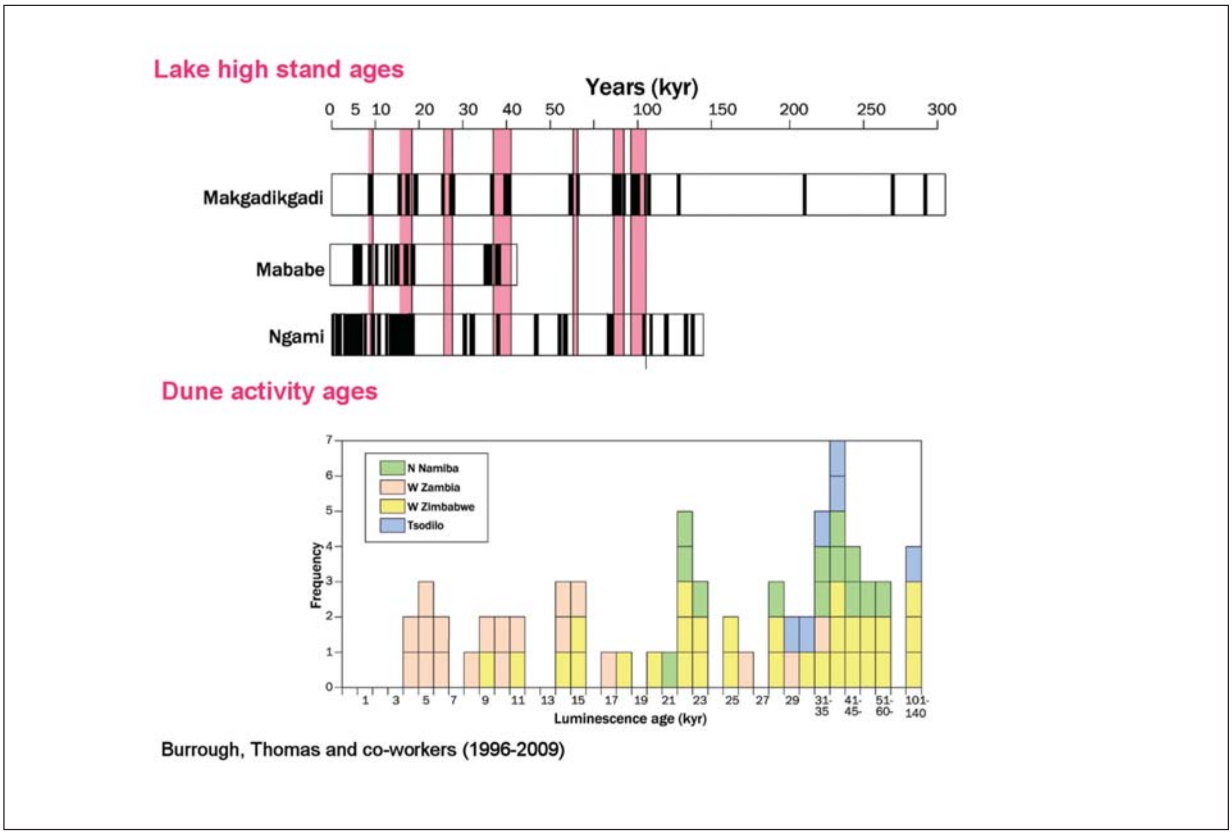

Figure 10. *Summary of luminescence ages of dunes and lake shorelines in and around the Okavango Delta (Thomas, Burrough and co-workers; see text for details).*

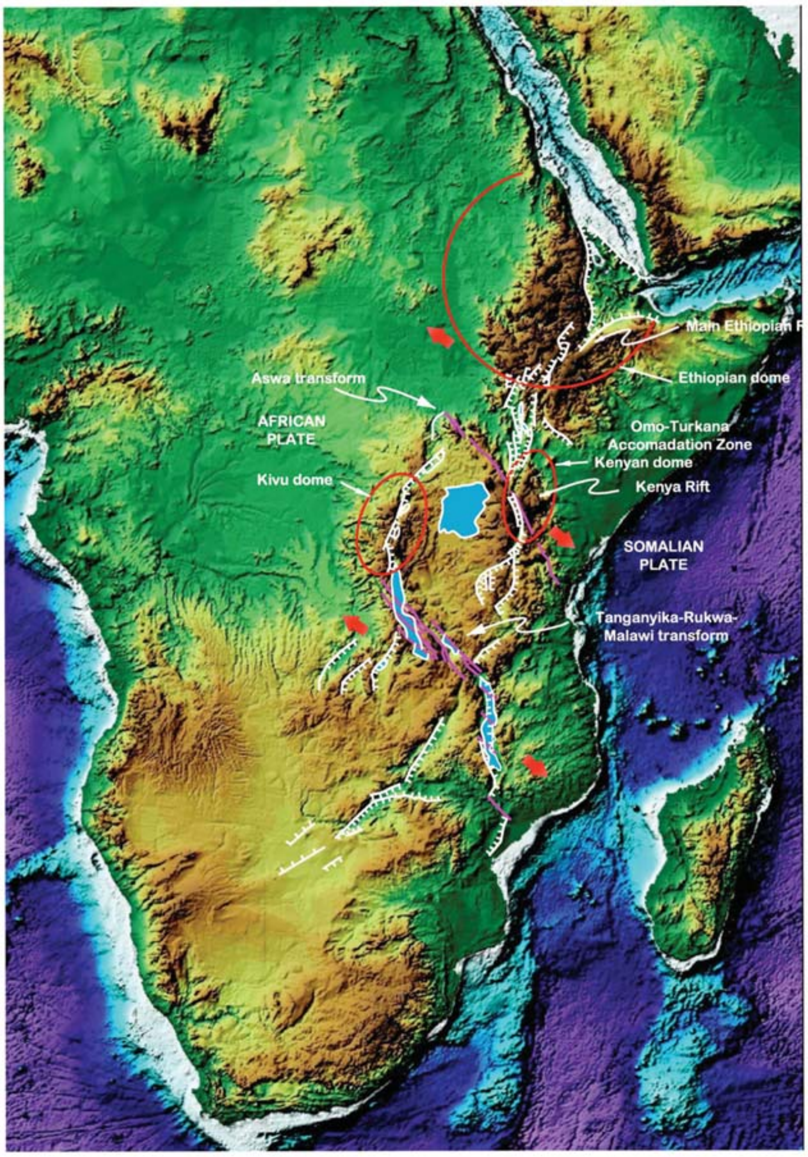

Figure 11. *The East African Rift system (structure based on Chorowicz, 2005; DEM by Nico de Koker, unpublished).*

The dune systems of the Kalahari have been the subject of detailed research over many decades particularly by Lancaster (see Lancaster, 2000, and references cited therein), Thomas and Shaw (1991, 2002) and Thomas et al. (2003) and references cited therein. The ages of dunes have been measured in many studies using luminescence methods. Linear dunes from the southern sector have yielded ages largely in the range 7 to 17 kyr (Thomas and Shaw, 2002), whereas those in the northern and eastern sectors are more distributed (Figure 10). It is likely that the ages measured do not reflect the full range of dune activity in the region. As pointed out by McFarlane and Eckardt (2007) the linear dunes are clearly very old features. The basic pattern evident today was most probably formed early in the Pleistocene, and these ancient dune have been repeatedly reactivated during dry periods and degraded during wetter times over the past two million years. The majority of the large linear dunes are very degraded, notwithstanding periods of reactivation, attesting to their great age.

Although the data base is small, comparison of the data compiled in Figure 10 suggests that dune-building episodes have dominated over lake high stand episodes since the Last Glacial Maximum. The data seem to indicate that arid conditions have prevailed most of the time, interspersed with brief wet periods.

The East African Rift System

Du Toit (1933, 1939) correctly surmised that the late Tertiary and Quaternary geological development of southern Africa was influenced by the East African Rift System (EARS) in the form of warping and active faulting. It is now well established that rifting has had a profound influence on drainage in the region, and particularly the Zambezi-Okavango-Limpopo system. It is therefore necessary to review briefly the origin and development of the rift system.

The rift commences in Afar triangle at a triple junction where the Gulf of Aden and the Red Sea meet, and separates the African (Nubian) and Somali tectonic plates (Figure 11). There are two main branches to the EARS termed the Eastern Branch in the northeast and the Western Branch in the southwest, linked by the northwest-southeast striking Aswa transfer fault zone. The rifts consist of a series of graben structures that are linked by transfer faults. According to Chorowicz (2005) the transfer faults define the overall sense of divergence along the EARS, viz. northwest-southeast. Lakes frequently occupy the grabens. Lake Victoria is an exception: it does not lie in a rift but occupies the depression between the elevated flanks of the Eastern and Western Rifts.

The location of the rift system closely follows lines of weakness in the crust, generally the Pan-African mobile belts, and it appears that many of the major graben faults are reactivated Pan-African thrusts. Cratonic areas are generally avoided by the rifts.

The Eastern Rift commences in the southern portion of the Afar triangle and extends southwards as a well defined series of half grabens and becomes very prominent in the Kenyan dome (Gregory Rift). After crossing the Aswa transfer, the Eastern rift swings to the southwest and terminates in a series of widely spaced half grabens. There appears to be a further offset of the Eastern Rift to the southeast with a southerly continuation of rifting occurring offshore along the Davie Ridge, indicated by seismic activity. Mozambique thus appears to form a sub-plate separated from the Somalian plate by the Davie Ridge which Stamps et al. (2008) have termed the Rovuma plate.

The main rift movement is transferred to the Western Rift by the Aswa transfer zone. This rift becomes increasingly more prominent southwards into the Kivu dome and beyond with shoulders rising to over 3000 m around Lake Tanganyika. South of Lake Tanganyika, the rift is offset by the Tanganyika-Rukwa-Malawi fault zone. The main rift continues southwards through Lake Malawi where further offset occurs along the Zambezi fault zone to the Dombe graben and the Urema rift. To the southwest of Lake Malawi lies the elongated Luangwa valley and its continuation through the Kariba

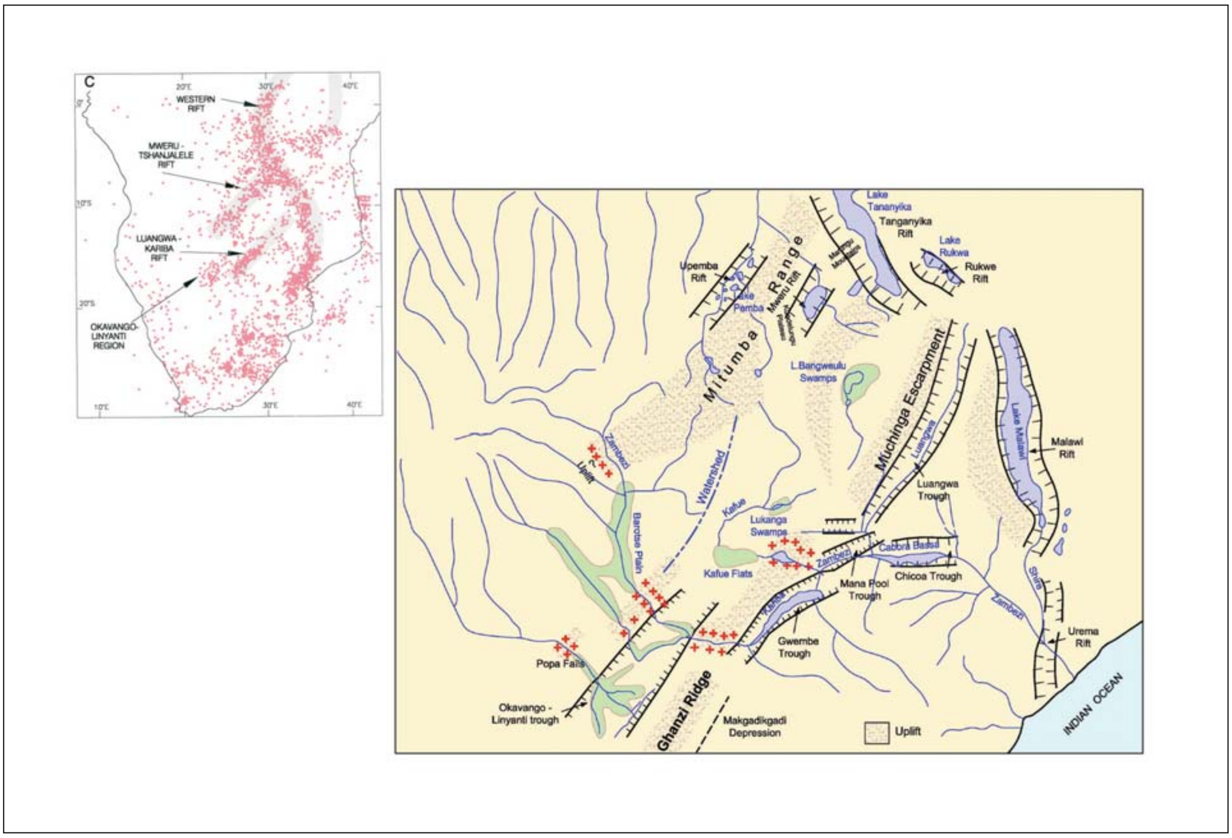

Figure 12. *The seismicity of the southern portion of the East African Rift system (inset) and its relationship to geomorphic features in the region.*

trough, which also occupy graben structures related to the EARS. Overall, the EARS has propagated southwards at the rate of between 2.5 and 5 cm/yr (Chorowicz, 2005).

Propagation of the East African Rift into southern Africa

Seismic activity was first detected in the Okavango region in 1952 when a series of tremors exceeding magnitude 5, and including one event of magnitude 6.7 (11 October, 1952), occurred in the area (Gane and Oliver, 1953). The regional significance of this activity was first recognized by Fairhead and Girdler (1971) and Reeves (1972), who suggested that it was linked to the seismic activity characterizing the EAR. A detailed study of seismicity in the region was undertaken by Scholz et al. (1976) who linked the Okavango activity to that around Kariba and the Luangua valley, and identified an association between the seismicity and northeasterly striking en echelon faults in the Okavango region. Fault plane solutions indicated dip slip motion on the faults and they suggested that the Okavango depression was a developing graben at the tip of an incipient zone of rifting that is following older basement structures, a view shared by Hutchins et al. (1976a,b), based on their analysis of satellite imagery and seismicity of the region.

The high level of seismic activity in the Kariba area led Fairhead and Girdler (1971) to suggest that it is reservoir-induced. However, Scholz et al (1976) pointed out that the activity remained at a high level long after filling of the lake and considered that gravitational loading was simply enhancing natural activity in the already active zone. This is also supported by the fact that the Cahora Bassa impoundment has a far lower level of seismicity than Kariba, notwithstanding its similar size and the fact that it is bounded by active faults (Moore et al., 2009). The possibility that the enhanced seismicity of the Okavango Delta is related to the gravitational loading effect of the seasonal flood was investigated by McCarthy et al. (1993a), who examined the timing of release of seismic energy in relation to the local hydrological cycle. No correspondence was found, suggesting that seasonal flooding does not significantly increase seismic activity.

The widespread distribution of swamps and lakes separated by bedrock river reaches in the region to the southwest and west of Lakes Tanganyika and Malawi (Figure 12) indicates that the rivers of the region are undergoing disturbance due to topographic shifts. There is also evidence that river courses are being influenced by zones of subsidence. The Luangwa, Mana Pools, Gwembe and Chicoa troughs are prominent topographic features and constrain major drainages. They are fault-bounded and generally flanked by elevated terrain. Their uplifted margins are transected by rivers that have cut significant gorges, such as the Batoka, Kafue, Kariba and Mupata gorges. They are also associated with seismic activity and represent reactivated Karoo age graben structures. In these respects they closely resemble active segments of the EARS such as the Lake Malawi and Lake Tanganyika rifts.

The Mweru rift to the north (Figure 12) strikes parallel to the Luangwa-Gwembe trough and is also seismically active. It is largely flanked by elevated terrain surrounding a central graben, which hosts Lake Mweru and terminates in an elevated tract that is seismically active (Kundalungu Plateau). To the northwest lies the Upemba Rift, also marked by somewhat elevated terrain with a central valley that accommodates several lakes, including Lake Upemba. Between the Mweru and Luangwa troughs are a series of depressions, which host swamps and lakes, the largest being Lake Bangweulu and its associated swamp. This depression has no associated seismic activity and appears to represent a hollow between the uplifts of the flanking rifts, similar to Lake Victoria. To the southwest lie the Lukanga swamps and the Kafue Flats, which, although aligned with the Bangweula depression, are in fact separated from it by more elevated terrain, and do not appear to represent an extension of the Bangweulu trough. To the northwest of these depressions is elevated ground which marks the watershed between the Kafue and Zambezi Rivers. Where the Zambezi and Cuando Rivers cross the projection of this watershed they form bedrock reaches, and along the Zambezi River in particular, there are several rapids. To the northwest of this reach is the Barotse Plain, a low gradient alluvial reach. A similar alluvial reach is also developed on the Cuando River, and the two are connected by the Mulonga-Matebele swamp zone. The presence of this alluvial reach indicates yet another depression, which shows some seismic activity. It is likely that this reach is actively subsiding and accommodates sediment brought down by the Zambezi in particular. The apex of the Panhandle region of the Okavango Delta is also represented by an uplifted region, and pre-Kalahari basement outcrops at Popa Falls. To the south lies a fault-bounded depression which hosts the Linyanti and Okavango Deltas, bounded in the south by the Ghanzi uplift. South of this uplift is the Makgadikgadi depression, also a zone of active faulting and seismicity (Mallick et al., 1981; Reeves 1972).

Whereas rifting in the main section of the EARS is concentrated in relatively narrow zones, it appears from the geographic features and the seismic record that the southerly extension of the EARS is much more diffuse. The main rifting is clearly taking place along the Luangwa-Gwembe trough, but other rifts are forming as well and in a rather diffuse pattern. It appears that rift-related NW-SE crustal extension is being spread over a broad zone in this southern section of the EARS. Rather than a single, narrow rift zone, there appear to be several small rifts of variable length in various stages of development, perhaps analogous the Basin and Range Province of the western USA. Similar conclusions have been made by Modisi et al. (2000) and Kinabo et al. (2007) who termed this diffuse zone the Southwestern Branch of the EARS. The notion that the Okavango-

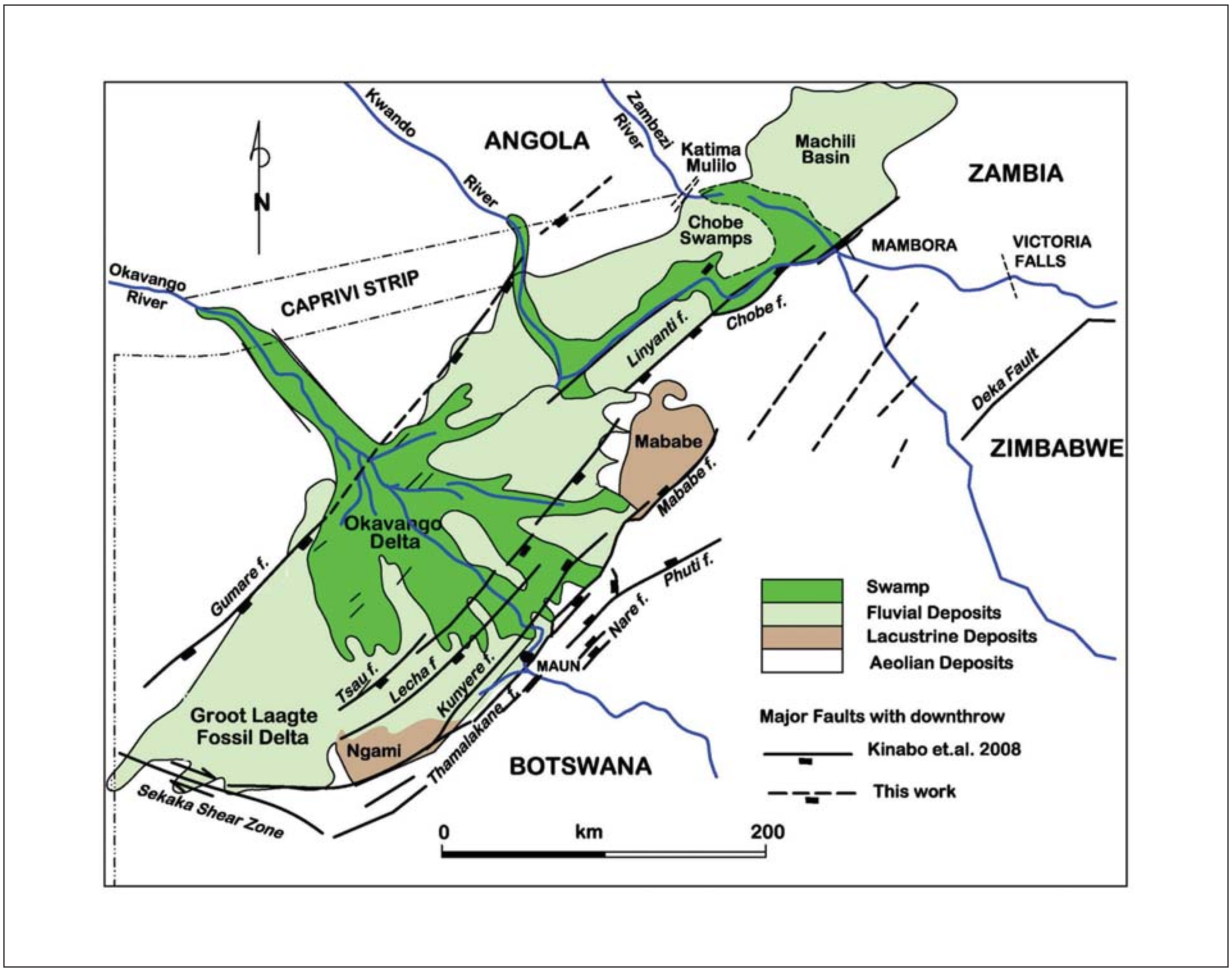

Figure 13. *The major structural features of the Okavango-Linyanti graben.*

Linyanti depression represents the active tip of the southerly propagating EARS is incorrect, but in fact represents one of several isolated zones of incipient rifting. It is possible that as rifting progresses, extension will gradually focus on one of these zones, whilst the others will become dormant.

The structure of the Okavango-Linyanti Depression

Du Toit first proposed that the Okavango region represented a graben structure related to the EARS. He stated: "It can be conjectured … that had it not been for another insidious enemy, the Okavango River might now be flowing majestically through the thirsty Kalahari to the Limpopo … But … crustal movements that had been operating … in Central and East Africa – disturbances that were connected with the development of the Great Rift Valley – extended to this part of the continent and slight warpings and faultings took place within it along axes directed south-west and north-east, that in some cases followed certain older lines of fracturing" (du Toit, 1927, p. 95). He stated further that " … owing to the operation of the earth movements referred to, an important section of the crust having a length of about 350 miles and a maximum width of 100, aligned in a north-east – south-west direction has … subsided athwart the Okavango-Linyanti-Zambezi river systems and disorganized them. A great hollow was thus formed in the red sand-veld, bounded in part by flexures, in part by fractures, and into it these three great rivers now had to discharge, dropping their load of fine grey sand and ultimately silting up the depression and producing a region of sandbanks, islands and swamps. Judging from what we see today, the rivers would have broken up into many meandering branches, repeatedly shifting their positions, while large areas would have been subject to seasonal inundation" (du Toit, 1927, page 96).

More detailed understanding of the structure of the Okavango region emerged from the analysis of aerial photographs of the region in the 1960s, and later, of satellite imagery and geophysical measurements. The first results from the analysis of satellite images were reported by Hutchins et al. (1976 a,b) who observed that the Okavango region was bounded by north-east – south-west trending faults, two in the south, the Kunyere and Thamalakane faults (Figure 13), and one in the north, the Gumare fault. They also identified a fault bounding the southern margin of the Linyanti swamps, which they termed the Chobe fault, and surmised that it may be an extension of the Gumare fault. Seismic refraction surveys conducted along the southern boundary faults (Greenwood and Carruthers, 1973, cited by Hutchins et al 1976a,b) indicated downthrows to the northwest on the Thamalakane fault and on the Kunyere fault.

Mallick et al. (1981) confirmed that the Okavango Delta region is an alluvium-filled graben structure and that it extends to the northeast across the Zambezi River to the partially fault-controlled Kafue Flats, a distance of 850 km, based on aerial photographic analysis. They noted that the bounding faults formed an en echelon pattern and their direction appears to be strongly influenced by the strike of Precambrian basement structures, and swings to a more east-southeasterly strike to the southwest of the basin. They note that Karoo strata underlie the graben fill, indicating that the graben formed by reactivation of Karoo to post Karoo-age structures, which in turn represent even older Damaride structures. They noted further that the Panhandle region of the Okavango Delta is probably controlled by northwesterly striking faults.

Thomas and Shaw (1991) and Shaw and Thomas (1992) introduced minor revisions, renaming the Chobe fault of Hutchins et al. (1976a,b) the Linyanti fault and assigning the name Chobe fault to a fault to the southeast, which borders the lower Chobe swamps (Figure 13). On the basis of these data, McCarthy et al. (1993a) interpreted the Okavango depression as a half graben bounded on its southeastern margin by the major Thamalakane and Kunyere listric faults, with the Gumare fault representing an antithetic fault. They proposed that the northwesterly striking faults represented a conjugate set of faults to the main listric faults.

The acquisition of high resolution topographic information and regional aeromagnetic and gravity data by the Botswana Geological Survey resulted in considerable refinement to the structural detail of the region (Modisi et al. 2000; Kinabo et al. 2007, 2008). In addition, very high resolution topographic, aeromagnetic and electromagnetic measurements made in the Maun region in the search for groundwater resources by the Department of Water Affairs further assisted in refining the structure (Modisi, 2000; Campbell et al., 2006).

The geophysical results revealed that the structure of the southeastern margin of the Okavango basin was more complex than previously envisaged. In addition to the Chobe, Linyanti, Gumare, Thamalakane and Kunyere faults, several other en echelon and/or parallel faults were recognized (Figure 13). The Thamalakane fault in fact consists of two parallel faults about 3 km apart (Campbell et al. 2006). Their combined scarp height varies from 5 to 12 m, whilst the Kunyere scarp is lower (5 m maximum) and decreases to zero in a northeasterly direction. The faults rotate to a east-northeasterly strike in the south and terminate against a west-northwesterly striking dextral structure, which Modisi et al (2000) termed the Sekaka Shear Zone. The land surface north of the Linyanti fault is tilting towards the northeast, diverting the flow in the Chobe River in that direction. This appears to be the most rapidly subsiding portion of the structure.

The geophysical signature of the Gumare fault suggests that it is of limited strike length and seems to terminate just south of the Panhandle section of the Delta. However, it is evident from the topography that this fault continues beyond the Panhandle to the head of the Linyanti swamp, as dune terminations characteristic of the fault in the Gumare area (McFarlane and Eckardt, 2007) also occur east of the Panhandle. The overall effect of the faulting in the Okavango region is to produce a set of grabens within grabens. The major displacements are taking place along the southern boundary of the Okavango graben, the Kunyere and Mababe being the most important in the south, confirming the half-graben structure. Kinabo et al. (2007) conclude that the en echelon faulting along the length of the graben has resulted in three depocentres: one centred on Lake Ngami, one around the Mababe Depression, and the third in the Linyanti-Chobe area. Faults towards the centres of the depocentres appear to have limited or no surface expression, indicating dormancy or waning activity, and suggest that the structure may have initiated as one or more fairly narrow grabens and has expanded by a process of fault piracy as en echelon faults become interlinked (probably both down dip and along strike). Kinabo et al. (2008) suggested that because of fault piracy, the Kunyere fault may be on the wane and its position is being usurped by the Thamalakane and Mababe faults, a conclusion based on the relative scarp heights of the faults.

The maximum width of the Okavango graben between the Gumare and Nare faults is 120 km, comparable to the widths of well developed rifts such as Tanganyika, but the topographic relief in the case of the Okavango is much less (120 m compared to >2000 m). Moreover, the fill in the case of the Okavango graben is modest (<600 m) compared to Tanganyika (>4 km), suggesting that during rifting the width of the depression is acquired early in its development (Modisi et al., 2000). Modisi et al. (2000) suggest that deep seated transverse structures such as the Sekaka Shear act as barriers to rift propagation and evolve to become accommodation or transfer structures that link separate components of the rift as it develops.

Modisi (2000) emphasized the role of northwesterly striking normal faults, particularly between the Thamalakane and Kunyere faults. These determine the locations of numerous small channels that cross between the two scarps (e.g. the Shashe and Boronyane channels). Detailed geophysical mapping by Campbell et al. (2006) confirmed the presence of a series of minor horsts and grabens between the two faults that are bounded by these northwesterly striking faults.

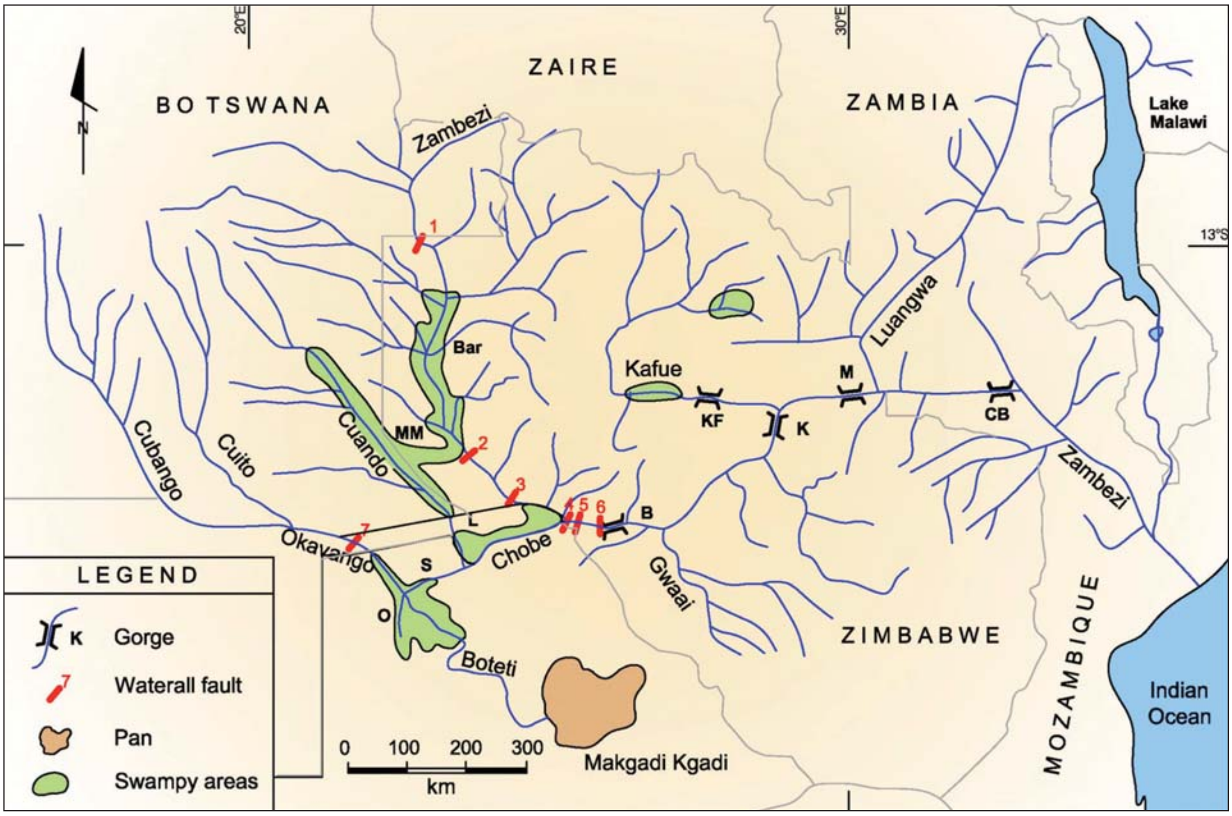

Figure 14. *The major geomorphological features of the Okavango, Cuando and Zambezi Rivers. Bar - Barotsi; MM – Mulonga-Matabele; S – Selinda; L – Linyanti; O – Okavango; B Batoka gorge; K – Kariba gorge; Kf – Kafue gorge; M – Mupata gorge; Mc – Machili; CB – Cabora Bassa gorge. Rapids and falls: 1 – Chavuma; 2 – Gonya; 3 - Katimo Mulilo; 4 – Mambova; 5 – Katombora; 6 – Victoria Falls; 7 – Popa Falls; (based on Moore and Larkin, 2001).*

Structures in this orientation are widespread in the region (Hutchins et al., 1976b; Sindling-Larsen et al., 1991; McCarthy et al., 1993a). The orientation of the Panhandle is determined by these fractures and there is a 5 m vertical displacement across this feature (Gumbricht et al., 2001). They also seem to determine the courses of the Zambezi River above Katima Mulilo and the Linyanti-Chobe link (Shaw and Thomas, 1988). Modisi (2000) suggested that extensional movement along these faults accommodates sagging along the length of the graben.

In studies of the southwest extensions of the EARS, workers have tended to focus attention on graben structures. However, horst blocks also form an important component of the region. Two in particular are important in the context of the Okavango-Linyanti Depression, namely the Ghanzi Ridge and the horst to the south of the Chobe fault. Uplift on the former is substantial, as almost the entire Kalahari Group has been eroded from the horst (Haddon, 1991), suggesting uplift of perhaps 200 m. Kalahari cover remains on the horst south of the Chobe fault, but topographic elevations indicate uplift of tens of metres. Both horst blocks have undoubtedly played an important role in influencing the outflow from the Okavango-Linyanti Depression.

The Okavango-Linyanti-Zambezi River system

The development of the Okavango, Linyanti and Zambezi Rivers has been strongly influenced by the tectonic history of the region. Although they came into existence with the breakup of Gondwana, their present configuration is evidently very young, (Lister, 1979; Moore and Larkin, 2001) Previously, the upper Zambezi formed part of the Limpopo drainage system, which also included the Okavango and Cuando as well as the Kafue and Luangwa Rivers (du Toit, 1927; Wellington, 1955; Moore and Larkin, 2001). The lower Limpopo River formed the distal trunk and occupied a rift valley along a failed arm of the Karoo triple junction.

The evidence for the former existence of this extensive drainage network has been compiled by Thomas and Shaw (1988, 1991), Moore and Larkin (2001) and Moore et al. (2007). In particular they note that the Cuando, Kafue and Luangwa Rivers all show abrupt changes in orientation of their courses, suggesting elbows of capture. The upper Zambezi also shows a marked inflexion in its course at Katima Mulilo, also suggesting capture (Figure 14). South bank tributaries of the Zambezi arising in central Zimbabwe flow in a northwesterly direction, unusual for tributaries of an easterly flowing river, and appear to have been captured via the Gwaai River and diverted into the Zambezi River. Finally, the inferred vast size of Lake Palaeo-Makgadikgadi suggests an additional source of water which only the Zambezi could have provided.

Wellington (1955) divided the Zambezi River into three discrete sections. The *lower* Zambezi lies downstream of Cahora Bassa Gorge on the coastal plain. The *middle* Zambezi lies between the Cahora Bassa Gorge and Mambova Falls and consists of a series of basins and deep gorges (Figure 14). The *upper* Zambezi extends upstream from Mambova Falls. In its lower reach it flows over an alluvial plain (the Ngami Depression of Wellington, 1955, and the Kalahari Rift of Shaw and Thomas, 1992), here termed the Okavango-Linyanti Depression. In this flat area the Zambezi is joined by the Chobe River, which arises in a belt of swamps to the southwest that includes the Linyanti Swamps. Upstream, this reach of the Zambezi ends at Katimo Mulilo, where the river traverses bedrock. This bedrock reach extends as far as Gonya Falls, upstream of which lies the Barotse Plain, another extensive, flat, alluvial reach. This region is linked to the Cuando River system to the west (where a similar alluvial reach is developed) by the Mulonga-Matabele floodplain. Upstream, the Zambezi again becomes a bedrock river to its source in a dambo near the DRC border.

Wellington (1955) and Thomas and Shaw (1988, 1991) noted that the course of the middle Zambezi closely follows a series of Karoo-bearing graben structures which host the Chicoa, Mana Pools and Gwembe troughs. Grabens also extend to the northeast to include the Luangwa valley. On the basis of relative elevations, Wellington concluded that the gorges at the exits to these various troughs could not be the result of superimposed drainages, which led them to favour a tectonic influence on the present course of the river. Thomas and Shaw (1988, 1991) also noted the important role played by capture of the former tributaries of the upper Zambezi, notably the Luangwa and Kafue rivers and several south-bank tributaries via the Gwaai River.

Moore and Larkin (2001) developed a detailed model of the evolution of the modern Zambezi River that incorporates many of the observations made by Wellington and later workers. They suggest that during the Cretaceous planation of southern Africa the upper Zambezi was linked to the Limpopo via the modern Shashi River (Figure 15). The Kafue flowed directly into the upper Zambezi via the Machili Basin. The Luangwa River continued in a southwesterly direction (via what is now the fault-bounded Gwembe Basin) to join the Zambezi near the present Makgadikgadi pans. They proposed that the Okavango, Cuando and upper Zambezi Rivers flowed in a southeasterly direction to join the Limpopo (Figure 15). At this time, the lower Zambezi River was a relatively minor coastal river which had formed on the eastern escarpment after opening of the Mozambique Channel, whilst rivers that today constitute the middle Zambezi were tributaries of the upper Zambezi and Limpopo. These river systems persisted throughout the Cretaceous and into the Paleogene, during which time the interior of the continent was reduced to a low-lying peneplain and the large Limpopo Delta was largely formed.

Moore and Larkin (2001) suggested that flexure along the Kalahari-Zimbabwe (Rhodesia) axis (the Ovamboland-Kalahari-Zimbabwe axis of Moore, 1999) severed the links between the Limpopo and the Okavango, Cuando and Zambezi Rivers and a large lake or lakes formed in the depression northwest of the axis. These rivers built deltas on the lake shore and fine-grained sediment began to collect on the distal parts of the lake bed, forming the clay deposits of the basal unit of the Kalahari Group.

Flexure also steepened the course of the lower Zambezi, steepening its course and causing it to erode headward into the interior, ultimately leading to its capture of the Luangwa River. The lowering of the base level of the Luangwa River and the increased discharge in the lower Zambezi accelerated erosion and the Luangwa valley, the Chicoa trough and the Cahora Bassa gorge were cut, thus initiating the formation of the middle Zambezi. Further headward erosion by the middle Zambezi led to the successive cutting of the Mupata and Kariba gorges, causing a reversal of flow in the now headless, lower reaches of the Luangwa River. A north bank tributary of the middle Zambezi captured the Kafue River and ultimately headward erosion by a tributary of the middle Zambezi resulted in capture of the upper Zambezi at Mambova Falls. The increased discharge downstream led to erosion of the Batoka gorge. The knick-point responsible presently forms the Victoria Falls. Moore et al. (2007) envisage that the capture of the upper Zambezi took place during the early Pleistocene, and Moore and Cotterill (2010) have estimated the rate of migration of the knick-point at somewhere between 0.042 to 0.052 m/yr and 0.067 to 0.080 m/yr. In this way, the upper Zambezi River became diverted and ceased to flow into Palaeo-Lake Makgadikgadi, although the link may have been periodically re-established by tectonic activity.

The implication is that the deep valleys representing the Luangwa, Gwembe, Mana Pools and Chicoa troughs formed mainly by erosion of Karoo sediments that occupy these late- to post-Karoo grabens, and it is for this reason that the Zambezi River has come to occupy a partially interconnected set of old graben structures – i.e. it is a superimposed system.

In contrast to Moore and Larkin (2001), Wellington (1955) and Thomas and Shaw (1988) placed more emphasis on the role of faulting in shaping the course of the middle Zambezi. The close correspondence between the course of the river and seismically active zones (Figure 12) suggest that the effect of tectonism on the middle Zambezi needs further examination.

The Okavango-Linyanti Depression is a northeasterly striking composite graben structure, flanked along its southeastern margin by horst blocks. Subsidence of the Depression, probably accompanied by northeast tilting,

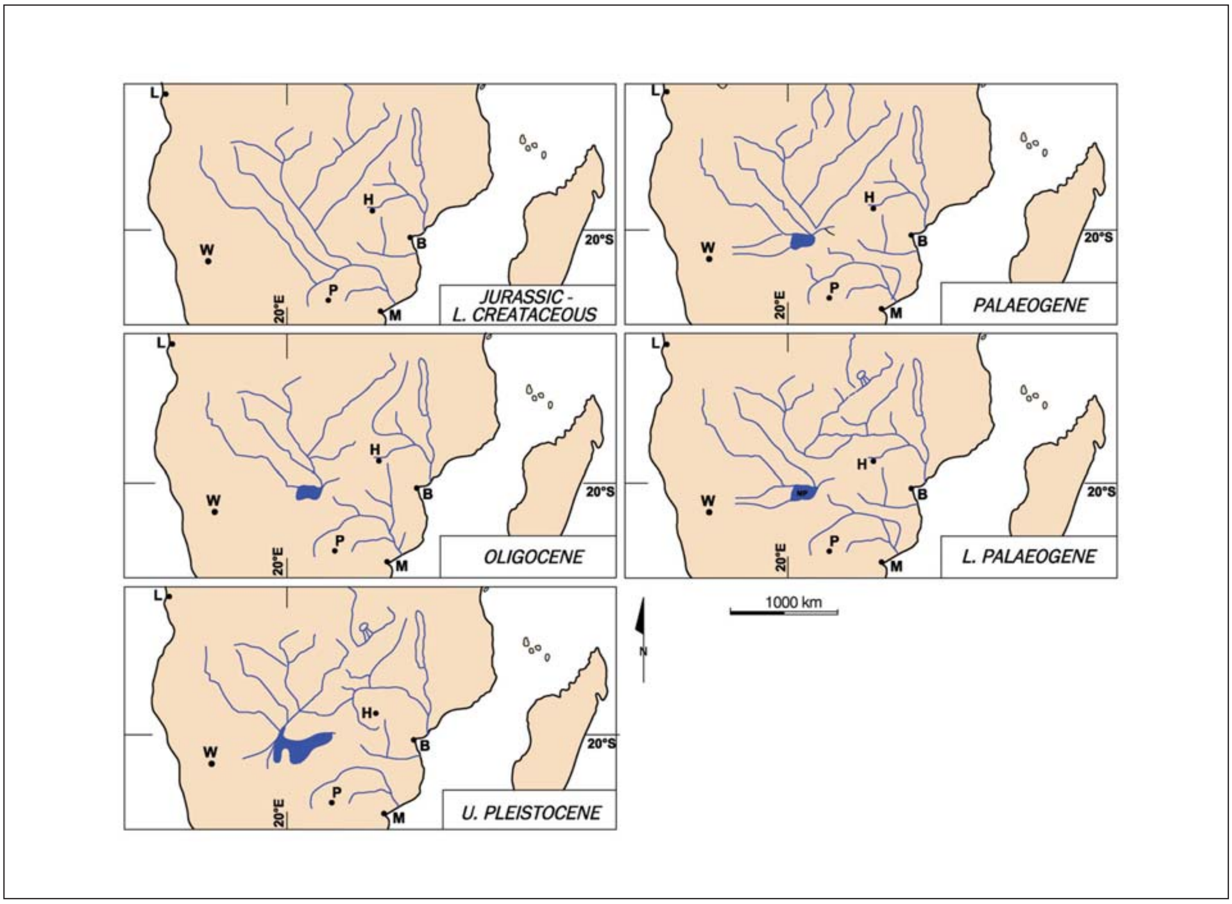

Figure 15. *Schematic diagrams illustrating the evolution of the Zambezi River system (after Moore and Larkin, 2001).*

is shifting flow of the Cuando River and, on occasion, the Okavango River (via the Selinda Spillway), to the northeast to join the Zambezi. The Zambezi River exits the graben across the Chobe fault at the Mambova Falls and is currently incising its course across the horst.

The Kafue River is believed to have formerly continued its course in a southwesterly direction across the Machili Flats to join the Zambezi above Mambova Falls (Thomas and Shaw, 1991). The course appears to have been severed, possibly by northwesterly striking faults (Moore et al. 2007), creating a closed basin which accommodated a lake (Lake Patrick; Moore et al. 2007). The Kafue has overtopped this depression on its southern side and outflow from the flats has carved the Kafue gorge.

These developments suggest the Middle Zambezi may well have been assembled in a piecemeal fashion by a series of tectonically-driven river captures in an analogous manner to the capture of the Okavango-Linyanti River that is currently taking place. The process appears to be initiated by graben formation. Water accumulates in the graben and then overtops a bounding horst or is captured by a tributary from an adjacent river that is eroding headward into the horst, thereby initiating outflow and gorge cutting. Subsidence of the graben continues and the gorge is gradually deepened, resulting in a basin (the graben) with an outlet via a gorge.

Tributaries discharging into the subsiding troughs became energized by subsidence of the trough, resulting in knick point formation, which migrate upstream, causing river capture. For example, the Gwaai River captured tributaries that had previously drained towards the west (Thomas and Shaw, 1988), whilst another tributary from the Gwembe trough breached the horst bounding the southern margin of the Okavango-Linyanti Depression, thereby linking the upper and middle Zambezi Rivers. Such a breach probably coincided with a highstand phase of Lake Palaeo-Makgadikgadi.

Water from the Zambezi was now able to flow across the Karoo basalt-covered horst to the Gwembe trough where it spilled over the faulted margin of the basalt (Wellington, 1955) initiating the formation of the Batoka Gorge. The resulting knick-point worked its way headward, exploiting joints and fractures, and currently manifests itself as Victoria Falls (Wellington, 1955). The capture process continues: the Linyanti (Cuando) River has only recently been captured and capture of the Okavango River is in its early stages (Wellington, 1955; Gumbricht et al., 2001), but will be fully completed once the Victoria Falls knick point breaches the Chobe fault.

It is thus suggested that the modern Okavango and Zambezi was assembled progressively, the latter by successive river captures, initially of its tributaries and finally of the upper Zambezi itself. It is proposed that the process was and still is being driven by tectonism associated with the southwestward propagation of the East African Rift Valley system, in this case mainly via the reactivation of late- to post-Karoo graben structures. The modern Okavango Delta and Zambezi River system are therefore young and their development probably only commenced once rifting had reached the sub-continent, commencing with the formation of the middle Zambezi, probably in the late Miocene to early Pliocene. However, the upper Zambezi is ancient and appears to date back to the break-up of Gondwana. Like many large rivers of the world, it therefore appears that the Zambezi is largely a rift-controlled river system (Potter, 1978).

The Okavango River catchment

The usual convention in river nomenclature is to assign a particular name to a river from its source to either its mouth or the point where it joins a more significant river. This is not the case in northern Botswana. Here, different reaches of the same river are often assigned different names. For example, the Cuando River, on entering the Linyanti swamps becomes the Linyanti River and, on turning southeast on its way to join the Zambezi becomes the Chobe River. According to Stigand (1923) this system arose because different river reaches were named according to the tribe or faction which occupied and controlled them. The name Okavango, in the view of Namibians, applies only that section of the river within Botswana. In Namibia, the river is known as the Kavango River (although Mendelsohn and el Obeid, 2004, refer to this reach as the Okavango). The name Kavango applies only in the Caprivi. In Botswana, the name Okavango only applies in the Panhandle region of the Okavango Delta, and downstream the name changes to Ngoqa. In Angola, the Kavango River becomes the Cubango River. This name, however, applies to one river that can be traced to its source, which is situated near Huambo in the Angolan (Bié) Highlands (Mendelsohn and el Obeid, 2004). This, then, is the true source of the Okavango River.

The source of the Cubango River lies at an elevation of about 1760 m, where it forms one of several parallel, north-south aligned tributaries. Each of these is fed by short lateral tributaries arranged perpendicular to the trunk stream in a trellis pattern. These tributaries appear to have a dambo-like character, and may be primarily groundwater tapping rather than carrying surface run-off. Further east, this drainage pattern is replaced by a more usual branching tributary network (Mendelsohn and el Obeid, 2004). To the east lies the second major tributary system of the Okavango, namely the Cuito River. It rises on the eastern flank of the Highlands at an elevation of about 1530 m (Figure 16), and also shows trellis drainage. The Cuando also arises in this area.

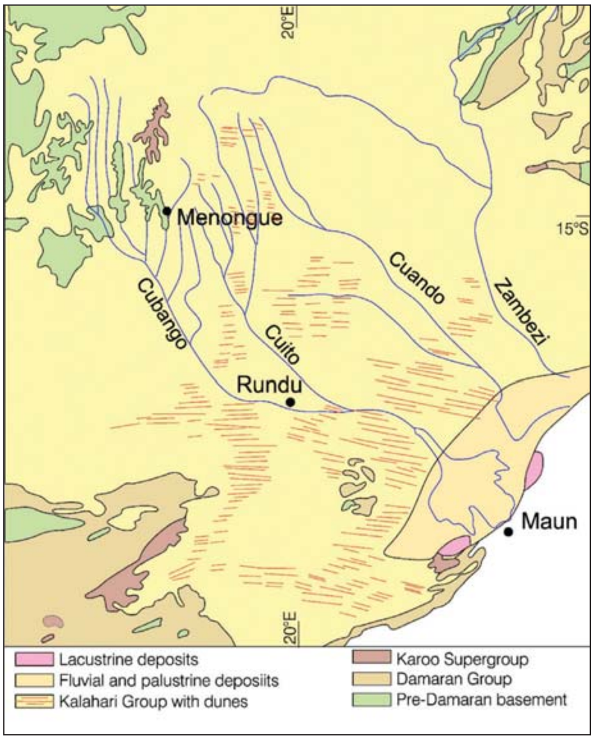

Figure 16. *Geology of the catchment of the Okavango River and Delta (after Mendelsohn and el Obeid, 2004). See Figure 9 for location.*

The gradient of the Cuito is lower than that of the Cubango and the mid reach of the lower Cuito in particular is characterized by a very low gradient (about 1:8000) with wide floodplains with fairly extensive marshes (e.g. Figure 14).

The upper catchments of the Cubango and Cuito provide virtually all of the discharge because of the higher rainfall in this region and there are no active tributaries to these rivers in their southern, lower reaches. The active areas of the two drainage basins are: Cuito 44,950 km^2; and Cubango 66,300 km^2, with a further 45,000 km^2 of combined inactive catchment (Mendelsohn and el Obeid, 2004).

Almost the entire catchment is underlain by Kalahari Group sediment (Figure 16), including both the upper sands as well as underlying calcretes and silcretes, some of which are exposed in the river valleys (Wellington, 1955; Mendelsohn and el Obeid, 2004). In portions of the upper catchment of the Cubango River, the river and several of its tributaries have incised through the Kalahari Group and exposed basement rocks, which consist largely of granite. In contrast, the entire Cuito catchment is underlain by the Kalahari Group.

The morphology of the Okavango Delta and its surrounding areas

The Okavango region was until recently very remote and although excellent map coverage was available (Botswana Department of Surveys and Mapping), there was very little in the way of elevation data, apart from isolated spot heights (e.g. du Toit, 1927; Jeffares, 1937), as well as elevations along the length of Chief's Island in the central part of the delta. The few data available made it possible to construct generalized contour maps of the region (UNDP, 1976; Cook, 1980), which clearly showed the curved form of an alluvial fan.

Although deltas and alluvial fans may appear to be similar in plan view, they differ fundamentally in their architecture and mode of formation. Alluvial fans have a sloping, conical upper surface whereas deltas have a relatively flat surface; a fan forms by the deposition of sediment due to loss of confinement of flow and builds vertically, whereas a delta forms by sedimentation caused by a decrease in flow velocity as a channel enters a standing water body and it builds outwards into the water body. The Okavango Delta is therefore an alluvial fan (McCarthy and Cadle, 1995), although the name Okavango Delta is so well entrenched, the term Delta continues to be used. Stanistreet and McCarthy (1993) examined the variety of fans that had been recorded and suggested that the Delta represented a new class of alluvial fan which they termed "losimean" (an acronym for *lo*w *si*nuosity *mean*dering) fans.

The broad features of the Okavango region are summarized in Figure 17 (Hutchins et al., 1976 a,b; Cook, 1980; Mallick et al., 1981; Thomas and Shaw, 1991). The Ngami Depression accommodates three alluvial fan systems: the easterly sloping Groot Laagte fan in the west; the southerly sloping Okavango fan in the centre; and the southerly sloping Linyanti fan in the east. The extreme eastern portion is traversed by the Zambezi River. The Groot Laagte and Okavango fans are separated by a discrete divide (Mallick et al., 1981), and Lake Ngami occupies the lowest position on this divide against the scarp of the Kunyere fault. The division between the Linyanti and Okavango fans is less clear, and it appears that the Linyanti system once flowed southwards into the Mababe Depression which occupies the low region between the Okavango fan and what would have been the distal end of the Linyanti fan. However, faulting appears to have truncated the Linyanti fan and tilted its surface towards the east, so the terminal distributary of this fan (Linyanti River) flows eastwards and then overtops the Linyanti fault and flows south as the Chobe River where it ponds against the Chobe fault scarp. From there it again flows east to join the Zambezi River at Mambova (Mallick et al., 1981).

Early differential GPS surveys revealed that the Okavango exhibits two discrete gradients: the Panhandle, with average gradient of 1:5570; and the fan portion of the Delta with a gradient of 1:3400.

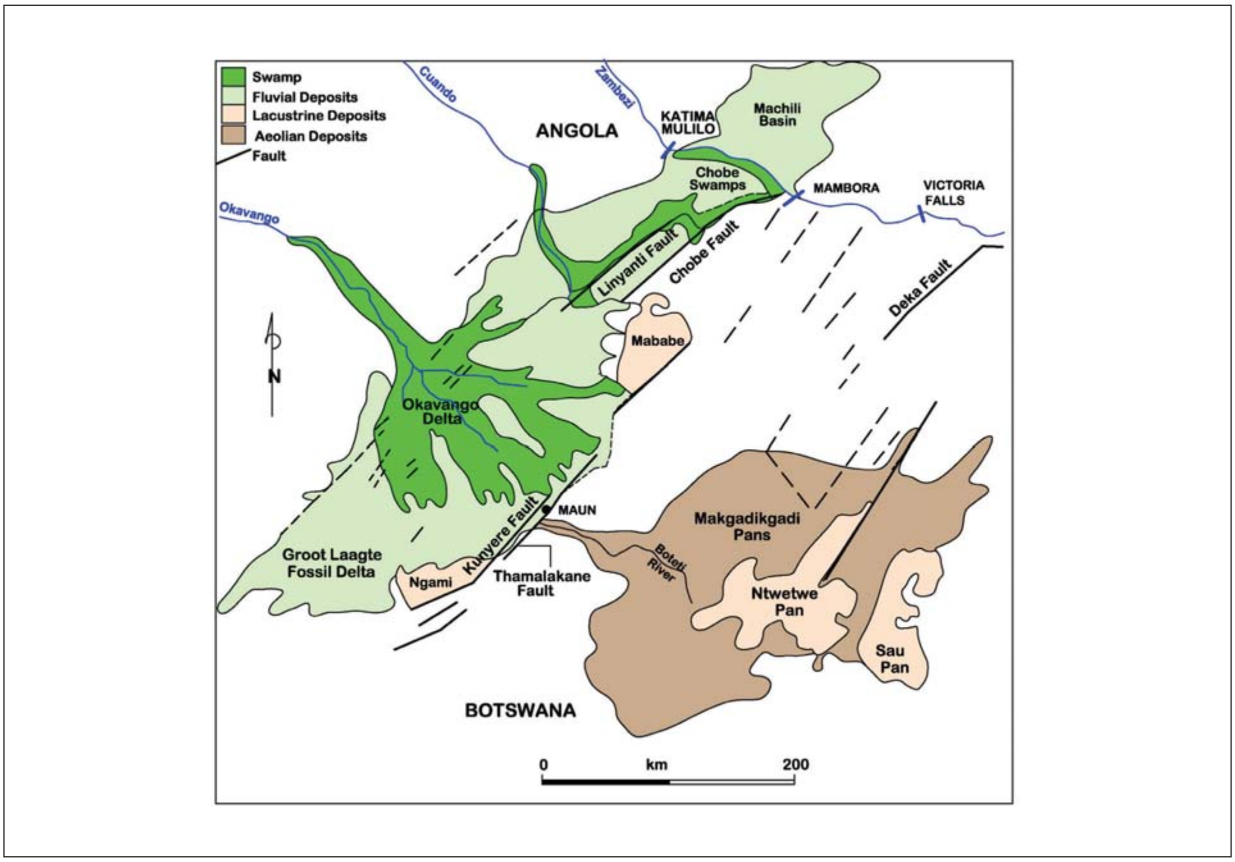

Figure 17. *Map showing the major geological and geomorphological features of the Okavango Delta and environs (adapted from Shaw and Thomas, 1992).*

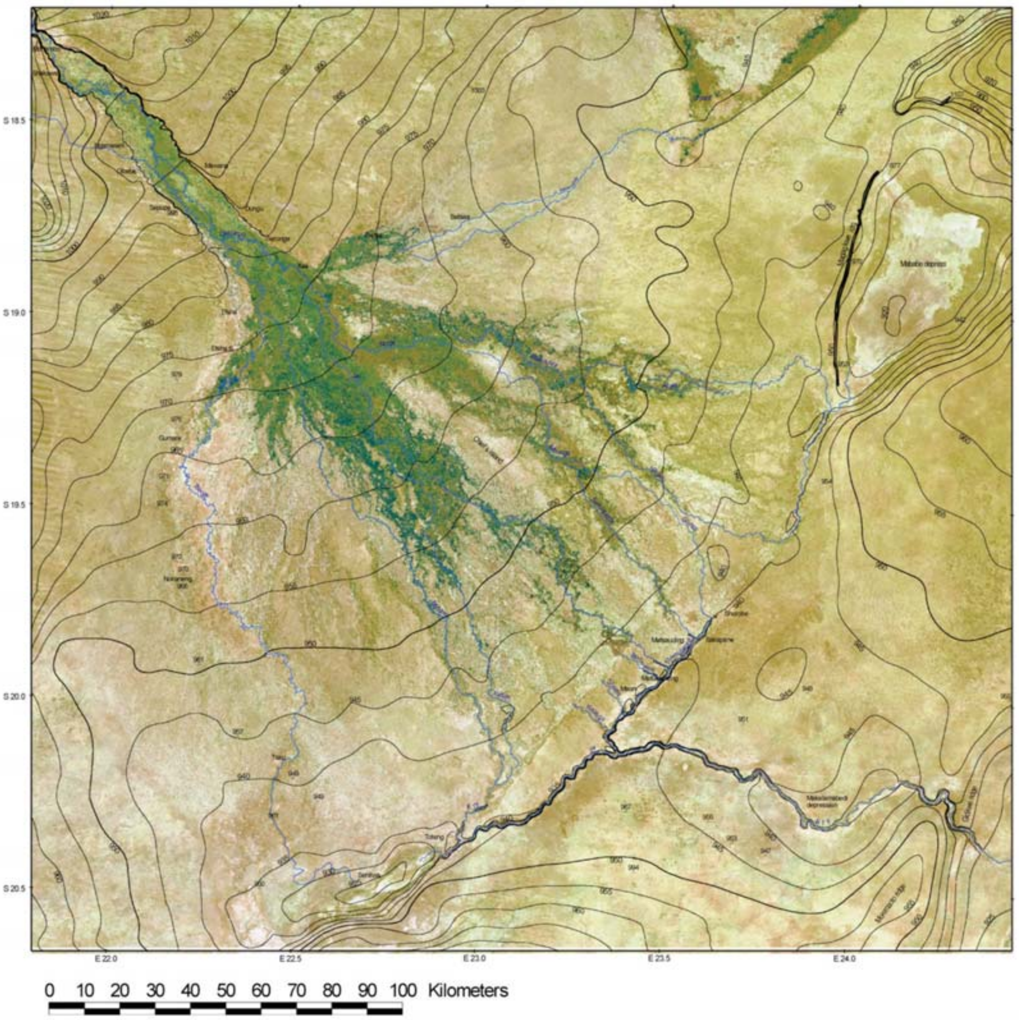

Figure 18. *Topographic contour map of the Okavango delta superimposed on a natural colour LANDSAT image (Gumbricht et al., 2001).*

Perturbations were evident along both segments. The total fall across the Delta is about 60 m over a horizontal distance of 250 km (Merry et al., 1998; McCarthy et al., 1997). Gumbricht et al. (2001, 2005) used GPS-derived elevation data to produce the first accurate contour map of the region (Figure 18). Subsequent to their work, NASA released the Shuttle Radar Topographic Mission (SRTM) data set for Africa which has provided topographic detail of the region not previously available.

Gumbricht et al. (2001) divided the Okavango region into three major structural domains on the basis of the topography: the northerly *Panhandle domain*; the central *Delta domain*; and the southern *Boteti domain*.

The terrain flanking the floodplain of the Panhandle in the *Panhandle domain* is characterized by the presence of west-northwest orientated liner dunes, which are especially prominent on the western side of the Panhandle. They are heavily degraded (Jacobberger and Hooper, 1991; McFarlane and Eckardt, 2007). The terrain flanking the Panhandle slopes to the southeast with a gradient of 1:2000, steeper than the Panhandle floodplain itself, and as a consequence the flanks of the Panhandle become more elevated to the north, culminating in the cleft at Popa Falls. The contours astride the Panhandle are off-set (Figure 18) which suggests a fault coincident with the Panhandle with a 5 m downthrow on the western side. The Panhandle terminates at the Popa Falls where pre-Kalahari Group basement is exposed. At its southern end, the Panhandle floodplain spills across the lower, western shoulder.

The Panhandle domain terminates abruptly in the south, and gives way to the curved contours of the Delta domain. The boundary coincides with the position of the Gomare fault and its inferred easterly extension (Figure 17). The contours in the Panhandle domain are parallel to the major northeasterly striking faults and suggest a roll-over structure into the rift graben (e.g Chorowicz, 2005).

The *Delta domain* is located between the Gumare and Kunyere faults (Figures 17, 18) and extends to the northeast to include the Linyanti swamps. The arcuate contours define the surface of the alluvial fan which is the dominant feature of this domain. In the west, the contours swing towards the south, reflecting the distal edge of the Groot Laagte fan, and Lake Ngami lies between the two fans. The Mababe Depression is located at the southeastern termination of the fan down-slope from the Linyanti fan. The gradient on the Okavango fan is remarkably uniform, averaging 1:3550. It is also remarkably smooth, and does not deviate by more than 2 m from a perfect conical surface. Exceptions occur where the arcs terminate in the east against the Gomare fault extension, where there appears to be a noticeable depression. This depression is occupied by the Selinda Spillway (Magwegqana), which links the Okavango and Linyanti swamps. The spatially more precise topographic representation afforded by the DEM reveals that the Delta surface is covered by a lace-like pattern of sinuous ridges and troughs (Figure 19).

The fan terminates against faults in the south. The Kunyere fault has a prominent scarp in the southwest, but scarp height decreases and the fault trace disappears to the northeast possibly due to the conical form of the fan, which actually overtops the Kunyere scarp. Further to the northeast, the Mababe fault appears virtually on an extension of the Kunyere fault, and its scarp increases in height to the northeast where it forms the southeastern margin of the Mababe Depression. Overtopping by the fan has created a local high point against the Thamalakane fault scarp, which acts as a flow-shed and flood water arriving at this point from the north splits, some flowing northeast towards the Mababe Depression and some southwest towards Lake Ngami (du Toit, 1926; Wilson and Dincer, 1976). The scarp height of the Thamalakane fault is greater than that of the Kunyere and Mababe.

Floodwater reaching the toe of the fan dams up against the scarp of the Kunyere-Mababe fault, except where the scarp has been overtopped. The scarp of the Kuyere fault has been breached at several points by fault-controlled channels that run perpendicular to the fault (Boro, Boronyane, Shashe, Nxotega; Figure 18), which allows some of the water to spill across into the Thamalakane River, which runs along scarp of the Thamalakane fault. The Thamalakane scarp has been

Figure 19. *SRTM-derived DEM of the upper Okavango Delta showing the lace-like pattern of ridges and depressions on the fan surface (prepared by Gordon Cooper).*

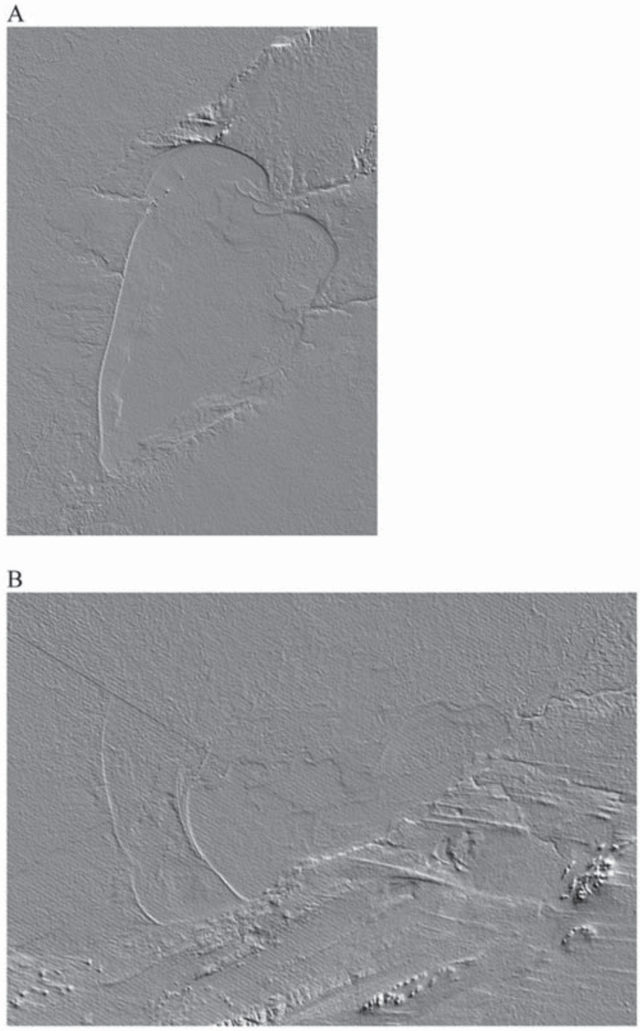

Figure 20. *SRTM DEMs of the Mababe Depression (**A**) and Lake Ngami (**B**) (prepared by Gordon Cooper).*

breached to form the Boteti River. The local gradients in the region of the breach are such that flow bifurcates, some flowing away from the Delta along the Boteti River and some continuing along the scarp to Lake Ngami via the Nhabe (Lake) River. The Boteti River does not cross the Thamalakane fault scarp at its lowest point, which suggests that erosion by the Boteti River has kept pace with on-going fault displacement.

In the northeastern portion of the Delta domain the contours of the fan turn abruptly to the northeast (Figure 18), the curvature reflecting the fan relating to the Cuando River (Linyanti fan). This was formerly much larger, but has been truncated by the Linyanti fault and the reduced fan is now represented by the Linyanti swamps. Rotational movement on the Linyanti fault has back-tilted the upthrown block to the southeast, resulting in a distinct lip along the fault line. Towards its western end the lip is breached by the Savuti channel, which flows southeastwards towards the Mababe Depression. During wet periods, flow in the channel reaches the Depression, forming the Savuti marsh. However, because of the dry period during the 1980s and 1990s, the channel and marsh largely dried up, but the situation has now reversed. Parallel to the Linyanti fault, but displaced to the east, is the Chobe fault, which has a much higher scarp. At its western end the fault terminates abruptly, as does the horst block, possibly due to truncation against a northwesterly striking fault. The Mababe Depression occupies the topographic low bounded by this horst in the northeast, the Mababe fault in the southeast and the distal edges of the Okavango and palaeo-Linyanti fans in the northwest. Gumbricht et al. (2001) noted that the surface of the Mababe Depression has been tilted towards the southwest.

The *Boteti domain* lies southeast of the Thamalakane fault and straddles the Ghanzi Ridge, which separates the Ngami Depression from the Makgadikgadi Depression. The domain is characterized by a central depression, termed the Makalamabedi Depression, bordered in the east by the broad Moremaoto Ridge (Figure 18) beyond which lies the Makgadikgadi Depression. The Boteti River flows across the depression and is fairly deeply incised over most of its length, especially where it crosses the Moremaoto Ridge. The river is underfit, indicating much higher discharges in the past (Cooke and Verstappen, 1984; Shaw et al, 1988).

The basins of Lake Ngami and the Mababe, Makalamabedi and Makgadikgadi Depressions all show the development of multiple palaeo-shorelines (e.g Figure 20a,b) (Grove, 1969; Cooke, 1980; Mallick et al., 1981; Cooke and Verstappen, 1984; Shaw, 1985; Shaw and Cooke, 1986). All possess a major sand ridge along their western margins. The east-facing slope of these sand ridges is generally fairly well defined. These are offshore bars formed by wind-generated wave action on the leeward side of the lakes that once occupied the depressions (Grove, 1969; Shaw, 1988).

Gumbricht et al. (2001) used the most prominent ridge on each of these depressions to estimate relative tectonic movement and concluded that the Gidikwe Ridge has been uplifted in the north and has subsided in the south; the Makalamabedi Depression has experienced a few metres up uplift, Lake Ngami has experienced minor downward displacement (<5 m) with no tilting; and the Mababe Depression has experienced downward displacement of between 5 and 10 m in the south and uplift of about 5 m in the north. They concluded that these movements reflect overall rise of the Ghanzi Ridge. They also suggested that the Thamalakane fault, with its curved strike, which forms an arc symmetrical to the fan, has also been influenced by gravitational loading.

Channel systems of the Delta

Wilson (1973) and Wilson and Dincer (1996) divided the Okavango wetlands into two broad divisions; the upper permanent swamp and the lower seasonal swamp (Figure 21). Both are characterized by a channel network. He divided these channels into three types:

a. Upper Channels are developed in the proximal portions of the permanent swamp and vary between 15 and 130 m wide and 5 to 7 m deep with sandy beds. Banks are dominated by papyrus, especially on the insides of bends, or by *Phragmites* or *Miscanthus*. They are perennial and strongly flowing with velocities of about 0.6 m/s.
b. Middle Channels are also confined to the permanent swamps and are separated from the Upper Channels, which supply them with water by filters, by floating debris blockages, or by restricted take-off channels, usually diverging from the main chanel at a high angle or even flowing backwards. Filters are zones where Middle and Upper channels overlap but without a channel linking them. Water is transferred between the channels by flow through the intervening vegetation. Middle Channels range from about 5 to 20 m in width and their flow velocities are less than about 0.5 m/s. Their beds may be sandy or vegetated.
c. Outlet Channels carry water away from the permanent swamps and are the most variable of the Delta's channels. They are very variable in terms of depth and flow velocity but are generally slow-flowing and seasonal with vegetated beds. They are usually flanked by extensive seasonal floodplains.

Unlike normal rivers, the banks of channels in the permanent swamps are formed by vegetal material and are semi-permeable (McCarthy et al., 1988a), allowing water to escape so that the areas surrounding the channels are permanently inundated. Nevertheless, the channel fringes offer sufficient resistance to flow to sustain a gradient in the water surface, thus permitting the water surface in a channel to rise above that in back-swamps behind the channel fringe.

Channel margins are dynamic in the sense that bank-forming plants tend to grow out into the open channel, but their outward growth is restricted by the flow of water. If flow velocity in a channel declines over time, then the encroachment of bank vegetation decreases the channel width. If, on the other hand, flow velocity rises over time, erosion of the banks occurs, initially by undermining, followed by breaking off of the unsupported vegetation cover. Thus, channels represent

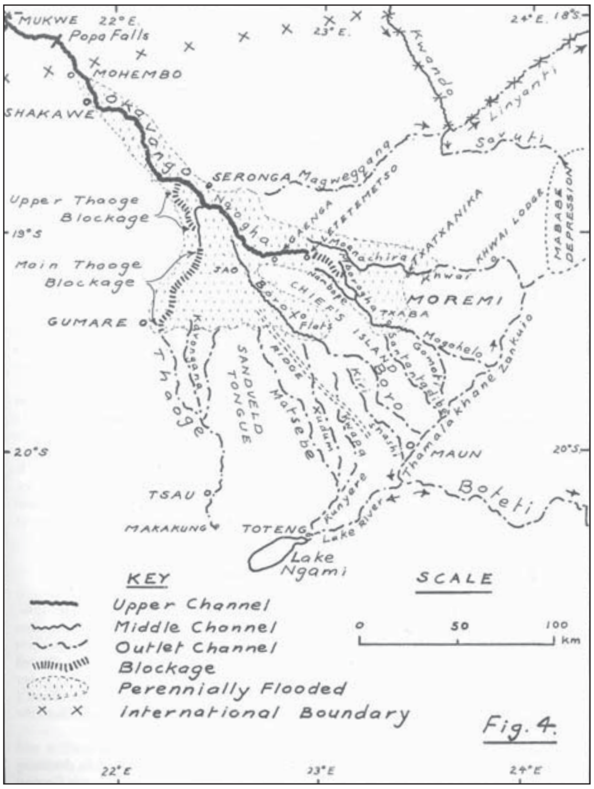

Figure 21. *Map showing the major channel systems of the Okavango Delta (from Wilson and Dincer, 1976).*

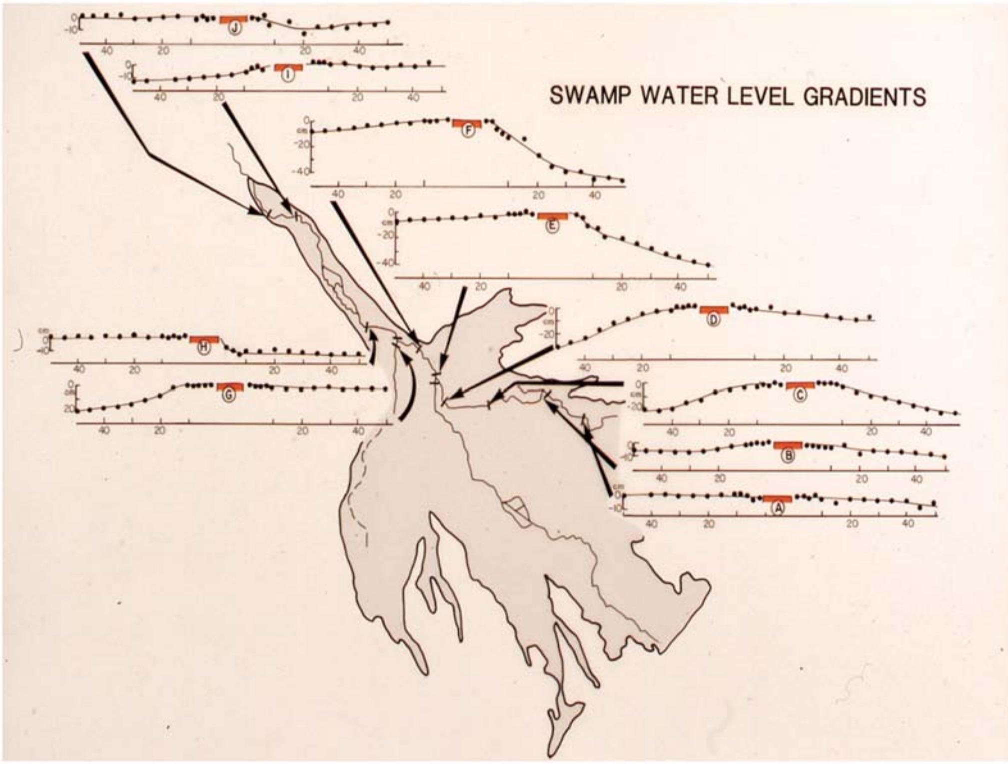

Figure 22. *Water surface gradients lateral to the Okavango, Nqoga and Maunachira channels (McCarthy et al., 1991). The red bar in each section represents the channel.*

a steady-state condition, but this is constantly changing due to changes in flow velocity.

The characteristics of the major channels of the Delta have been measured by W. and K. Ellery and T. McCarthy and their co-workers over several field campaigns (Ellery et al., 1990; McCarthy et al., 1991a; McCarthy et al., 1997; Smith et al., 1997; McCarthy and Ellery, 1997; McCarthy, unpublished data). The results have been summarized by Tooth and McCarthy (2004). The features of the major channels will be briefly described.

The Okavango River.

In the upper portion of the Panhandle reach, the Okavango River is about 90 m wide and about 3 to 5 m deep, decreasing to about 50 m wide and about 4 m deep at the lower end. The channel is actively meandering and highly sinuous, and point bars, scroll bars, cut-offs and oxbows are common features. The floodplain surface is covered by meander-related landforms and in many places the shoulders of the Panhandle are scalloped as a result of erosion by the river. In the middle of this reach, the channel anastomoses and the Filipo channel forms a loop on the eastern bank of the Okavango. Surveys of the water surface elevation along the channel fringes revealed that the water surface often slopes steeply away from the channel (Figure 22; McCarthy et al., 1991a).

Towards the lower end of the Panhandle, the Thaoge and Jao distributaries diverge from the Okavango. It is at the Thaoge take-off that the name of the main channel changes to Nqoga.

The Thaoge channel.

In the mid 1800s, the Thaoge channel was the major continuation of the Okavango channel and extended down the western side of the Delta as far a Lake Ngami (Figure 21; Wilson, 1973; Shaw, 1983). By about 1880, the channel had ceased to discharge into the lake due to papyrus blockages along the channel, and over the following years its reach progressively declined.

The Nqoga channel.

The demise of the Thaoge evidently coincided with the rise of the Nqoga channel (Stigand, 1923). At the time of Stigand's surveys (1910 to 1922) the Nqoga formed a continuous channel with the Okavango, in its lower reaches becoming the Mboroga and then the Santantadibe, which discharged into the Thamalakane channel northeast of Maun (Figure 21). The bulk of the

Thamalakane's discharge at that time was derived from this source. By 1973, the Nqoga channel below Letetemetso Island had become blocked and overgrown (Wilson, 1973) and by 1989, the reach above Letetemetso had also failed (McCarthy et al., 1992; McCarthy and Ellery, 1995a). The Ngoga channel now ends at Hamoga Island a few kilometres upstream. The changes in this section of channel provide important insights into mechanisms of channel failure and water diversion and have been well documented (McCarthy et al., 1992; Ellery et al., 1993a).

At the Thaoge take-off, the Nqoga is 55 m wide and 4 m deep. The Nqoga channel becomes narrower downstream and is 33 m wide and 4 m deep at the take-off channel leading to the Jao (Boro) channel. The Nqoga continues to narrow downstream to about 20 m at Hamoga Island, although depth remains more or less constant. The lateral water level gradients become particularly steep along the Nqoga, rising to more than 40 cm over a horizontal distance of 50 m (Figure 22). Jeffares (1937) found that the water level difference between the Nqoga channel and the back-swamps at Gaenga Island (see Figure 21) was 1.3 m (Wilson and Dincer, 1976).

The Nqoga channel is sinuous (sinuosity about 1.5; Tooth and McCarthy, 2004), but is not actively meandering and the river bends have not changed since aerial photography of the Delta was first undertaken in 1937 (Wilson, 1973; McCarthy, personal observations).

The Maunachira channel.

The Maunachira is classified as a Middle Channel and arises in the permanent swamps close to the termination of the Nqoga channel. The channel is slightly sinuous to straight (Stanistreet et al., 1993; Tooth and McCarthy, 2004). Its source originally lay in a series of shallow open water bodies near Hamoga Island, which were fed by water leaking from the lower Nqoga until a channel was dug across the edge of Hamoga Island in 1973 by Mr. P. Smith (Smith's channel) linking the two. The rush of water rapidly widened the dug channel due to the slope on the water surface. Smith's channel did not persist and required regular clearing to maintain boat passage. It was ultimately abandoned. The upper reaches of the Maunachira are now moribund and no longer navigable by boat. The Maunachira channel becomes narrower downstream. Lateral gradients are modest, reaching a maximum of 20 cm over 50 m horizontal distance (McCarthy and Ellery, 1997). Some of this water flows through a filter region to supply the Mboroga channel to the south (now also linked by a hand-dug channel, which has become wider over the years due to erosion).

The Maunachira flows through several large oxbow lakes. The uppermost oxbow, Dxherega, has experienced substantial change over the past few decades due to sedimentation (McCarthy et al., 1993b) and is now almost completely overgrown. These oxbow lakes are unrelated to the present channel system. In its lower reaches the Maunachira channel anastomoses into broad, open channels which ultimately connect to the lake at Xaxanaka.

The Mboroga channel.

The Mboroga was at one time the continuation of the Nqoga, but the connection was severed by blockages on the Nqoga. It is now supplied with water by leakage from the south bank of the Maunachira and from a poorly defined channel-swamp system, which is supplied by water leaked from the south bank of the Nqoga (Wilson and Dincer, 1976). The Mboroga is a relatively small channel, ranging in width from 10 to 15 m (McCarthy, unpublished data) and its bed is largely covered by submerged vegetation. The Gomoti and Santantidibe once formed important links between the Mboroga and the Thamalakane, but are now blocked and overgrown.

The Boro – Jao system.

The Jao (the name given to the upper reaches of the Boro channel; Figure 21) is connected to the Nqoga by a narrow take-off channel. Much of the flow in the Jao is obtained from lakes situated to the north which, in turn, are fed by water leaked from the Nqoga (Porter and Muzila, 1988). The Jao is narrow (ca. 15 m) and its bed is intermittently vegetated, indicating limited bedload sediment movement. It terminates in an anastomosed network of channels and broad expanses of shallow water known as the Jao and Xo Flats (Wilson and Dincer, 1976; Shaw and Thomas, 1992), which are probably fault-controlled. Most of the discharge in this region probably occurs outside of the channels. Downstream of Xo Flats, the character of the channel (now called the Boro) changes abruptly. For a distance of some 10 km the Boro is deeply incised into the firm, sandy substratum (McCarthy et al., 1997). Downstream the channel becomes less incised and the Boro becomes a sinuous depression flanked by extensive floodplains.

The Xudum system.

The predominantly seasonal Xudum system arises in the Jao and Xo Flats and consists of three channels: the Matsibe, the Xudum and the Xwaapa/Moraphe (Wilson and Dincer, 1976; Figure 21), which flow in a southerly direction west of the Boro. They supply water to the Kunyere channel which flows along the Kunyere fault scarp to Lake Ngami. They are Outflow channels and resemble the lower Boro described previously.

Upper and Middle Channels in the Delta are generally rectangular in form with no evidence of deepening in the thalweg region. Tooth and McCarthy (2004) examined their hydraulic geometry and found that main contributor to variable discharge is width (78%), followed by changes in velocity (22%). They found that channel margins have a significantly greater drag effect than channel beds (Manning's n = 0.11 to 0.15 and 0.02 to 0.03, respectively; James and Makoe,

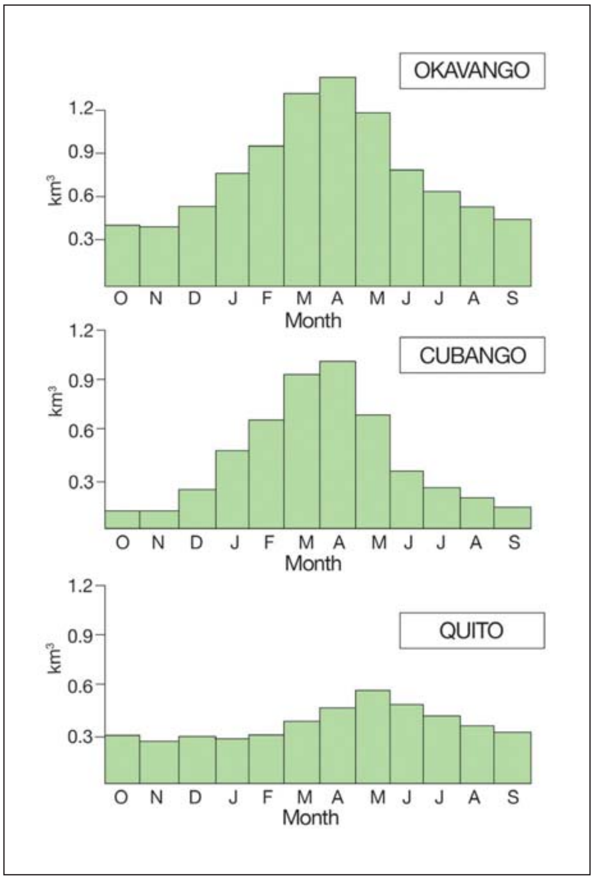

Figure 23. Mean monthly discharge of the Okavango, Quito and Cubango Rivers (Mendelssohn and el Obeid, 2004).

2006). It is surmised that the regular form of the channels and their constant depth must reflect some optimum state.

The Okavango channel in the Panhandle is actively meandering whereas the Nqoga, Maunachira and other channels on the fan are sinuous but are not meandering. However, there are indications that there were actively meandering channels on the fan in the past, which formed the oxbow lakes along the Maunachira. Tooth and McCarthy (2004) found that Okavango channels with a width to depth ratio greater than 10 meandered, whereas those with a lower ratio did not. In narrower channels the margins exert a strong influence on total resistance (James and Makoa, 2006), reducing velocity and hence stream power below the threshold where bank erosion is possible and hence meandering does not occur if the width to depth ratio falls below 10.

Hydrology of the Delta

The Okavango River receives its water from its two main tributaries, the Cubango and the Cuito. The average discharge of the Cubango is 5590 Mm^3/yr and of the Cuito is 4360 Mm^3/yr (McCarthy et al., 2000). Virtually all of the water comes from the upper catchments of these rivers, and the more southerly tributaries seldom flow (Mendelsohn and el Obeid, 2004). The total catchment is large and thus well damped, and sudden, episodic floods do not occur. The Cubango shows stronger seasonality than the Cuito (Figure 23). Discharge in the Cubango peaks in April, whereas in the Cuito the peak occurs in May, although rainfall in both catchments peaks in January to February (Mendelsohn and el Obeid, 2004). The Cuito is flanked by extensive floodplains in its lower reaches which retard the advance of the seasonal flood pulse. The combined discharge is dominated by that of the Cubango and generally peaks in April. Average discharge of the Okavango River measured at Mohembo is 10,100 Mm^3/yr and is quite variable. The recorded annual discharge ranges from a low of 6,000 Mm^3/yr to a high of 16,400 Mm^3/yr.

The seasonal flood pulse passes into the Panhandle region of the Delta, causing a rise in water level of about 1.8 m in the upper to middle reaches (Okavango Delta Management Plan, 2004). Channel margins are permeable and an increasing proportion of water is lost to the floodplains. At the lower end of the Panhandle the seasonal water level range is 40 to 50 cm and declines to about 20 cm along the Nqoga channel. The shoulders of the Panhandle confine the extent of lateral leakage, especially in its upper reaches, but at the lower end of the Panhandle all lateral confinement ceases. Water leaked from the channels simply flows away across the fan, and channels become progressively smaller down the fan. Some of the water gathers and is conveyed in Middle Channels, but the bulk of the seasonal flood flows outside of the channels.

Base flow in the Okavango River is sufficient to sustain permanent inundation in the upper reaches of the fan (permanent swamps). In the period of regular satellite image cover (post 1972) the smallest flooded area recorded (i.e. permanent swamp) was 2450 km^2, which occurred in February 1996 (J. McCarthy et al, 2003). Flood water that leaks from the Upper Channels flows out into the surrounding swamps, causing an expansion in the area of inundation (seasonal swamps) (Figure 24). However, before the swamp area can expand, the groundwater table has to be raised to the surface, and hence flood water flowing onto the seasonal floodplains initially infiltrates into the ground. Most of the flood progresses by unchannelized overland sheet flow. The maximum area of inundation (seasonal plus permanent swamp) during the driest year on record was 5094 km^2 (August, 1996; J. McCarthy et al., 2003) and the largest inundated area recorded was 11,400 km^2 during the flood of 1979 (the latter figure relates only to the period during which satellite imagery has been available). Seasonal water level fluctuations are generally low on the fan, except in the Boro channel downstream of Xo Flats (due to channel incision) and in the Thamalakane (2 m), due to water ponding against the fault scarp.

Progress of the flood wave across the fan is slow, and it takes about four to five months for the flood peak to traverse the 250 km from Mohembo to Maun.

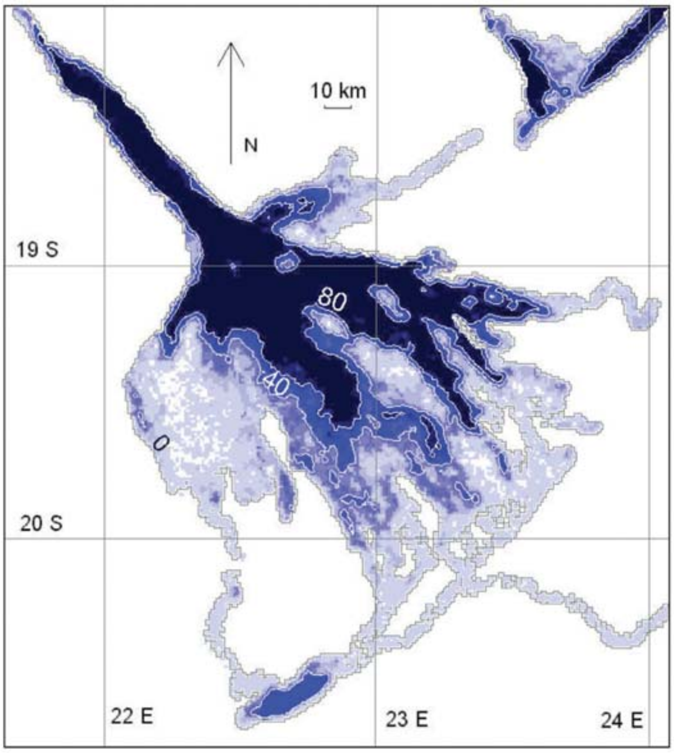

Figure 24. Map showing the percentage annual inundation in the Delta (J. McCarthy et al, 2003).

The maximum area of inundation occurs in July to August, as shown by mean monthly discharge for the Thamalakane which peaks at maximum flood (Figure 25). The flow velocity in channels is such that the flood should take only a few days to traverse the distance between Mohembo and Maun. The reasons for the slow progress are: the slow rate of infiltration of the advancing flood water into the ground to raise the water table; the retarding effect of vegetation; and the very shallow depth of water on the fan (Wolski et al, 2005), averaging probably no more than 20 cm. The largest area of seasonal swamps occurs to the west of Chief's Island (Figure 24). Only about 1 to 2% of the total annual inflow into the Delta leaves as surface flow via the Boteti and the mean annual discharge in the Thamalakane is 236 Mm^3/yr (McCarthy et al., 1998a; Wilson and Dincer, 1976; Gieske, 1997). Very little water appears to leave as groundwater and the bulk of the water is believed to be lost by evapotranspiration.

The extent of inundation of the fan depends on four factors: the quantity of water arriving from the catchment, the amount of rain that fell on the Delta during the summer, the amount of water already in the system, known as the antecedent condition, and the amount of evaporation that takes place over the season. The rainfall and the antecedent condition affect the ground water table in particular. If the water table is high, less flood water is needed to raise it to the surface, and a given quantity of inflow will result in greater inundation (McCarthy et al., 1998a).

Using satellite imagery (NOAA AVHRR) J. McCarthy et al. (2003) measured the area of maximum inundation in most years for the period 1985 to 2000. Gumbricht et al. (2004a) established that these variables are related as follows:

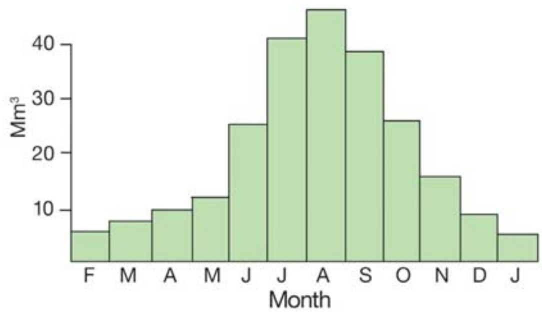

Figure 25. Discharge in the Thamalakane River at Maun.

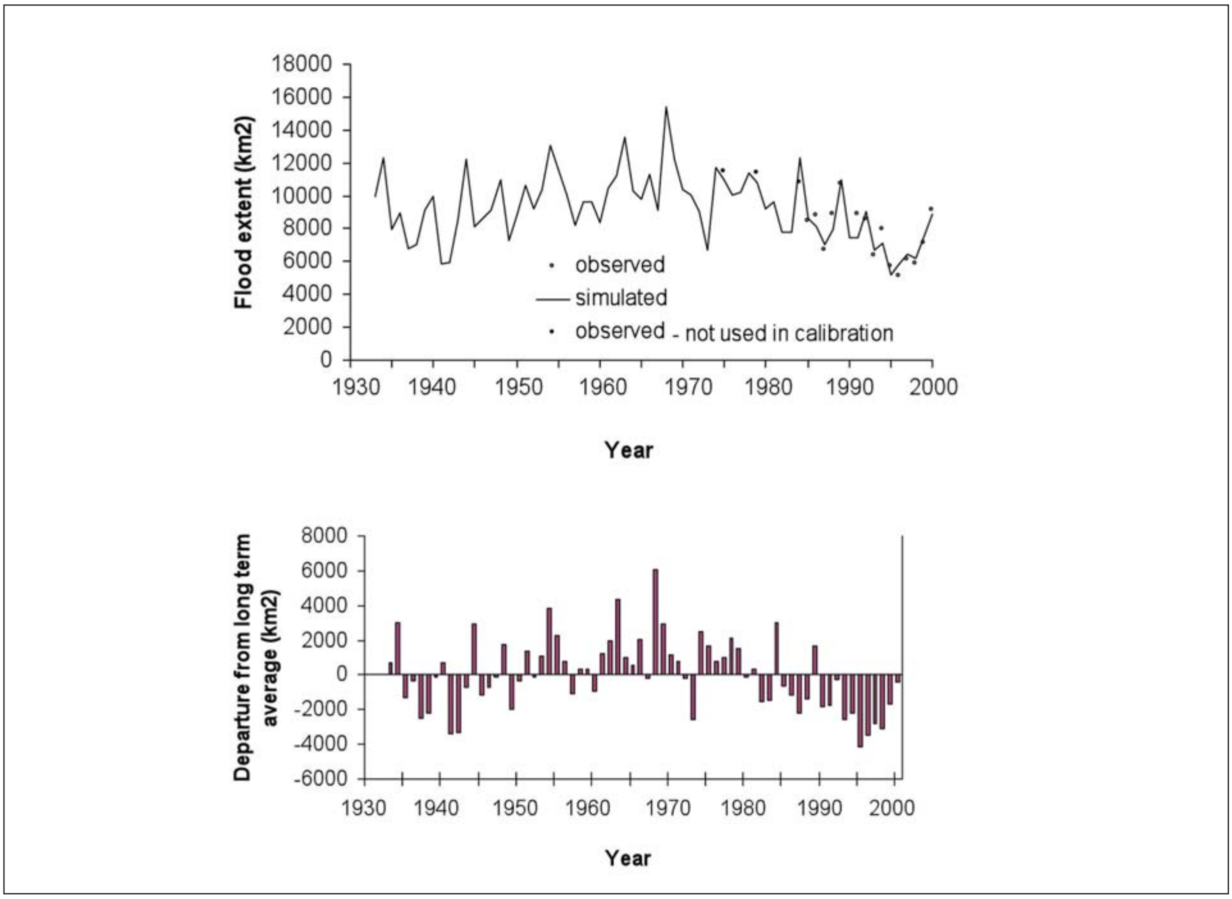

Figure 26. *Calculated maximum area of inundation based on the regression equation of Gumbricht et al. (2004a). The actual inundated area is shown in the upper diagram and the lower diagram shows the maximum inundated area as a deviation from the long term mean. Prior to 1950, the inundated area was below the long term mean, between 1950 and 1980 it was above and after 1980 below, reflecting the quasi 80 year climatic oscillation identified by McCarthy et al., (2000).*

$$A\ (km^2) = 0.79 \times Q\ (Mm^3) + 5.59 \times P\ (mm) + 0.059 \times L\ (km^2) - 817$$

where A is the area flooded, Q is the inflow and L is the area flooded in the previous year.

Inflow and rainfall records for the Delta extend back to 1934 and Gumbricht et al. (2004a) were thus able to calculate the area of maximum inundation back to that time (assuming a value for the flooded area for 1933). Their estimated areas of inundation are shown in Figure 26 as the actual area and as a deviation from the long term average maximum inundated area (9970 km^2). Gumbricht et al. (2004) also devised flood maps showing the area flooded at various flood sizes (Figure 27).

During the course of their investigation, Gumbricht et al. (2004a) also noted that there was a shift in water distribution taking place in the lower Delta, with increasingly more water moving into the Xudum system at the expense of the Boro. Wolski and Murray-Hudson (2006) investigated this change in more detail and found that the Xudum started receiving more water from 1997. The event was abrupt rather than progressive and gradual. After the large flood of 2004 the Xudum was able to make a significant impact on Lake Ngami, which began filling.

The Delta is subject to substantial year-on-year fluctuations in inflow and rainfall, but in addition there are longer term trends, the most important being a quasi 80 year oscillation, with about 40 years of below average rainfall and inflow and 40 years above average (McCarthy et al. 2000). The Delta responds to this cycle by changes in the area of inundation and in the outflow both to Lake Ngami and to the Boteti. The effect of this oscillation on the lower Delta and especially the lower Boro and Thamalakane is quite dramatic. In the decades of the 1960s and 1970s, the Thamalakane River in Maun was perennial but water levels and discharges declined into the 1980s, reaching their lowest in the mid-1990s (Figure 28). During this dry period, flood waters would arrive in the Boro in late July or August, and creep along the dry river bed at walking pace, eventually flowing past Maun. Wetter conditions are currently returning and the Thamalakane River is again perennial. The Savuti Channel appears to be responding to the same climatic oscillation.

Clastic sedimentation and its influence on channel stability

Wilson and Dincer (1976) realized that water leaks from the channels and the bulk of the water discharges across the permanent swamps and floodplains outside of channels, whereas the bedload is confined to the channels. They concluded that channel beds must therefore aggrade over time and went so far as to suggest that the only reason channels exist in the Okavango is to convey sediment.

Suspended load concentration in the upper Okavango and Nqoga channels, amounts to between 0.009 and 0.011 kg/m^3 but appears to vary seasonally. The concentration is much lower in the distal Middle Channels (0.001 to 0.0004 kg/m^3) (McCarthy et al., 1991a). Suspended load is carried into the channel margins by water that leaks from the channels. Here it is filtered out and incorporated into the peaty material that forms the channel banks. The suspended load contributes allochthonous mineral matter consisting mainly of kaolinite (40%) and quartz (20%), and allochthonous and autochthonous silica of plant origin (phytoliths) to the peat (McCarthy et al., 1989, 1990).

Bedload determinations were carried out in the Delta over a number of field campaigns at many different sites (McCarthy et al., 1991a; 1992; McCarthy, 2003). The primary variable influencing bedload movement in Okavango channels was found to be the flow velocity (McCarthy et al., 1991a, 1992), described by the regression equation:

$$Q_b = 0.154\ U^{3.4}$$

where Q_b is the total bedload discharge and U the mean flow velocity.

Using this equation for bedload sediment discharge and the suspended load concentrations referred to above, the estimated total annual bedload entering the Delta at Mohembo is 170 000 tonnes, and total suspended load is 39,000 tonnes (McCarthy and Ellery, 1998).

The bedload consists of very well sorted fine sand with grain size of about 2Φ (0.25 mm) and does not vary systematically down the channel system. Bedload moves as small dunes with ripples on their backs (Coles et al., 2003).

The bulk of the bedload is being deposited in the anastomosed reach of the Panhandle (McCarthy et al., 1991a). Downstream of this reach, bedload flux increases slightly and thereafter declines. The Maunachira channel does not receive bedload from upstream but generates it by erosion of the channel bed (McCarthy et al., 1992). In the more distal reaches of the Maunachira the channel bed is vegetated and there is no bedload movement.

On the basis of these results, McCarthy et al. (1992) formulated a model for the processes leading to failure in the distal sections of Upper Channels, which is illustrated in Figure 29. As an Upper Channel loses water by leakage through its banks, vegetation encroaches and it becomes narrower, decreasing its capacity to convey bedload, which accumulates, raising the bed and encourages more leakage. The leaked water promotes growth of vegetation in the channel fringe, which grows upwards and becomes denser, forming a vegetation levee (McCarthy et al., 1993c). Water surface of the channel thus rises, reflecting this vertical aggradation. This in turn steepens the lateral water surface gradients (Figure 29b), further promoting leakage. The down-channel gradient is also flattened by channel bed aggradation, reducing flow velocity and the channel-flanking vegetation (papyrus in these situations) encroaches into the channel and other plant species take hold, further constricting the channel. Floating papyrus debris now begins to form blockages. The steep water

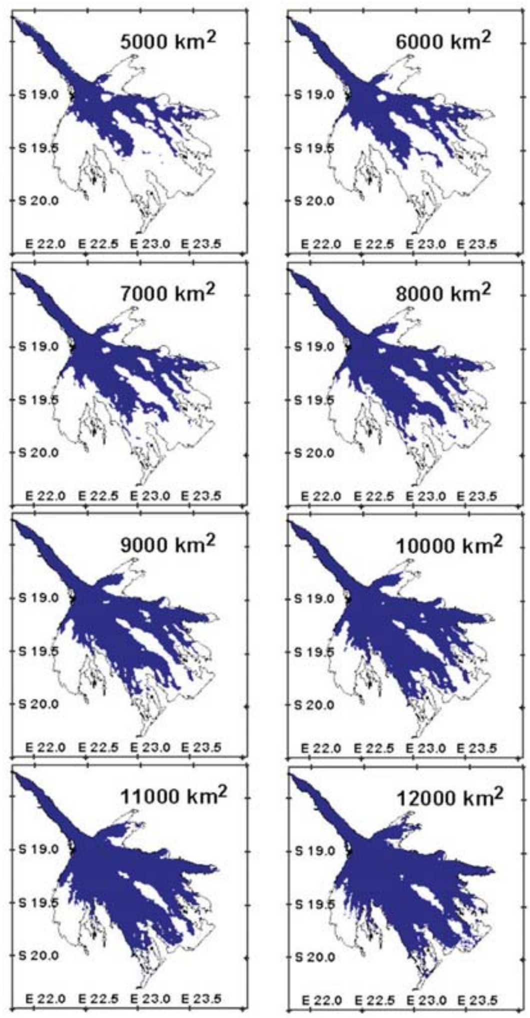

Figure 27. Maps showing the area inundated at various flood sizes ranging from 5000 km^2 to 12000 km^2 (from Gumbricht et al., 2004a).

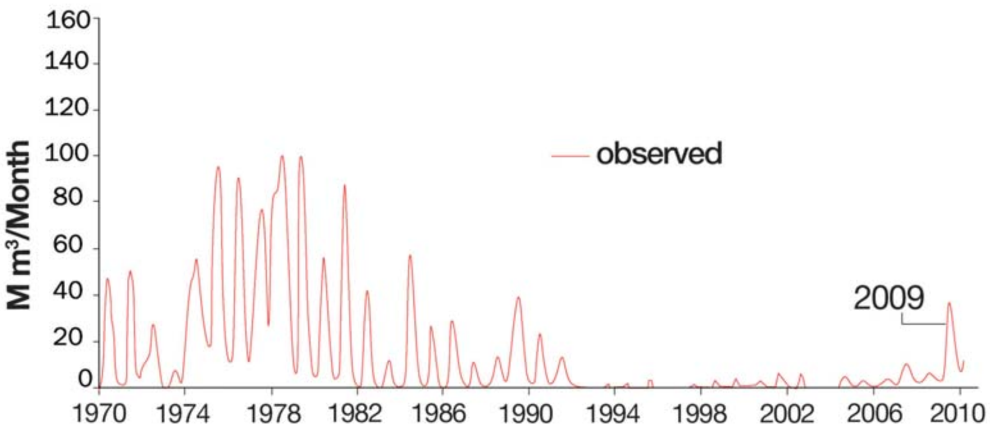

Figure 28. *Discharge record for the Thamalakane River at Maun (P. Wolski, personal communication).*

surface gradient in the flanking swamps promotes flow in suitably orientated pathways nearby, which are provided by hippo trails and a Middle Channel starts to form (Figure 29c) adjacent to the failing Upper Channel, usually nucleating along hippo trails (McCarthy et al., 1998b). It receives no sediment from the failing Upper Channel, and enlarges by erosion of its bed and margins. Meanwhile, the lower reach of the Upper Channel becomes completely blocked and is overgrown. Initially water flows beneath the surface blockages, but this pathway eventually becomes totally blocked and all flow ceases. Deprived of water, the swamp surrounding the failed channel now begins to desiccate (McCarthy et al., 1988c). A proportion of the flow is thus transferred to the flanking Middle Channel but the two do not normally connect directly because of the vigorous growth of papyrus in this setting. The overlapping reach between a failing Upper Channel and a receiving Middle Channel was termed a 'filter' by Wilson (1973). As the Upper Channel continues to fail, so it is followed by headward propagation of the adjacent Middle Channel.

Channel sedimentation is also taking place in the Panhandle and in fact most of the sediment introduced by the Okavango River is being deposited there in an anastomozing reach (McCarthy et al, 1991a). Sediment storage is primarily by bed aggradation, and has resulted in the Okavango channel lying as much as 1 m above the adjacent Filipo channel. As a consequence, flow is being diverted from the Okavango to the Filipo both by direct connections between the two and also by leakage through the channel banks. The Filipo appears to be a composite channel, with some segments representing former oxbows and others representing former minor channels, possibly hippo trails (Smith et al., 1997). Water diverted from the Okavango has linked these various segments to form the now continuous Filipo channel. Aerial photographs of the region taken over several decades indicate that the lower Filipo has been growing at the expense of the Okavango channel and a large section of the Okavango channel has failed. The Filipo will ultimately replace the adjacent reach of the Okavango River.

Peat fires and their environmental implications

Surface fires are a regular and important feature of the Okavango ecosystem (Heinl et al., 2006) and clear out accumulated dead plant material. Fairly regular burning is essential, especially in the seasonal swamps, as it prevents the accumulation of excessive dead plant material. Surface fires are also important in the permanent swamps and assist in nutrient recycling. Fires in the permanent swamps destroy both accumulated dead plant material and those portions of the living plants that are above water, but the rhizomes are unscathed and rapidly shoot to replace the material destroyed by fire.

Peat fires are different from surface fires. As previously discussed, the permanent swamps are characterized by a carpet of emergent vegetation growing on a substrate of peat. Water is supplied to the permanent swamps on the fan by a network of channels, but these are prone to failure. An area formerly supplied with water by a channel begins to desiccate in the dry Delta environment once the channel fails. Drying of the peat occurs from the top down (Ellery et al., 1989). The dry peat catches fire, possibly ignited by surface fires, by lightning, or perhaps by spontaneous combustion. The combustion is very slow, more smoldering than burning, and produces very little smoke. It nevertheless destroys all surface vegetation. The fire heats and chars the moist peat below, causing drying and further combustion. The ash produced is porous and has a very low density.

Combustion proceeds down to a depth where the moisture content and possibly the limited oxygen supply stop further burning. Summer rain may also extinguish the fire (Gumbricht et al., 2002). The ash becomes compacted, facilitated by rainfall and possibly by trampling by animals, and a soil-like crust forms above the unburned peat. The compacted ash layer is colonized by various opportunistic pioneer species. Further desiccation may be delayed by wetter conditions, but eventually will continue and peat burning will recommence. Successive burns will eventually destroy the peat, leaving a layer of compacted ash on the underlying pre-swamp substrate (Figure 30). Detailed studies along the lower Nqoga channel, where peat fires had been burning about a decade earlier, revealed that the original peat layer, which would have been about 4 to 5 m thick, had been reduced to a layer of compacted ash about 15 cm thick overlying 15 cm of charred and unburned peat, representing about a 90% volume reduction (Stanistreet et al., 1993). The former channel bed in this area stands about a metre higher than the compacted ash layer (Figure 31).

Given the nature of the climate in the Okavango, peat fires are an inevitable consequence of channel failure. Peat fires fulfill an important function in the Okavango Delta. The ecosystem is very deficient in nutrients and the soils in the region are sandy and of low inherent fertility. Peat accumulates fine particulates including clay minerals carried into the Delta from the catchment as suspended sediment. Bacterial activity and ion exchange also contribute to the accumulation of nutrients by the peat. Peat ash becomes mixed into the generally sandy substrate material by bioturbation by insects and animals, resulting in soils of superior quality than the normal soils of the Kalahari region in terms of texture and nutrient status.

Groundwater and its importance in the structure and function of the ecosystem

The depth to the water table in and around the Delta was compiled from boreholes and dug wells by Aquatec (1982) and is shown in Figure 32. The data were collected over several years and probably reflect the average condition at the end of the relatively wet decades of the 1960s and 1970s, possibly not very different from the present situation. The water table is shallow beneath the Delta but deepens to the west and south. The gradient in the west is particularly steep. Overall, the form of the piezometric surface (water table) resembles a recharge mound centred on the Delta.

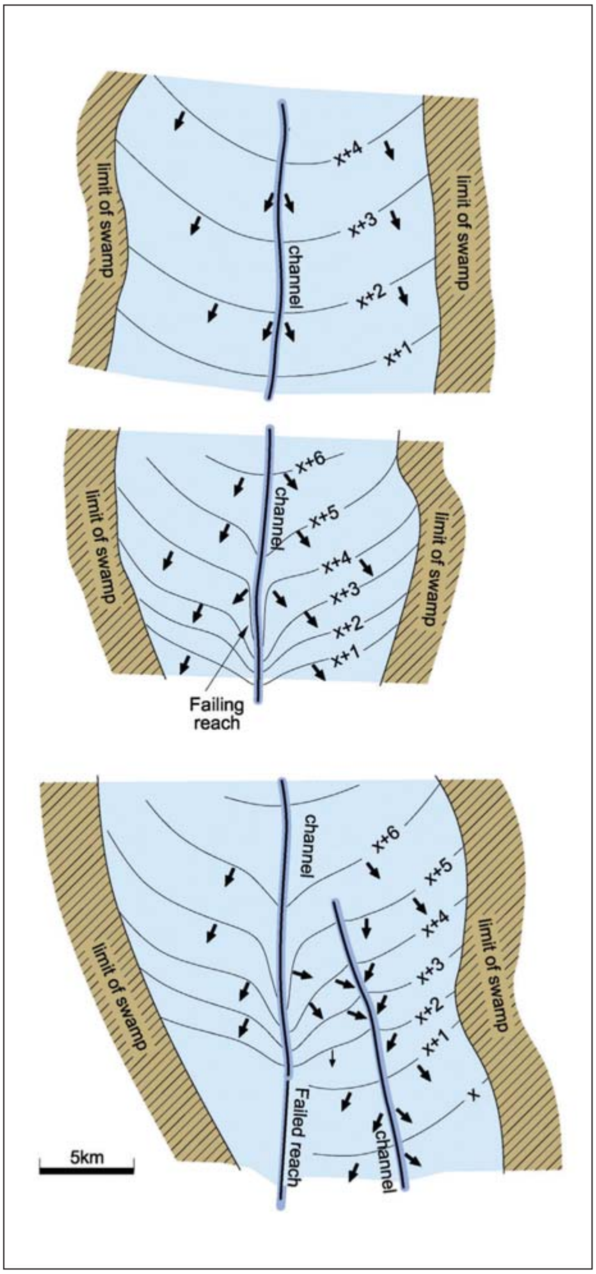

Figure 29. *Schematic diagrams illustrating the manner in which bedload aggradation leads to local steepening of the water surface gradient around a channel, the formation of a secondary channel (Middle Channel) and channel failure (McCarthy et al., 1992).*

McCarthy et al. (1998a) found that there are three distinct compositional types of water. Groundwater within the Delta is bicarbonate dominated and low in sulphate and chloride and resembles the surface water although the salinity is very variable. Electrical soundings and drilling carried out around the fringes of the Delta and especially in the Maun area indicate a marked increase in salinity with depth and it appears that fresh water is floating on underlying saline water, especially beneath the distal floodplains (Aquatec, 1982; Bureau de Recherches Geologique et Miniers, 1986;

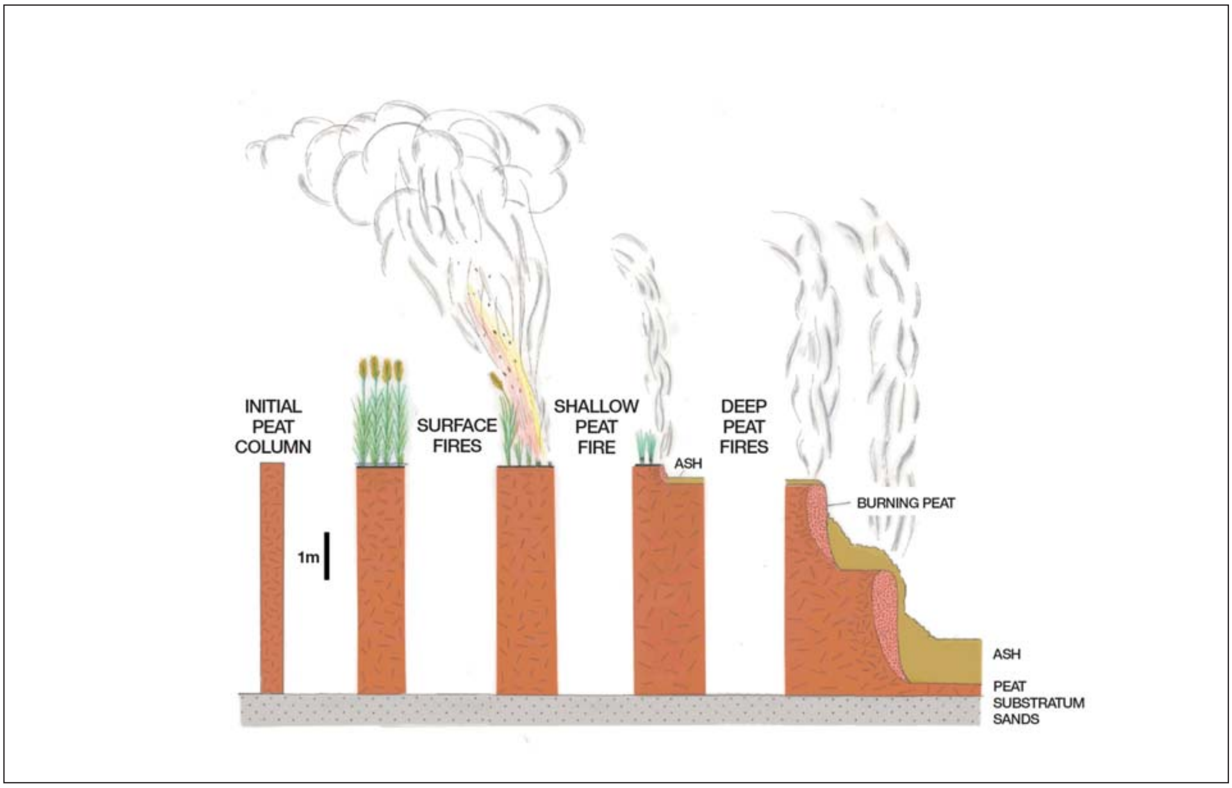

Figure 30. *Peat burning usually proceeds in a number of passes as the material dries progressively from the top.*

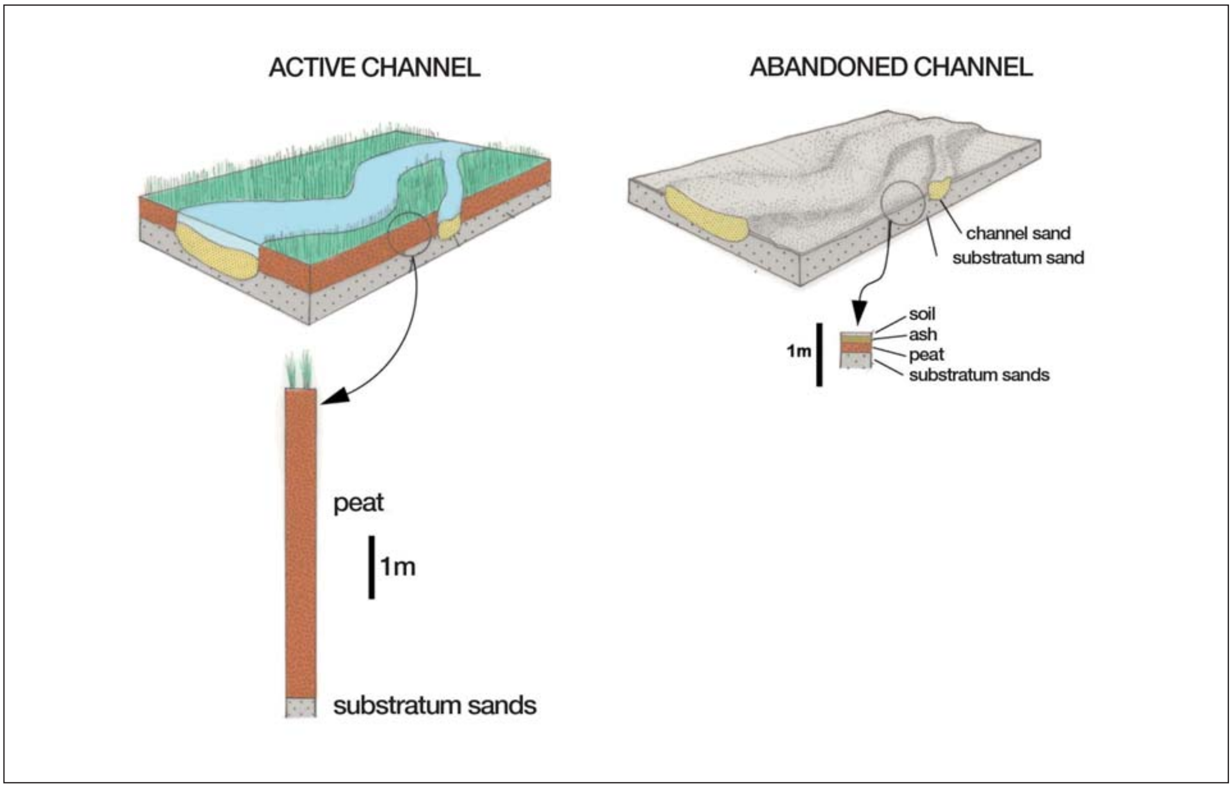

Figure 31. *Peat burning causes a volume reduction of about 90% which results in a substantial lowering of the terrain formerly occupied by swamp. Former channels in which bedload had accumulated often stand in positive relief as a result.*

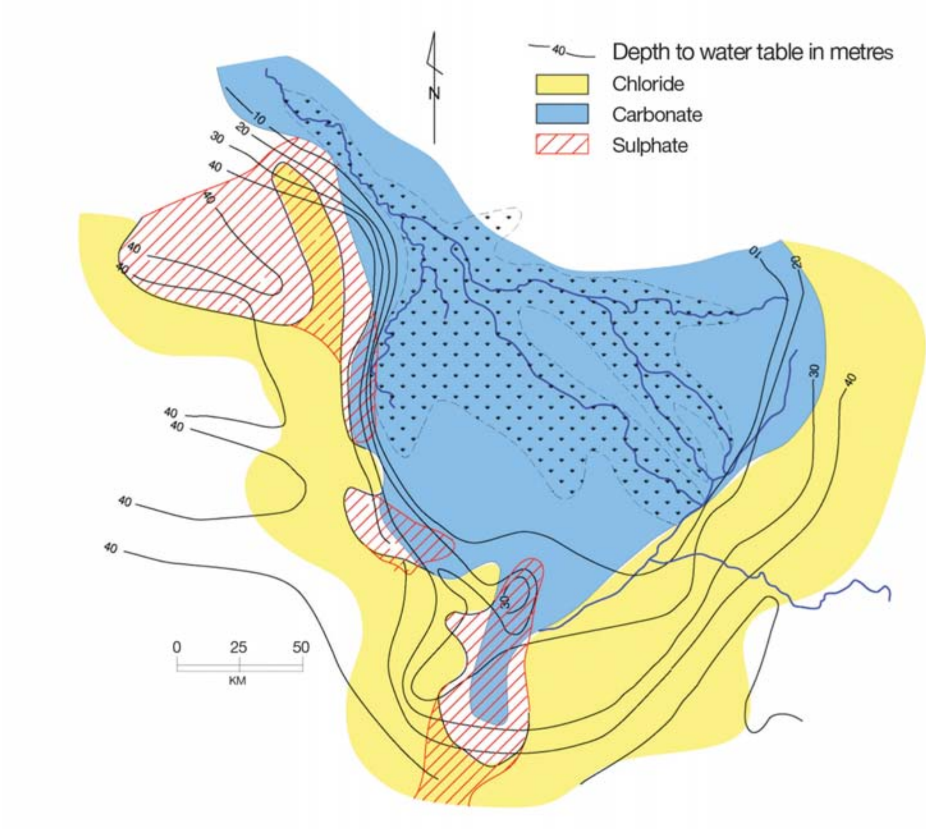

Figure 32. *Map showing the depth to the water table and the chemical character of the groundwater in and around the Delta (Aquatec, 1982; McCarthy et al., 1998a).*

Maun Groundwater Development Project, 1997; Campbell et al., 2006). This situation may exist across the entire Delta (Bauer, 2004).

Salinity of the groundwater surrounding the Delta is generally high (3000 to 30,000 mg/l) and the water is either chloride- or sulphate-dominated (Figure 32). It is not related to Okavango Delta water. In spite of the resemblance to a recharge mound, water chemistry and isotopic abundances suggest that there is no widespread groundwater outflow from the Delta, although the possibility that groundwater is discharging along discrete fault zones cannot be excluded (McCarthy et al., 1998a).

The annual flood wave flows through the permanent swamps and extends the inundated area mainly by sheet flooding. Water discharging onto dry floodplains rapidly infiltrates into the sandy soils, raising the water table beneath both the floodplains and its islands. Two studies examining the infiltration process have been carried out, one at Beacon Island on the floodplains west of Chief's Island (Dincer et al., 1976; see McCarthy, 2006, for a synopsis) and the other at Nxaragha on the edge of Chief's Island (Ramberg et al., 2006). These studies revealed that during the initial stages of the flood infiltration was rapid, but this tapered off as the flood season advanced. At Beacon Island 80% of the water discharging onto the floodplain was lost by infiltration and 20% by evaporation. At the Nxaragha site, 90% was infiltrated and 10% evaporated. It thus appears that as the flood advances infiltration is fast initially, raising the water table and this consumes much of the flood. The amount of infiltration undoubtedly varies spatially: on the fringes of the permanent swamp, the water table is high so the quantity of water infiltrated is small. However, as the flood advances further from this fringe into the seasonal swamps, the depth to the water table increases and more of the flood is lost to infiltration. Antecedent conditions also play a role: a large flood in the previous season or very heavy summer rains will result in a high water table and hence less infiltration will occur. The effect of a single season of heavy rainfall may actually persist for several years (Milzow, 2008). Infiltration near extensive land masses such as Chief's Island is particularly large and does not diminish as much as on floodplains in the waning flood, because of the large groundwater reservoirs provided by these land masses.

Figure 33. *Island vegetation shows a distinct zonation: the outer fringes are characterized by evergreen trees, followed inwards by deciduous trees, ivory palms, shrubs and grass and finally barren soil. The swamp surrounding the island is densely vegetated with sedges and grasses.*

Islands offer important insights into the interactions between the terrestrial and the aquatic realms of the Delta and have been subject to many detailed investigations.

Islands in the permanent swamp.

In the permanent swamps on the upper fan, seasonal water level variations are small. Islands generally rise little more than a metre above the swamp water level. They show a characteristic vegetation zonation (Figure 33): the outer fringes support evergreen trees (riverine forest species), which pass inwards to a deciduous tree zone with the distinctive ivory palm on the inner fringe, followed by a zone dominated by grasses and shrubs that finally gives way to a barren interior where the soil surface is usually coated with efflorescent carbonate salts (trona and thermonatrite) (Ellery et al., 1993b; McCarthy et al, 1986a). The water table is generally lower than the water level in the surrounding swamp, sometimes by as much as 50 cm. The steepest slope on the water table generally occurs beneath the island fringe, whereas the water table beneath the island interior is generally flat (Figure 34). Electrical conductivity (salinity) of the groundwater rises sharply towards the island interior where conductivity is typically three orders of magnitude higher than in the surrounding swamp.

Both the lowering of the water table and increase in salinity of the groundwater beneath islands is the result of the loss of water to the atmosphere. Some of this loss occurs via evaporation from the capillary zone, which generally extends to surface in the island centres, but it is mostly due to transpiration by the trees, especially those in the riverine forest fringe. They draw down the water table, forming a cone of depression beneath the island. Equilibrium is established between groundwater recharge and the transpiration rate, which determines the extent of the draw-down of the water table. Transpiration rate declines during the night, and the water table rises accordingly as the equilibrium shifts, resulting in a diurnal fluctuation in the depth to the water table. McCarthy and Ellery (1994) measured diurnal fluctuations in the water table of 7 cm under the forest fringe of an island on the edge of the permanent swamp while that in the island centre remained static. Longer term measurements on several

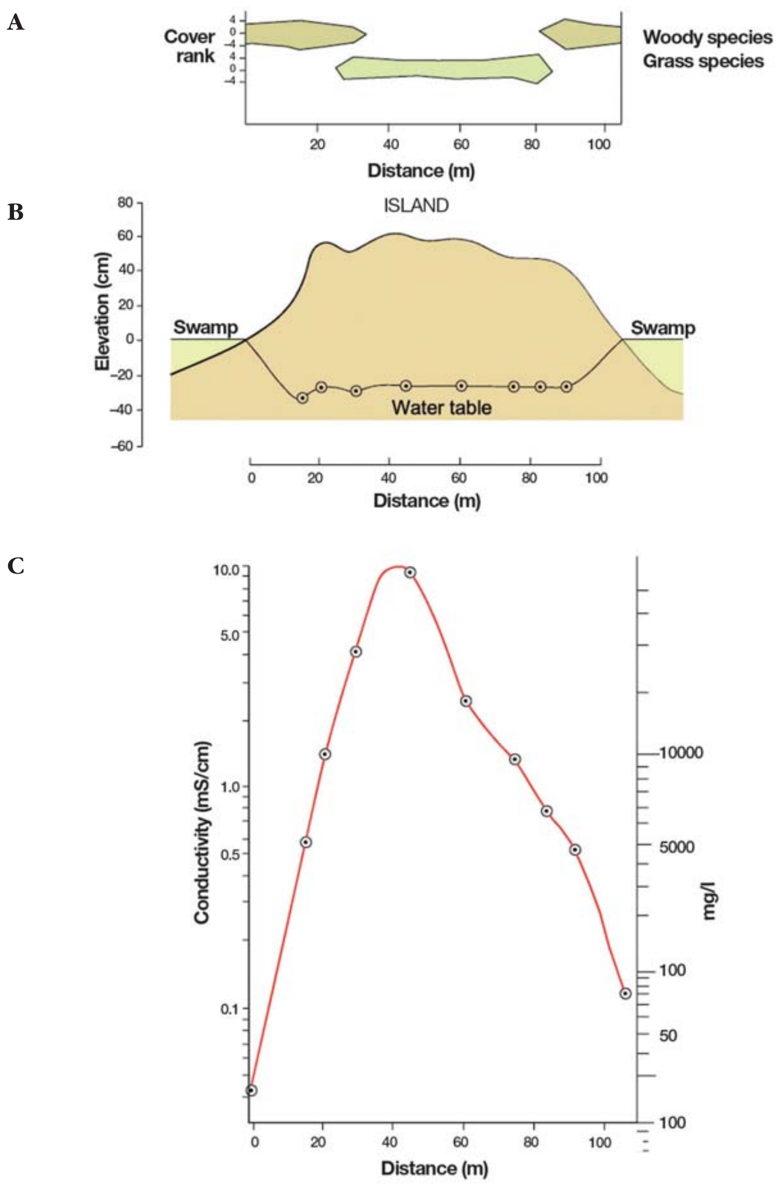

Figure 34. Diagrams showing the distribution of vegetation (A), topography and depth to the water table (B) and groundwater conductivity (C) across an island in the permanent swamps (McCarthy et al., 1993d).

islands by Bauer et al. (2006) recorded diurnal ranges of between 3.4 and 7.5 cm.

Islands on the fringe of the permanent swamp.
These islands show pronounced variations in groundwater level, although the floodplains retain water throughout the year (McCarthy and Ellery, 1994; Ramberg and Wolski, 2008). At all times the water table beneath the island remains lower than the surrounding swamp level, but the extent of the differential varies during the year, reaching a maximum just prior to the arrival of the seasonal flood. Although the water table fluctuates in depth, the saline zone remaines anchored beneath the island centre.

Islands in the distal seasonal swamps.
In the distal seasonal swamps the floodplain dries up completely during the summer. In the dry season the water table becomes horizontal but a depression develops beneath the island once the flood water arrives

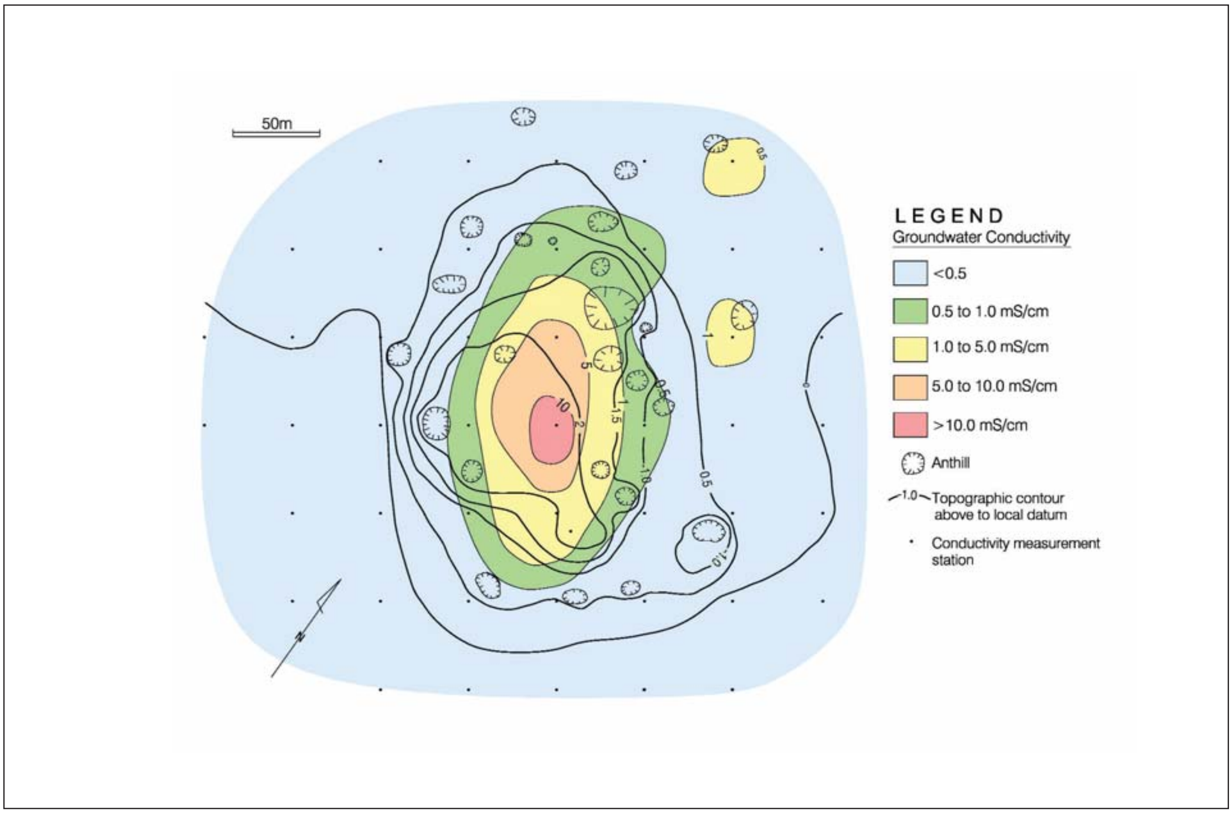

Figure 35. *Diagram showing the relationship between the topography of an island and the conductivity (salinity) of the underlying ground water (McCarthy, 2006).*

(McCarthy and Ellery, 1995; Wolski and Savenije, 2006; McCarthy, 2006). Typically salinity varies by three orders of magnitude between floodplain and island centres but the most saline water remains centred beneath the middle of the islands throughout the seasonal cycle (Figure 35), indicating the absence of lateral flow.

The water table is continuously drawn down beneath islands partly by evaporation off the capillary zone but mainly by transpiration, especially by trees, leading to diurnal fluctuations in the water table. Bauer et al. (2006) used these diurnal fluctuations to estimate the evapotranspiration rate, and found that the figure varies from 0.06 mm/day in the barren centre to 4.3 mm/day under the vegetated fringe. Based on modeling of the groundwater - surface water interaction beneath an island in the seasonal swamps Wolski and Savenije (2006) estimated the evapotranspiration rate from an island, which equates to about 3 mm/day for an entire island. The maximum transpiration rate of the islands is thus in the region of about 1600 mm/yr, compared to the evaporation rate for the area of 2172 mm/yr and average rainfall of 490 mm/yr.

Studies of the infiltration of the advancing flood in the middle and lower seasonal swamps have revealed that 80 to 90% is lost to infiltration and only 10 to 20% evaporates. The dissolved solid content of water leaving the Delta is about double that flowing in (Cronberg et al., 1995), meaning that the maximum evaporative loss by surface water that traverses the entire Delta is 50%. The average evaporative loss for the seasonal swamps lies between these figures, and given that only 2% of inflow leaves as surface outflow, it is likely that the evaporative loss from the seasonal swamps is closer to the 10 to 20% range. The remainder of the flood water infiltrates the ground and is lost mainly by transpiration, especially by trees as there is very little or no groundwater outflow.

Based on a hydrological model that incorporated

Table 1. Water balance for the Okavango Delta based on a hydrological model (Wolski et al., 2006)

		Mm^3/yr	mm/yr	%
Inputs	Inflow	8780		58
	Rainfall	6260	467	42
	-Floodplains (9327 km^2)	4370	469	29
	-Islands (4090 km)	1890	463	13
Total inputs		15 040		100
Outputs	Outflow	260		2
	Evapotranspiration	14 920	1111	98
	-water surface	8720	1496	57
	-dry floodplains	2540	773	17
	-islands	3660	894	24
Total outputs		15 170		
Change in storage		-130	-9	

measured inflow, rainfall and knowledge of the spatial distribution of flooding, Wolski et al. (2006) derived estimates of the evapotranspirative loss from the water surface, the dry floodplains and the islands (Table 1). They did not, however, attempt to separate transpiration and evaporation. They estimate that 57% of the water is lost from the water surface, representing a mixture of evaporation and transpiration. McCarthy and Ellery (1995b) speculated that emergent plants on the floodplains (and in the permanent swamps) would suppress evaporation by shading and by reducing air flow over the water surface but at the same time would actively transpire water into the atmosphere. They suggested that transpiration from floodplains exceeds evaporation. The evapotranspirational loss from dry floodplains and islands, amounting to 41% of the total, is almost entirely due to transpiration. The bulk of the total evapotranspirative water loss therefore appears to be via transpiration.

The distinction between transpiration and evaporation is extremely important because evaporation results in an increase in the dissolved solid content of open water and ultimately leads to the development of surface saline brines. Such brines are rare in the Okavango, even though most of the water is lost to the atmosphere. In contrast, transpiration involves the uptake of water by plant roots. Plants selectively exclude dissolved solids when taking up water so that the dissolved solid concentration in the root zone rises. In the case of the permanent swamps, this occurs in the peat, but the pore water is continually being flushed by through-flow, so metals do not accumulate to any significant extent. Plants growing on flood plains induce an increase in the dissolved solid content of pore water in the root zone, but this has not been systematically studied. Pore water in the root zone on the floodplains is also continually being replenished by through-flow as fresh water infiltrates from the surface, so significant increase in dissolved solids will probably only occur once the surface water is no longer present.

The situation is different beneath islands where the high transpiration rate lowers the water table, resulting in centripetal flow of groundwater. Islands thus represent groundwater sinks where solutes accumulate. Increasing solute concentration results in successive saturation and precipitation of silica, followed by low magnesium calcite grading to high magnesium calcite followed by potassium feldspar formation (McCarthy et al., 1991b; McCarthy et al., 1993d; McCarthy and Ellery, 1994; McCarthy and Ellery, 1995b). This process has the effect of removing dissolved silica, calcium, magnesium and potassium from the water, leaving only sodium bicarbonate remaining in solution. McCarthy et al. (1991b) found that the increase in salinity was accompanied by an increase in density, and they suggested that this may result in density-driven subsidence of the saline groundwater. It was also found that the silica and calcium carbonate precipitate mainly beneath the forested island fringe and that there is a broad correspondence between topography and calcium carbonate content of the underlying soil (Figure 36). This suggests that the precipitation of calcite (and probably silica) causes vertical growth of the islands. The highest salinity generally occurs in the centre of the islands.

The increase in salinity from margin to centre is accompanied by changes in vegetation (Figure 33). Ellery et al. (1993b) found that this zonation reflects the degree of tolerance of different plant species to salinity of the groundwater in the case of the tree species and of the soils in the case of the shrubs and grasses.

The efflorescent carbonate salts in the barren zone form as groundwater brought to surface by capillarity evaporates, precipitating its solutes. It is likely that these are redissolved by rain and are flushed down into the soil, only to be drawn back to surface as the water evaporates. The solutes thus steadily accumulate in the shallow groundwater. The initial increase in concentration of salts in the groundwater is thus caused by plant action, but the final, highly saline water in the interior is produced by repeated cycles of precipitation and dissolution.

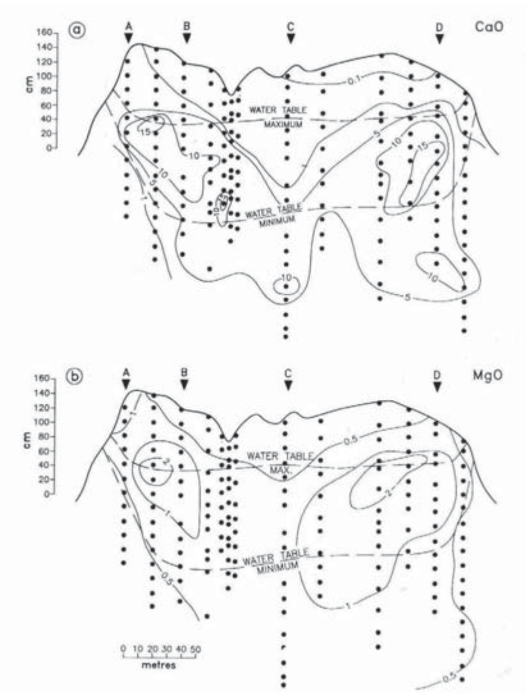

Figure 36. *Cross sections of an island showing the abundances of CaO and MgO in the soils (McCarthy and Ellery, 1994), which reflect the distribution of magnesian calcite.*

The high salinity sets up density-driven advection (McCarthy et al., 1991b; Gieske, 1996), which has been studied in detail using geophysical imaging of groundwater conductivity and numerical modeling by Bauer (2004) and Bauer-Gottwein et al. (2006). Numerical modeling suggested that density driven flow should commence at a dissolved solid concentration above about 17 g/L and that once started, the salinity of the groundwater would stabilize at about this value. Bauer calculated that this would happen after about 200 years. Analyses of groundwater from many islands by McCarthy and co-workers have shown that the maximum concentration of sodium attained is generally in the range 4000 to 5000 mg/L, which corresponds to 14 to 18 g/L sodium bicarbonate, consistent with the predictions made by Bauer and suggest that such density driven flow might be a fairly common phenomenon in the Okavango. Bauer (2004), however, concluded that such density-driven flow could not account for the salt present in the deep aquifer below the Delta. Further investigation of this topic is needed.

Microtopography

The surface of the Okavango is not perfectly smooth but contains undulations that rise above water level to form islands. Because of the shallow average depth of water on the fan surface, tracts of land elevated by as little as a metre can form islands. These islands are an extremely important component of the ecosystem. Islands range in size from about 1 m^2 to many km^2 and originate in two ways: by fluvial processes and by biotic processes.

Fluvial processes.

There are two types of fluvial process that result in locally elevated topography. The first is by deposition of bedload on point bars. The height of the crest of the point bar is determined by the water level. Periods of sustained high water levels lead to more elevated point bars, which stand out as curved ridges known as scroll bars (Figure 37). The second arises from accumulation of sediment on channel beds due to water leakage through the permeable margins of channels. Channels aggrade and ultimately fail and become overgrown. This process leads to desiccation of the peat surrounding and encasing the former channel. The peat then burns away and the bed of the former channel is left as a slightly elevated ridge (Figure 31). The process occurs on scales ranging from former hippo trails to major channels over 200 m wide (Figure 37). When the area is reflooded, the former channel bed stands as an elongated island. Transpiration by the trees growing on islands initiates precipitation of calcite and silica in the soil and the island expands. Some of the larger inverted channel islands are visible on the SRTM DEM of the Okavango (Figure 19).

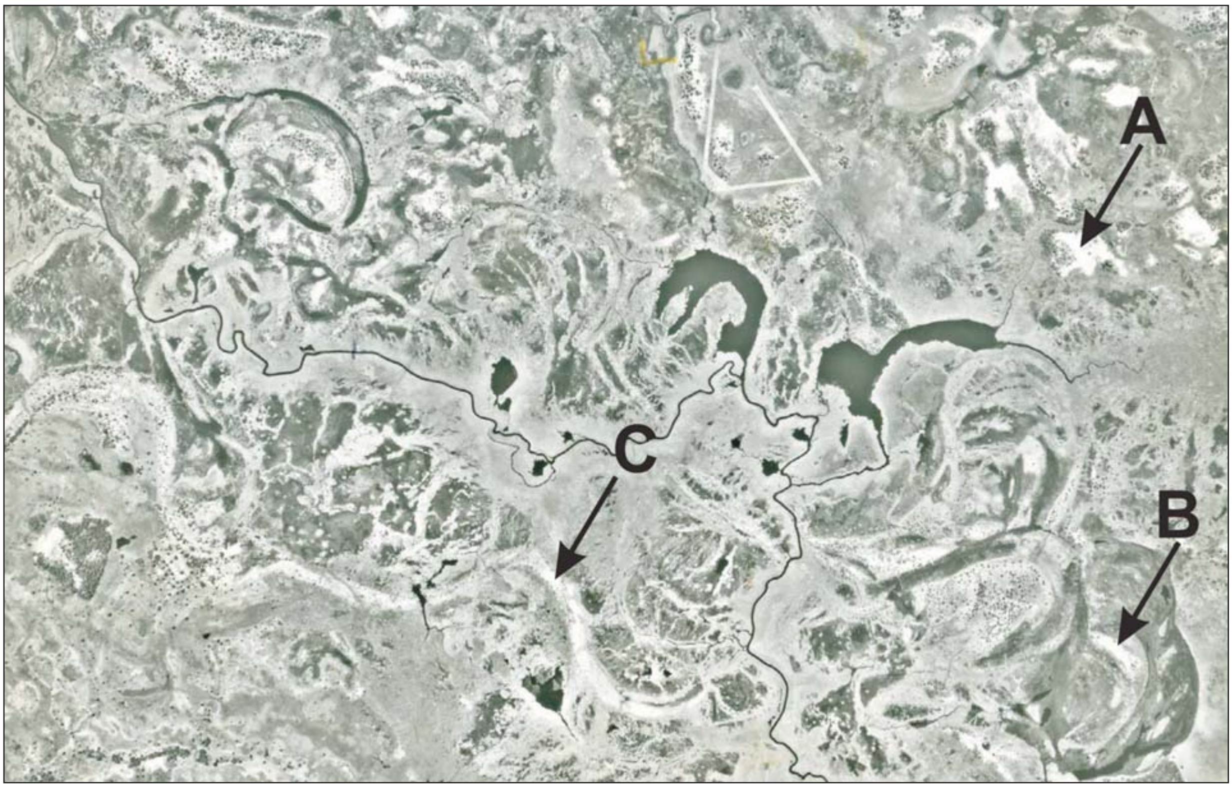

Figure 37. *Aerial photograph of a portion of the Xugana area in the northwestern Delta showing the diversity of island types: (**A**) amoeboid; (**B**) scroll bar; and (**C**) inverted channel. The field of view is about 9 km wide.*

Table 2. Terrain classification of the Okavango (Gumbricht et al., 2004b)

Ecoregion	Panhandle	Permanent	Seasonal	Occasional
Total area (km^2)	817	2507	3287	7082
Forest and grassland (%)	5	3	17	55
Flooded grassland (shrubs and non-aquatic grass species) (%)	12	2	15	37
Floodplain (sedges and aquatic grass species) (%)	37	13	42	8
Permanently inundated (%)	41	75	6	0
Open water (%)	5	7	1	0

Biotic processes.

Topographic relief is also created by the termite species *Macrotermes michaelseni* (McCarthy et al, 1998c; Ellery et al., 1998; Dangerfield et al., 1998). A termitarium consists of a spire-like central turret containing large air passages, which is surrounded by a broad pediment made of material eroded from the turret. The height of the turret can reach 5 m or more. In order to build the structure the termites import material from an extensive tunnel network beneath the surrounding area. When a colony dies the turret is no longer maintained and becomes eroded, merging with the pediment, which survives as a mound a metre or more high. The mound not only differs from the surrounding area in terms of elevation, but in addition contains more fine particulates and thus retains moisture for longer than the surrounding, sandier soils. The mound is also enriched in nutrients because termites import plant material. The mound is thus a local "hot-spot" and is colonized by shrubs and trees, which attract birds and animals whose faeces further raises the nutrient status of the soil.

When the surrounding area becomes flooded, precipitation of calcite and silica from the groundwater occurs beneath the mound, causing it to grow. Therefore, once a mound has formed it sets in motion a cascade effect, which causes it to grow ever larger and become more nutrient-rich (Dangerfield et al., 1998; McCarthy et al., 1998c). Not only do the soils become enriched in nutrients, but the groundwater beneath the island also becomes nutrient enriched (Ramberg and Wolski, 2008; Wolski et al., 2005). Precipitation of calcite and silica occurs preferentially around the fringes of the islands where evergreen forest grows and as a result the outer edge may, over time, become slightly elevated relative to the island interiors, so that these islands often have an atoll-like appearance. The islands grow laterally over time and adjacent islands probably coalesce, forming larger platforms of irregular ("amoeboid") shape (Figure 37). As they grow they may also merge with fluvially formed islands.

Both types of island support trees and during the dry season, wind raises dust clouds across the barren floodplain surfaces. Most dust movement occurs on the floodplains and least in the barren areas in the interior of islands (Krah et al., 2004). Dust particles falling on the floodplains are re-entrained many times, whereas the tree fringe surrounding islands acts as a dust trap. Island interiors are sheltered by the forest fringe and little dust movement takes place there. Dust accumulation thus probably contributes to island growth, particularly around the edges.

Gumbricht et al. (2004b) used image analysis techniques and Landsat imagery (28.5 m x 28.5 m pixels) to delineate and classify the Okavango terrain. At the largest spatial level, they divided the Delta into four domains: Panhandle (817 km^2), Permanent swamp (2507 km^2); Seasonal swamp (3287 km^2); and Occasional swamp (7082 km^2). Each of these domains was further subdivided as shown in Table 2. The total active area of the Delta amounted to about 13,500 km^2, about 9000 km^2 of which has been flooded at one time or another during the past 30 years. In contrast, the area of the Okavango fan as a whole is about 40,000 km^2.

Gumbricht et al. (2004b) classified islands into the following three primary genetic classes: scroll bar islands, inverted channel islands and amoeboid islands. The amoeboid islands were those with irregular to circular shapes and these were further subdivided into several sub-classes based on the type of land cover. Forested islands less than four pixels in area were separately delineated as 'termite mound' islands.

Table 3. Distribution of island types (Gumbricht et al., 2004b)

	Panhandle	Permanent	Seasonal	Occasional
Average island size (km^2)	4.6	2.8	8.0	14
Number of large islands	3139	4198	11043	24733
Scroll bar islands (%)	3	2	0	0
Inverted channel islands (%)	28	25	30	25
Amoeboid islands (%)	69	73	70	75
Number of termite mound islands	13685	16517	29009	43942
Total number of islands	16842	20715	40052	68675
Number of islands/ km^2	20.6	8.3	12.1	9.7
Islands showing salt crusts		966 (22%)	1766 (16%)	3462 (14%)

Island size distribution follows a power law (Pareto distribution). The lack of any size clumps amongst islands indicates that the island-forming processes are scale invariant and islands of all sizes have a proportional growth rate.

Gumbricht and McCarthy (2003) investigated the orientation of elongated islands longer than 50 m. Most are preferentially orientated radially on the Delta, but as size becomes smaller, greater dispersion of orientations occurs. Only about 30% of islands are of fluvial origin, and the majority of the rest arise from termite activity, yet these too tend to be preferentially orientated radial to the fan. The tendency is weak amongst smaller islands but becomes stronger as the islands elongate. Gumbricht and McCarthy (2003) suggest that this is a consequence of feedback between islands and surface water flow. As islands enlarge they impede flow and create a hydraulic head across the island. The effect of this is to displace the salinized portion to the downstream end and to focus calcite and silica precipitation on the upstream end, thus causing preferential growth in the azimuth direction.

Gumbricht et al. (2001) showed that the Delta surface is remarkably smooth, approaching a near perfect conical shape, with small deviations generally less than 2 m high, which give rise to the micro-topography (Gumbricht et al., 2005). The average water depth in the Okavango swamps is less than about 1.5 m (possibly about 0.2 m on average), so the elevated ground forms the islands. Some of the islands are of fluvial origin but most islands are amoeboid in form and probably originated as anthills, but have become enlarged by calcium carbonate and silica accumulation. From knowledge of the total volume of islands and an estimate of their average calcite content, Gumbricht et al. (2005) estimated that the present surface of the active Delta contains about 6 x 10^9 tonnes of calcite. Since 1.15 x 10^5 tonnes is introduced annually, they inferred that the presently active Delta represents accumulation over a period of about 50,000 years and that the surface of the entire Okavango fan could have formed in about 150,000 years. Ramberg and Wolski (2008) suggested this may be an underestimate based on a revised estimate of the quantity of accumulated carbonate.

In the upper part of the fan, sedimentation is primarily clastic in nature, whereas in the lower seasonal swamps, calcite and silica precipitation is the dominant form of sedimentation. Chemical and clastic sedimentation appear to be in balance, and the uniform slope of the fan is maintained. The amount of vertical aggradation is restricted by water depth to less than 2 m, producing limited local relief on the fan, and in turn, the limited local relief maximizes the area of inundation. Accumulation of sediment in one section of the fan will cause local cumulative aggradation and the water will shift elsewhere on the fan. In this way, the microtopograhy is maintained and water distribution and hence sedimentation is dispersed uniformly over the fan, maintaining it's near perfect form. Mass balance estimates suggest that over a period of 150,000 yr all parts of the fan surface (about 40,000 km^2) will experience aggradation.

Lakes of the Okavango Delta

Lakes are a minor but nevertheless important feature of the Okavango Delta. The largest is Lake Ngami, which occupies a topographic low area on the southwestern flank of the conical Okavango alluvial fan, as discussed earlier. The vast majority of the remaining lakes are situated in the permanent swamps, especially at the head of the fan and lower Panhandle and along the Maunachira channel system. These lakes are termed *lediba* (pl. *madiba*) in Setswana, or are sometimes referred to as lagoons in English. Only those along the Maunachira channel system have been subject to detailed study (McCarthy et al., 1993b).

The lakes along the Maunachira are oxbows but are not related to the present Maunachira channel. The channel system that gave rise to these oxbows appears to have been about 200 m wide, substantially greater than the Maunachira (20 m) and even the Okavango River today at the apex of the Panhandle (ca. 100 m), and reflects a period of much greater discharge onto the fan. The clear definition of the meander-related features suggests that they are relatively young, and luminescence dating currently in progress (G. Duller and S. Tooth, personal communications) indicates an age of around 6000 years BP for this channel system.

Lakes in the backwater areas of the upper reaches of the permanent swamps are largely unstudied (Child and Shaw, 1990). The location of these lakes in the lower Panhandle and upper fan suggests that the most likely modes of formation is as oxbow lakes or as segments of large channels that became isolated as surrounding channels were abandoned. However, more study is needed to determine their origins.

Not all lakes represent former channels. Lakes can form when a channel system is diverted and causes flooding of an area previously occupied by seasonal swamp, resulting in permanent inundation (McCarthy et al., 1993b). This produces areas of open water typically less 1 to 2 m deep. Lakes formed in this way have an irregular outline. Their beds are gently undulating and colonized by a variety of submerged aquatic plant species. An example of a lake of this type is Xakanaxa Lake. The water is derived from back swamp areas and is low in nutrients and there is little inflow of particulate sediment. Consequently the filling of these lakes is extremely slow.

The manner in which these shallow, distal lakes ultimately fill has been described by K. Ellery et al. (1990). The process commences with the accumulation of fine organic detritus on the lake bed, forming organic-rich sludge. Plants take root in this material, especially water lilies (*Nymphaea* spp.), but because the lakes are distal they are generally poor in nutrients and hence plant cover is sparse. Anaerobic decay of the sludge produces methane gas, some of which escapes, bubbling to the surface. However, some gas remains trapped in the organic layer, especially where it densely impregnated by lily rootlets. This renders the organic layer buoyant and clumps typically a quarter to half a square metre in area tear away from the bed and rise to the surface. They may remain partly attached or may be free-floating and are known as sudds. Free-floating bodies are generally blown to the leeward side of lakes by the prevailing wind. The subaerially exposed surface of the floating sudds is colonized, initially by sedges (particularly *Pycreus nitidis*) and then by the grass *Miscanthus junceus*, which stabilizes them. As the process proceeds, the lake may take on a patchwork appearance, reflecting the range of successional stages, but ultimately is converted to a homogeneous, *Mischanthus* grassland. Initially the grassland is floating but in time peat accumulates in the water-filled void below. The process is extremely slow, especially in the very distal areas where nutrient levels are very low, and the lakes probably take many decades to centuries close over.

How the Okavango system functions

As a result of research over the past five decades we now have achieved a fairly comprehensive understanding of the way the Okavango ecosystem functions.

The nature of its catchment contributes more than any other single factor to the character of the Okavango wetland. The low relief catchment is largely underlain by aeolian sand of the Kalahari Group, except for a small area of granitic basement near the headwaters of the Cubango River (Figure 16). The upper catchment lies in a region of high rainfall (Figure 9), which infiltrates the sand and is then released as groundwater seepage (via dambos) to the Cuito and Cubango Rivers. These various factors have the cumulative effect of delaying, modulating and extending the seasonal flood pulse, so that there is a steady seasonal rise and fall of discharge in the Okavango River. Short-lived flood pulses simply do not occur. The seasonal flood pulse of the Cuito River is further delayed and attenuated by extensive floodplains along its lower course, so that the peak discharge in the Cuito occurs a month later than that of the Cubango (Figure 23) thereby extending the duration of the seasonal flood wave into the Okavango Delta. The fact that much of the supply of water to the Cuito and Cubango is via groundwater seepage means that there is substantial base flow in the Okavango River, which sustains the permanent swamps in the Delta downstream.

The sandy nature of the catchment and the general low level of chemical weathering results in a very low suspended load and the bulk of the particulate sediment emanating from the catchment consists of aeolian sand that is carried as bedload. Moreover, the low level of chemical weathering results in a low solute load, and the total dissolved solid concentration in Okavango River water is in the order of 40 mg/L (Hutchins et al., 1976b; Hutton and Dincer, 1976; Sawula and Martins, 1991), half of which consists of silica and the remainder of bicarbonates of Ca, Mg, Na and K. The annual water and sediment budget is shown in Table 4.

Table 4. Estimated annual sediment and water budget for the Okavango Delta system (McCarthy and Metcalfe, 1990; McCarthy and Ellery, 1998)

	Inflow	**Outflow**
Rainfall	6.14×10^9 m^3	
Surface flow	9.20×10^9 m^3	0.24×10^9 m^3
Bedload sediment	170 000 tonnes	nil
Suspended sediment	39 000 tonnes	nil
Total solute load	381 100 tonnes	23 450 tonnes
Breakdown of solute load		
$CaCO_3$	114 900 tonnes	5 310 tonnes
$MgCO_3$	14 100 tonnes	1 640 tonnes
SiO_2	147 000 tonnes	8 300 tonnes
$NaHCO_3$	67 100 tonnes	5 600 tonnes
$KHCO_3$	33 000 tonnes	2 600 tonnes
Aerosol fallout	250 000 tonnes	

It is evident from the data in Table 4 that aerosol fallout contributes significantly to the Okavango Delta system and may be especially important to nutrient supply (Garstang et al., 1998) but as it is uniformly spread over the entire Delta it has little site-specific impact and will not be considered further. Although the dissolved solid concentration in Okavango River water is low, the total solute load substantially exceeds the total particulate load because of the very large annual inflow. The clastic sediment load is dominated by bedload.

The Okavango River system straddles a marked climatic gradient (Figure 9). The Delta is situated on the fringe of the semi-arid Kalahari Desert, where annual evaporation is three to four times rainfall. As a consequence, very little of the water leaves as surface flow and there there is very little groundwater outflow. No clastic sediment is exported from the Delta and only a limited quantity of the solute load leaves as surface outflow. It is evident from the Table 4 that the mass of solutes deposited annually exceeds that of particulate sediment.

Because of the very low abundance of suspended load in the Okavango River, its role has been taken over by plants that form the channel banks. Unlike normal river banks, the banks of Okavango channels are porous to water, but confine bedload. In addition, they are active in the sense that bank-forming plants strive to extend into the channel because it offers open space, and can also grow upwards, forming vegetative levees. Nutrients are in short supply in the Okavango, and become rapidly depleted in standing water by plant growth. Through-flowing water constantly replenishes nutrients, enhancing the growth of bank-forming plants and promoting channel invasion and levee formation by vegetation. This results in constriction of channels and the raising of the water surface in the channels relative to the surrounding back-swamps, further promoting water loss. The net effect is a steady decline in channel discharge downstream. As a consequence, sediment accumulates on the channel bed, slowly elevating the channel relative to its surroundings. This creates a feed-back loop that promotes further water loss and encourages upward growth of bank-forming plants, further raising the height of the levees.

The consequences of this aggradation seem to differ between the Panhandle and the fan proper. In the Panhandle, water leaked from channels gathers in the backswamps and flows along the network of interconnected lakes (mainly oxbows and sloughs) and hippo trails, creating in effect yazoo streams, which ultimately reconnect to the main channel. In time, the elevation differential between a yazoo stream and the main channel encourages a direct connection, again exploiting hippo trails. These become widened by the rapid flow out of the main channel, resulting in an anastomosing channel network, as can be seen in the central section of the Panhandle (e.g. Filipo channel). The ultimate outcome is an avulsion as portion of the former main channel is abandoned and flow is diverted into the subsidiary channel. In its terminal stages, the abandoned channel is rapidly overgrown by vegetation.

On the fan, the outcome of channel bed aggradation is different. Here, water loss from an aggrading channel may give rise to a new channel system in the flanking back-swamps, also exploiting hippo trails, but the two channel systems do not connect. Transient connections might form along hippo trails or dug channels, but these are short-lived. The channel-flanking vegetation grows vigorously in these situations because of nutrients provided by through-flowing water. Runaway aggradation in the failing main channel causes complete failure, and as the main channel fails progressively from its distal end, so the new channel may propagate alongside the failing channel, the two overlapping but always separated by a filter. The Nqoga channel is failing in this way. It seems that a subsidiary channel such as the Khiandiandavhu or Maunachira may not always form as an aggrading channel fails. Water is currently leaking through both the north and south banks of the Nqoga, but a through-going channel system has only formed to the north. To the south, no distinct, integrated channel system has formed, and the flow is more diffuse. Similarly, the progressive failure of the Thaoge was not accompanied by secondary channel formation. It is not yet understood why subsidiary channels form in some cases but not in others.

The ultimate outcome of the failure of the Nqoga channel is uncertain at this stage. It has become elevated relative to the Jao/Boro system (Gumbricht et al. 2001; Ellery et al., 2003) and the elevation differential is probably steadily increasing as the Jao erodes its bed whilst the Nqoga aggrades. As the terminus of the Nqoga approaches the Jao take-off, which at the current rate of failure will occur in about 60 to 80 years from now, the elevation differential will probably be substantial and it possible (likely?) that the Nqoga will divert into the Jao. The implications of such a diversion could be profound: the entire swamp area to the north and northeast of Chief's Island may desiccate and revert to dry land or perhaps seasonal swamp, depriving the Moremi Reserve of all of its permanent swamp.

The distributary channel systems on the Okavango fan appear to have a life span of between 100 and 200 years. The main mechanism responsible for redirecting the channels is bed aggradation. The swamp flanking the channel aggrades sympathetically, and captures suspended load from the water leaked from the aggrading channel. Once the channel fails, the surrounding swamp is deprived of water and desiccates in the dry climate. Peat fires ultimately destroy the organic material, leaving a veneer of ash derived principally from the suspended sediment load. The former channel bed is, on average, slightly elevated relative to this ash-covered surface. Peat fires thus lower the surface, but nevertheless the inorganic component remains and constitutes an increment of aggradation on the land surface. The removal of the peat primes the surface for re-flooding in the future. The ash residue from peat burning is relatively rich in nutrients, and once the summer rains fall theses areas become highly productive grasslands and attract large herds of grazing animals. Insects also flock to these areas, amongst them the *Macrotermes* termites, which begin construction of their massive mounds.

The main sediment load carried by the Okavango River is its solutes, the bulk of which are also deposited within the Delta. The processes responsible are well understood. As the seasonal flood advances, water infiltrates into the ground, raising the water table to surface, a necessary requirement for further advance of the flood. The further the water progresses down the fan, the deeper the water table and hence the greater the amount of infiltration. In addition to infiltration, some of the water evaporates. As the floodplains become inundated, aquatic plants that have been lying dormant in the soils, as well as their seeds, sprout and cover the surface in a carpet of vegetation. These plants transpire water into the atmosphere. The gently undulating topography results in a profusion of islands, which become ever more abundant towards the distal parts of the fan. There islands support trees, which have a high transpiration rate, so much so that they depress the water table below the islands. Evaporation and especially transpiration result in a steady rise in the dissolved solid concentration in both surface and ground water, leading ultimately to saturation in solutes in the groundwater. Saturation very seldom occurs in surface water.

The nature of the solute load plays a critical role in the outcome of prolonged evapotranspiration. Silica is the most abundant solute and is therefore the most abundant chemical precipitate. Sulphate and chloride abundances are vanishingly small and hence the major cations precipitate as carbonate species. The Delta would have been very different had chlorides dominated the solute load. The dissolved solutes saturate at different stages of evapotranspiration. The first to saturate is silica, which precipitates in the soils of floodplains and islands. Because of the low concentration of solutes, about 80% to 90% of the water has to have been lost before the water becomes saturated in silica and precipitation commences. By this stage there is little surface water left, and precipitation occurs primarily from the groundwater. Groundwater is drawn towards the islands because of the lowered water table where further loss occurs, resulting in saturation in calcite, which precipitates in the island fringe. This occurs after about 95% of the original water has been lost by evaporation and transpiration. At this stage, the solute load is dominated by potassium and sodium carbonates and bicarbonates. The pH rises, approaching 9, and kaolinite in the soils reacts with the potassium forming microcline, leaving only sodium bicarbonate in solution. Sodium bicarbonate is extremely soluble, and although it is not toxic, it begins to impact on the vegetation as its concentration rises due to osmotic effects, destroying all vegetation in the interiors of the islands. Capillary rise draws the saline water to surface where it evaporates, leaving a salt crust on the surface. This is re-dissolved and washed back into the soil during rainstorms, and the salinity of the groundwater steadily rises. The density of the water also rises, and eventually the dense brine begins to sink down into the deep groundwater. This process puts a ceiling on the maximum salinity of the groundwater. Islands reach this stage after about 200 years, about the same time as the life of a channel system.

Approximately 85% of the solute load is precipitated as amorphous silica and carbonate and silicate species, all of which are benign. Their precipitation causes swelling and the islands grow. It is this swelling that creates the gently undulating microtopography of the Delta. The sodium salts, which become toxic to plants by virtue of the high concentrations attained, sink into the deep groundwater. There is therefore minimum impact on the surface, and in particular, the surface water in the Delta remains fresh at all times. Over time, island interiors become extensively salinized and the barren area gradually expands. In extreme cases all of the island vegetation may be destroyed. However, this takes considerable time and diversion of the surface water as a result of upstream channel failure invariably occurs before islands on the floodplain are completely salinized.

Once floodwater is diverted, the islands cease to grow. The water table beneath the floodplains falls and the only means of groundwater replenishment is via infiltration of rain water. Such infiltration dissolves sodium salts held in the soil and dilutes the salts in the groundwater. The calcite and silica in the soils are, however, unaffected and the increment of island growth they have caused is permanent. In time, flooding will return to a particular region and once this occurs, the islands will be rejuvenated. They will become populated by trees and the island growth process will begin anew. Over repeated cycles of growth in this way, islands slowly enlarge and eventually merge, forming extensive platforms, which are essentially very large islands.

Although islands are present on the upper fan, they are aerially of less importance than the wetlands and aggradation in this region is primarily caused by clastic sedimentation, although even the upper fan is from time to time freed from permanent swamp cover. The mid and lower fan regions aggrade mainly by chemical sedimentation. Despite their very different nature, the two processes remain in balance so that the uniform gradient of the fan is maintained. Both processes are dependent on the distribution of water. If an area becomes too elevated, water will be excluded and sedimentation will cease. If the area is of low elevation, water and sediment will continue to collect there until the elevation is raised.

The constantly shifting water distribution across the fan surface, coupled with the many feed-back loops that operate in the seasonal and permanent swamps, result in constant environmental change. The system is constantly being disturbed on a time scale of decades to centuries and climax vegetation communities are never able to develop. It is this constant change that creates the wide diversity of habitats and beauty of the Okavango. This constant change and self-renewal is essential for the long term well being of the Delta ecosystem.

Conclusions

We now have a fairly comprehensive understanding of the development of the southern Africa landscape since the break-up of Gondwana and how it has been influenced by climate change and epeirogenic uplift. Quantitative analysis of the high resolution digital topography has shown that Alex du Toit's original notion of the interior of southern Africa consisting of an extensive, planed land surface located essentially at the erosional base level (now referred to as the Africa Surface) appears to have been correct. His notion of epeirogenic uplift, commencing at about 20 million years ago, has likewise been vindicated by recent work. The landscape we see today therefore has a complex structure: some large scale features of the landscape such as the Great Escarpment are relatively young, whereas others are ancient and date back to the time of Gondwana.

While we have a grasp of the broad picture of landscape development, there are large gaps in our knowledge of key elements of the history of the region. Perhaps foremost amongst these is our poor understanding of the palaeo-environments that prevailed during the early stages of deposition of the Kalahari Group and the timing of that deposition. This is partly a consequence of poor exposure, but lack of concerted research is possibly the main reason. The Kalahari basin offers a unique window into climate change through the Cenozoic and especially during the Neogene and the Pleistocene. To date, the only firm information we have relates to late Pleistocene and Holocene climates and information on earlier climates, and especially the origin and timing of the Kalahari sand sea and its massive dune systems, is sparse. It is evident from the chronology that has emerged that the region has been generally dry with brief wetter periods, but the number of age determinations is far too few, and many are insecure. More age determinations are needed, especially of the early stages of dune systems, and on lacustrine features, especially lake deltas, that provide unambiguous information about conditions prevailing during deposition. The availability of the SRTM data set provides new opportunities for geomorphological analysis and investigation, and coupled with luminescence dating should enable new insights to be obtained on climatic changes in the region. Also needed are more studies of the Benguela current and especially ocean temperatures off the west coast, as this system is clearly an important determinant of climate in the interior. Ultimately what is needed is a high resolution climatic record for southern Africa through the Neogene and Pleistocene and to determine the relationship between local climatic switching and global climate events.

As far as the Okavango Delta is concerned, the functioning of the system is now generally well understood. It is a highly complex system and its present form is a product of interactions between past and present climate, geomorphology, tectonics, sedimentation, geohydrology and the biota.

Perhaps the most remarkable aspect to emerge from this review is how many of the key ideas in our understanding of the post-Jurassic history of southern Africa stem from observations first made by Alex du Toit eighty or more years ago. Amongst these are the details of Gondwana break-up and the formation the coastline of the sub-continent; the idea of the post-break-up planation of southern Africa and the fact that it had been achieved by the early Paleogene; the notion that the Limpopo drainage basin was formerly far larger than today and included the Okavango and Linyanti; the notion of large-scale continental warping, which influenced the major watersheds of the sub-continent and led to the formation of the Kalahari basin; the notion that the Kalahari basin once hosted great lakes; the desiccation of southern Africa and the formation of the Kalahari sand sea, the largest continuous aeolian sand sheet on Earth; the notion of uplift of the subcontinent and the timing of that uplift; and the propagation of the East African Rift system into southern Africa and its role in the formation of the Okavango-Linyanti depression. The fact that so many of du Toit's ideas have stood the test of decades of subsequent scrutiny, some involving technology that to him would have seemed in the realms of science fiction, attests to his extraordinary observational skills, his perceptive mind and his deep understanding of geological processes.

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