

THE GEOLOGY, PALAEOLOGY AND EVOLUTION OF THE ETOSHA PAN, NAMIBIA: IMPLICATIONS FOR TERMINAL KALAHARI DEPOSITION

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ABSTRACT

A mid to late Tertiary lake, Palaeolake Etosha, was the single collective end point of three drainage systems feeding southwards off the central Angolan highlands, the Cubango and Kunene Megafans and the Cuvelai Drainage System. The lake beds consist of 50 m of saline, olive green clay, the Etosha Pan Clay Member, containing authigenic analcime, monoclinic K-feldspar and glauconite. Fossil suites in two local sandstones near or at the top of the clay, the Oshigambo and Ekuma Delta Sandstone Members, include various antelope, pig, zebra, quagga, rhinoceros, elephant, lion, spring hare, ostrich, flamingo, crocodile, hippopotamus, freshwater turtles, catfish, bivalves and trees, which have respective ages of 6 ± 1 Ma and 4 ± 1 Ma. An oolitic, pan-margin limestone of limited lateral extent, the Poacher's Point Carbonate Member, was the last unit deposited in the palaeolake. Flooding and complete desiccation were common occurrences in Palaeolake Etosha before it dried up completely at about 4 Ma under conditions of progressively increasing aridity. Over the ensuing 2 Ma, a huge groundwater calcrete, the Etosha Calcrete Formation, advanced northwards over the dry lake deposits and the surface Kalahari sands were reworked to produce the regional Kalahari dune fields.

Resumption of periodic flow in the Cuvelai system resulted in cycles of flooding and desiccation in an ephemeral Cuvelai end-point lake atop the Etosha Calcrete Formation in the Etosha area. Fragmentation of the calcrete by salt enabled strong easterly winds to remove the calcrete fines from the Etosha depression. This initiated the 'excavation' of the present Etosha Pan by wind ablation. Lunette dunes on the western side of Etosha Pan contain an inverted pan stratigraphy. Flooding of the present pan occurs occasionally and has been as deep as 10 m in the past. White magadi-type chert nodules have formed locally on the pan floor in animal footprints.

Introduction

Three palaeo river systems dominated upper Kalahari deposition in the Owambo Basin. One of these, the Cubango Megafan, and the present drainage basins that evolved from the palaeo systems are shown in Figure 1. The Etosha Pan is located in the topographically lowest part of the Owambo Basin of northern Namibia near its southern margin (Figure 2). Most of the floor of the pan has an elevation of 1080 m or close to it. Nevertheless, the two present perennial river systems of northern Namibia, the Cubango-Okavango in the east and the Kunene in the west, bypass the Owambo Basin (Figure 1). Consequently, Etosha Pan is usually dry but during occasional periods of heavy summer rains, it can be partially or almost completely flooded, mainly by inflow from the Cuvelai Drainage System via the Ekuma River to the northwest but also minimally from marginal rivulets and very rarely from westward-directed

flow in the Omuramba Owambo into Fischer's Pan (Figure 3).

Stuart-Williams (1992) and Marsh and Seely (1992) suggested that the saline, pale olive green clays that form the bed of the pan were deposited in a Tertiary palaeolake that extended up to and was fed by the Kunene River. Rust (1984; 1985), Buch et al. (1992), Buch and Rose (1996), Buch and Trippner (1997) and Hipondoka (2005) have studied the surroundings of the pan and the upper three metres of the clay. These authors provided evidence of the pan being an ablation feature, having been excavated by strong easterly winds. Hipondoka (2005) discovered and was the first to describe mammalian fossils at the northwestern margin of the pan.

This paper investigates how three megafans originating in the southern highlands of central Angola have contributed the stratigraphy in the Etosha Pan area.

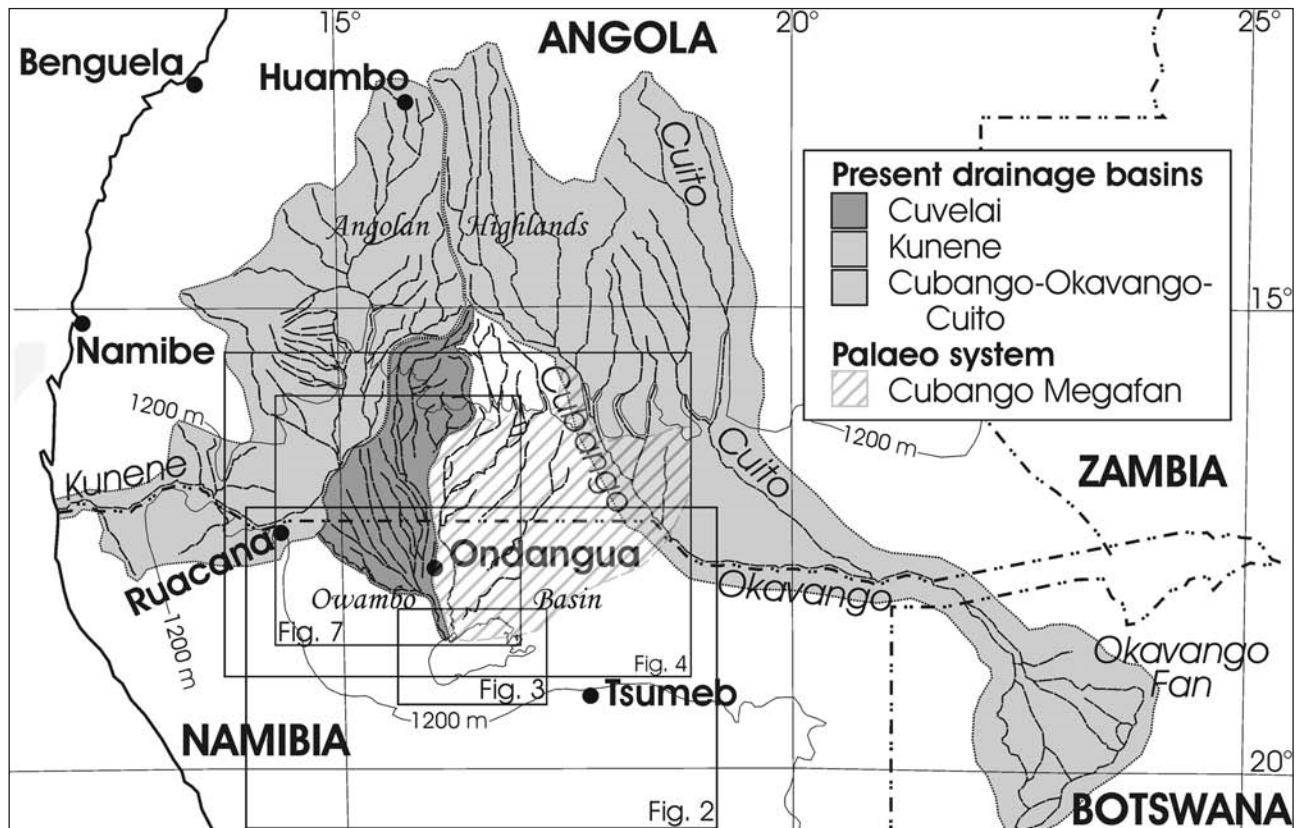


Figure 1. Map showing the location of the Etosha Pan relative to the main present-day drainage basins to the north of Etosha Pan as well as the location of the older Cubango Megafan (modified from Stengel 1963; Mendelsohn et al., 2002; 2004). The locations of the other map figures in this paper are also shown.

This area was the single, collective, low-elevation end point of each of these fans from approximately the mid Tertiary. The fossils listed and briefly described in this paper have enabled certain aspects of the depositional history and evolution of the area from a Tertiary Palaeolake Etosha to the present, normally dry Etosha Pan to be dated.

Structure of the Owambo Basin

Wellington (1939) suggested that the Owambo Basin was structural in origin. Aeromagnetic surveys (Eberle et al., 1995; 1996), a few deep, widely spaced boreholes (Hugo, 1969; Hedberg, 1979) and some widely spaced seismic survey lines shot in the 1960s (Hedberg, 1979) reveal only limited faulting (Miller and Schalk, 1980). The aeromagnetic surveys and the boreholes reveal that it is a deep, long-lived sedimentary basin containing thick Neoproterozoic sedimentary rocks, Karoo sedimentary rocks and basalts and thick Kalahari sediments. These units are separated from each other by major erosional unconformities (Miller, 1997; 2008). The aeromagnetic surveys reveal displacement of the Karoo rocks by a few roughly east-west faults and a northward-fanning spay of faults, some of which have been intruded by dolerite dykes (Miller, 2008, Figure 16.8). A north-northeasterly trending fault with a displacement of <50 m occurs parallel to and just south

of the southern edge of Etosha Pan (Miller, 2008, Figures 16.8; 24.11). The surveys do not reveal any major basin-margin faults. The seismic surveys and the subsequent exploration wells reveal the Neoproterozoic fold structures occurring at ever greater depths as one moves towards the centre of the basin (Hugo, 1969; Hedberg, 1979). Thus, the Owambo Basin appears to have attained its basic shape as a result of uplift along its margins during late Neoproterozoic continental collision in the Kaoko Belt to the west and the Damara Belt to the south (Miller, 1983; 2008; Goscombe et al., 2003). This shape has scarcely changed since then. The deepening of the bedrock floor in the centre of the basin was enhanced by pre-Karoo and pre-Kalahari erosion (Miller, 2008). Borehole data in the southeastern part of the basin suggest the presence of a deep, northwest-trending valley in the pre-Kalahari bedrock (Figure 2).

Alluvial megafans in the Owambo Basin

One of the earliest usages of the term 'megafan' was by Gohain and Parkash (1990) in their description of the Kosi Fan of northern India. With the aid of SRTM imagery, Wilkinson et al. (2006, 2008) recognised up to 150 megafans within inland sedimentary basins. They describe a megafan as a low-angle, partial cone of fluvial sediment that arises where major rivers enter an unconfined basin. Fan radii reach several hundred

kilometres. One such fan identified by Wilkinson et al. (2006; 2008) is the well developed Cubango Megafan in the Owambo Basin (Figures 4 and 5). Two additional fans occur in the basin, the Kunene Megafan and the Cuvelai Drainage System (Figure 4). These fan systems contributed to the late stages of sediment accumulation within the Owambo Basin but in the absence of numerous deep and well logged boreholes, it is impossible to ascertain how important these were during earlier stages of Kalahari deposition. Nevertheless, the size of the Cubango Megafan suggests that it, at least, must have played a major role throughout much of the Kalahari depositional history within the basin.

The Cubango Megafan

Recognised initially by Wilkinson et al. (2006), this megafan has a gentle but well defined convex-up cross section with curvilinear contours which define an almost north-south orientation of the fan axis (Figure 4). The fan arises at the southeastern foot of the central

Angolan highlands and its form and orientation indicate that towards the end of Kalahari aggradational accumulation, the Cubango/Okavango River built a huge megafan that prograded almost due southwards and reached the eastern edge of the Etosha Pan. The north-south length of the megafan along section line CDEF is 350 km and the east-west width at least 300 km at the widest point. The southward slope in the northern parts of the megafan is approximately 0.00056 over a distance of 95 km ($\sim 0.034^\circ$, section CD in Figure 4). Downstream, this decreases to 0.00028 over a distance of 255 km ($\sim 0.017^\circ$, section DEF in Figure 4). The approach to the depression of Etosha Pan becomes apparent some 40 to 50 km from the pan through an increase in slope (Figure 6). Most of the palaeo water courses on the western slopes of the Cubango Megafan surface rarely flow since they are choked by aeolian sands deposited after formation of the fan. Many of the larger river channels in this region terminate before reaching the Etosha Pan (Figure 7). Nevertheless, a large number of pans occur across the axis of the megafan just

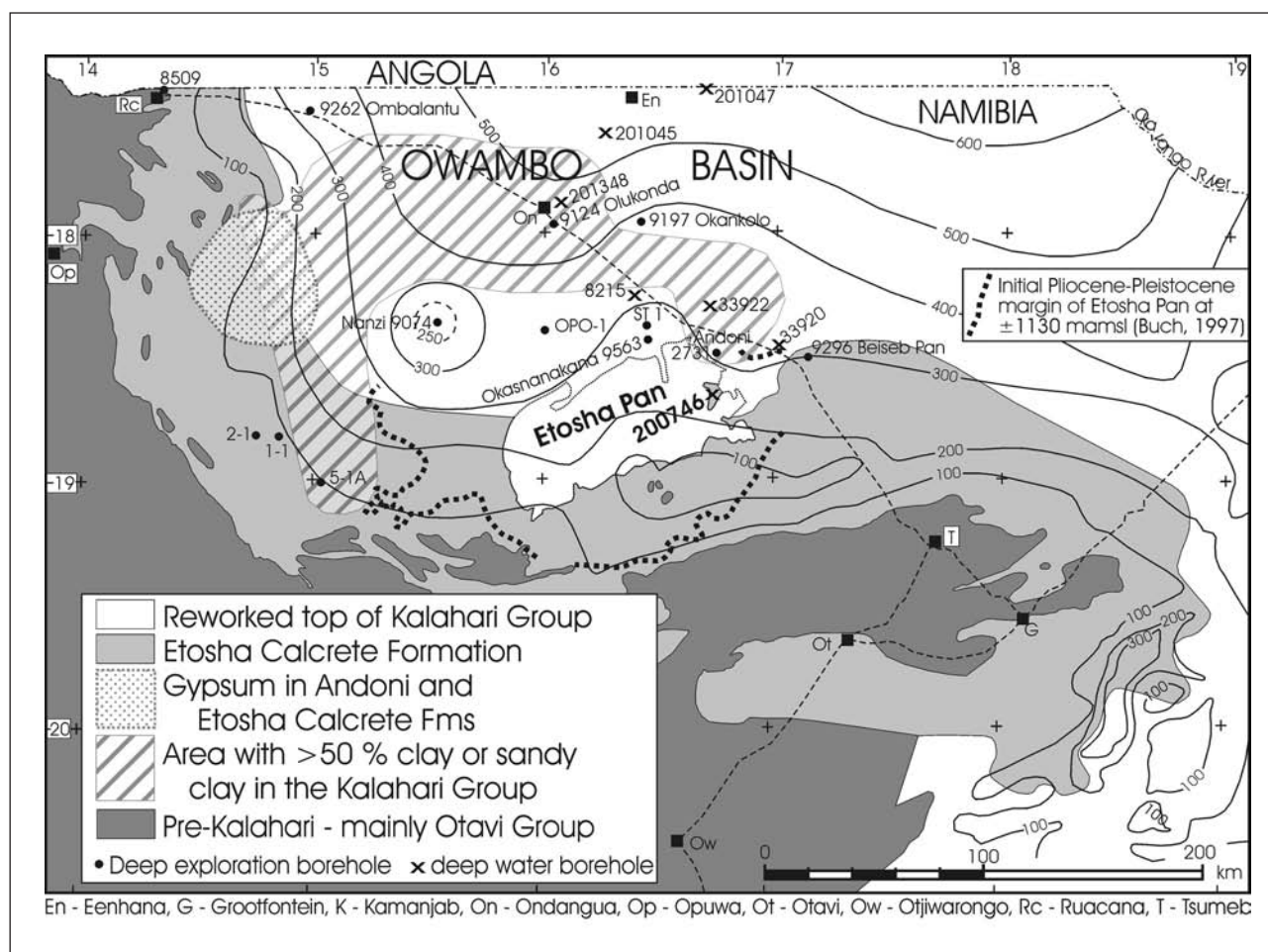


Figure 2. Map showing the location of the Etosha Pan and deep boreholes in the Owambo Basin (modified after Miller, 1997; 2008). Isopachs of the Kalahari Group are from Miller and Schalk (1980) and are based on deep oil and coal exploration boreholes (Hugo, 1969; Hedberg, 1979), seismic data shot in the 1960s (Hedberg, 1979), numerous water boreholes in the regions of thinner Kalahari, a few deep water boreholes in regions of thick Kalahari (numbered only where referred to in text), and recent 400-m deep water boreholes in the Eenhana area (En in the figure; Miller, unpublished data).

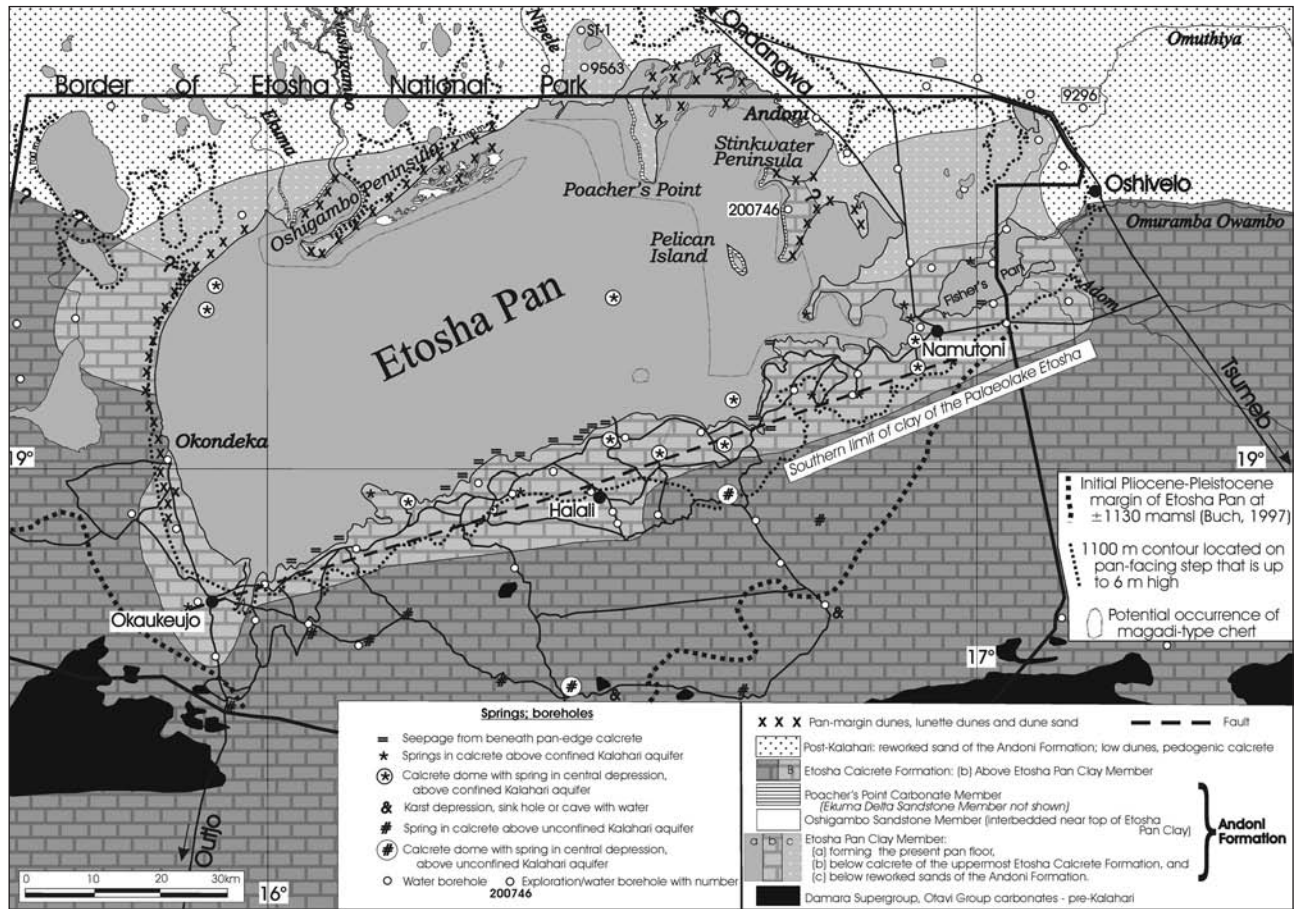


Figure 3. Map of the Etosha Pan area showing: (1) The distribution of the Etosha Pan Clay Member, (2) The approximate location of the Oshigambo Sandstone Member, (3) The extent of the Etosha Calcrete Formation, and (4) The Ekuma and Gwashigambo Rivers within which the Ekuma Delta Sandstone Member is exposed. The location of boreholes used in the construction of the figure are shown (modified after Miller, 2008).

north and south of the Angolan border (Surveyor General, 1976; Mendelsohn et al., 2002).

The initial steeper gradient of 0.00056 in the proximal section of the Cubango Megafan and the gentler gradient of 0.00028 in the central and distal parts of the fan both fall into the field for braid-dominated fans of Stanistreet and McCarthy (1993). Aeolian sands obscure the older river courses so it is not possible to determine whether they were braided or meandering.

The present Cubango/Okavango River stands out in stark contrast to the broad north-south megafan the river initially produced. The present river has carved a relatively straight, southeasterly directed valley 60 to 70 m deep into the eastern sediments of the megafan.

The Kunene Megafan

Wellington (1938) suggested that the Kunene River was the primary source for the clays of the Etosha Pan. However, this is not obvious from borehole data. Nor does the borehole data (unpublished borehole records, Department of Water Affairs, Windhoek) support the contention that the palaeo Hoanib River formed a southern outlet to such a lake or the Kunene River as

proposed by Stuart-Williams (1992). In contrast to the Cubango Megafan, the Kunene Megafan is much smaller and far less obvious. The 10 m contours defining it are only slightly curvilinear and have an almost straight east-west orientation. In the north, the fan occupies a rather narrow valley coming out of the southwestern part of the central Angolan highlands but further south it fills a broader north-south depression, 100 to 115 km wide between the western edge of the Cubango Megafan and the elevated Otavi Group carbonates forming the western margin of the Owambo Basin. The slight southward bulge of the elevation contours appears to cease at the 1120 m elevation contour just south of point H on the section line GHJ (Figure 4). But the slight curvilinear trace of the contours could as well be due to the superimposed Cuvelai Drainage System. This superimposition makes the southern limits of the Kunene Megafan difficult to define.

Clay and/or sandy clay make up more than half of the Kalahari Group in the southwestern part of the Owambo Basin (Figures 2; 4). Most of this clay-rich region is located at the terminal end of the Kunene Megafan. This region also contains evaporites near the top of the Andoni Formation (Figure 2; Miller, 2008).

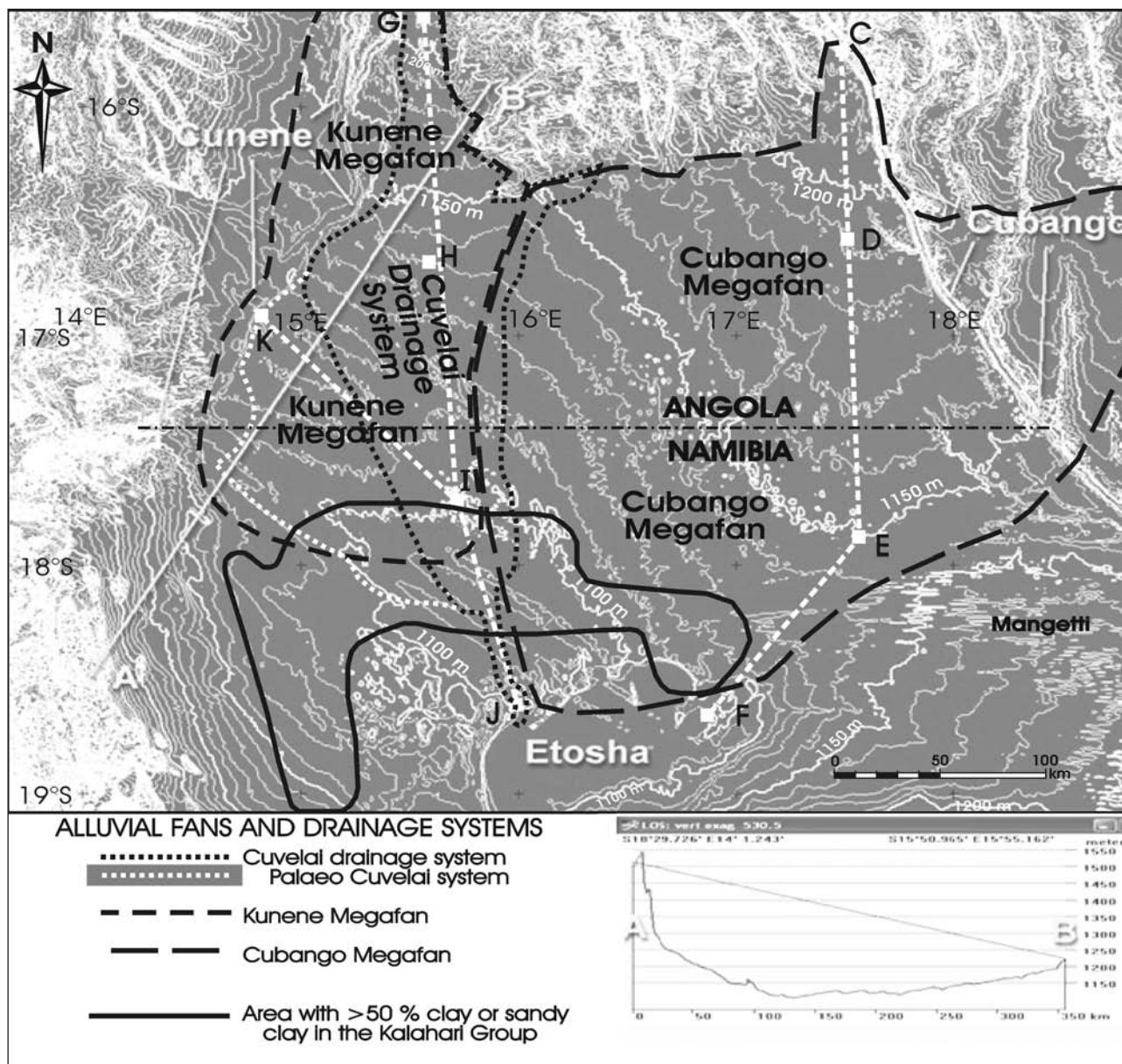


Figure 4. Contoured SRTM image of the Owambo Basin extending from the central Angolan highlands to the southern edge of the Etosha Pan; 10 m contour intervals. The Cubango and Kunene Megafans and the Cuvelai Drainage System are outlined (original contoured image provided by M.J. Wilkinson). Section lines CDEF, GHIJ and KIJ are plotted in Figure 6. (Figure modified after Miller, 2008).

The Cuvelai Drainage System

The Cuvelai Drainage System (Figures 4 and 7) was initiated by flooding of the Kunene River in a southeasterly direction into the Owambo Basin from roughly 50 km upstream of Xangongo to approximately Ruacana (Figure 7). The flooding incised the characteristic broad, shallow channels of the system. These channels are broad right up to the watershed that now separates the Kunene River from the Cuvelai system (Figure 7). The watershed developed as the Kunene River gradually deepened its present valley and, in so doing, eventually cut itself off from the Cuvelai system.

The Cuvelai Drainage System is superimposed on the Kunene Megafan and overlaps slightly onto the western

edge of the Cubango Megafan (Figure 4). The north-south length from the Angolan highlands to its end point at the mouth of the Ekuma River at the northwestern edge of Etosha Pan is just over 330 km. The Cuvelai system consists of two parts (Figure 7). The eastern part is the larger of the two and is the presently active system. It is fed primarily by the flooding of rivers flowing off the southwestern foothills of the central Angolan highlands but also by local rains within the system (Stengel, 1963). The western part does not receive any water from Angola and is a palaeo system fed only by local rains. Consequently, the Etaka River (Figure 7) flows infrequently and contributes only occasionally and in limited amounts to the flow via Lake Onopono into the Ekuma River.

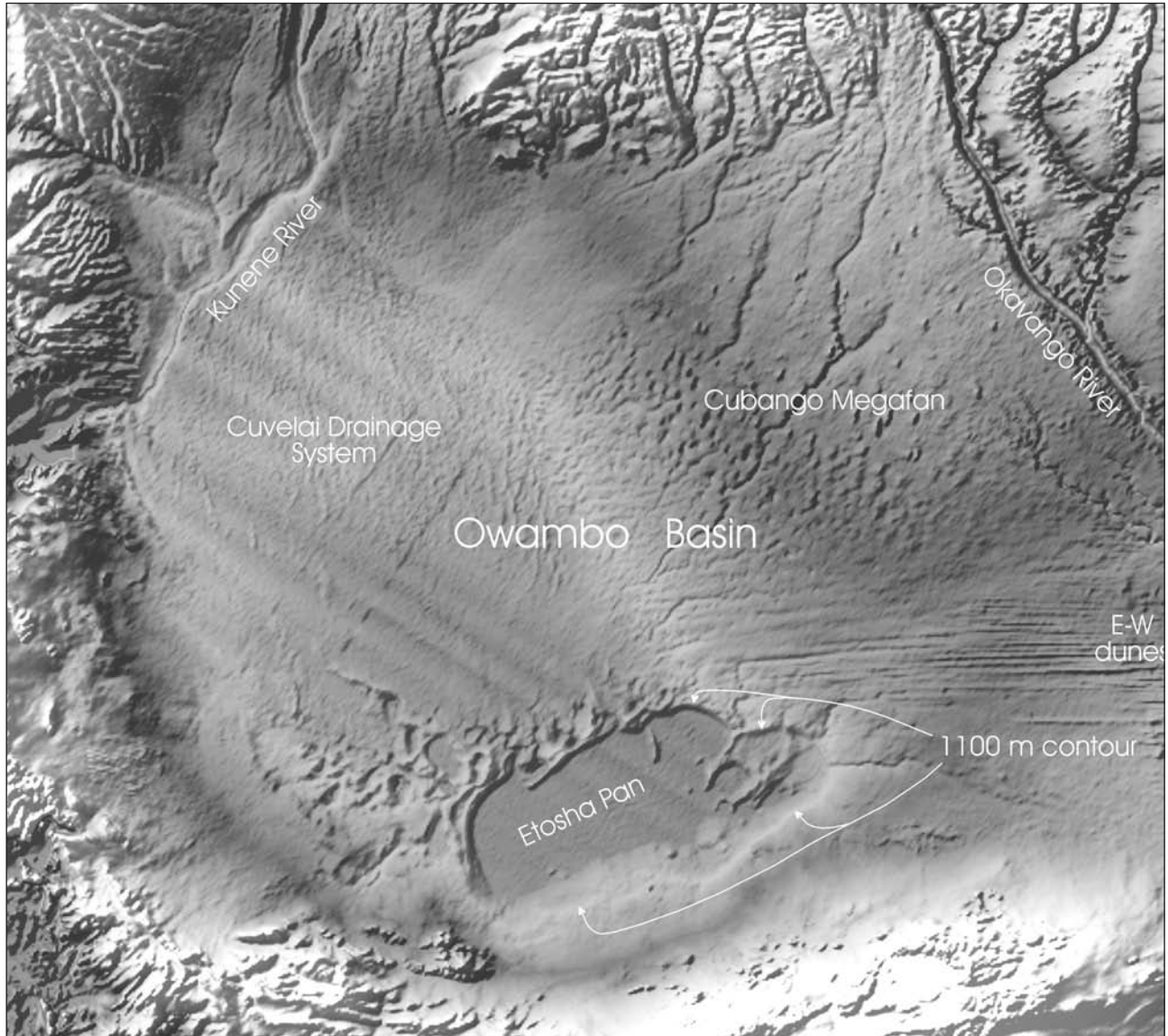


Figure 5. Digital terrain model of the Owambo Basin, northwesterly sun angle. A large elevation stretch has been applied between the elevations of 1096 m and 1177 m in order to emphasise small differences in elevation within this elevation range. Within the basin, the image shows the Cubango Megafan with a few southwesterly flowing rivers and east-west trending dunes on its southern termination. Further west, the 1 to 2 m deep channels of the Cuvelai Drainage System are generally below the resolution of the stretch and are poorly defined. Widely spaced, low-elevation dunes with north- to northeasterly trending crests are discernable along the northern margin of the pan. A pan-facing step of up six metres in the landscape follows the 1100 m contour outboard of the pan margins giving the impression that the pan outline in this figure is large than it is in Figure 3 (1100 m contour shown in Figure 3). On the Cuvelai Drainage System, the broad, linear, northwesterly to southeasterly trending shadows and the two thin northeast-southwest trending linears are artifacts. (Image kindly provided by Branko Corner of Corner Geophysics).

Because of the superimposition of the Cuvelai system on the Kunene Megafan, it is almost impossible to ascribe the 10 m contours to either one or the other. However, the Cuvelai system has a well defined, southeasterly-orientated subfan 130 km long and 100 km wide centred on the point K on the Kunene River (Figure 4). The section line KI roughly bisects this subfan and point I marks its distal termination. Although there is only an elevation difference of 30 m between points K and I, the contours have a marked curvilinear bulge away from the Kunene River. The subfan is an

early Cuvelai feature that formed during flooding of the Kunene River since it encompasses both the palaeo and the present parts of the system

The upper part of the present Cuvelai system drains southwards, is about 30 to 35 km wide at its apex and stays approximately this width down to about 16.25°S. It widens to some 100 km where rivers draining off the southernmost edge of the central Angolan highlands join it and stays approximately this width down to 16.75°S. An extensive region around Evale approximately 100 km north of the Angola/Namibia border (Figure 7 and

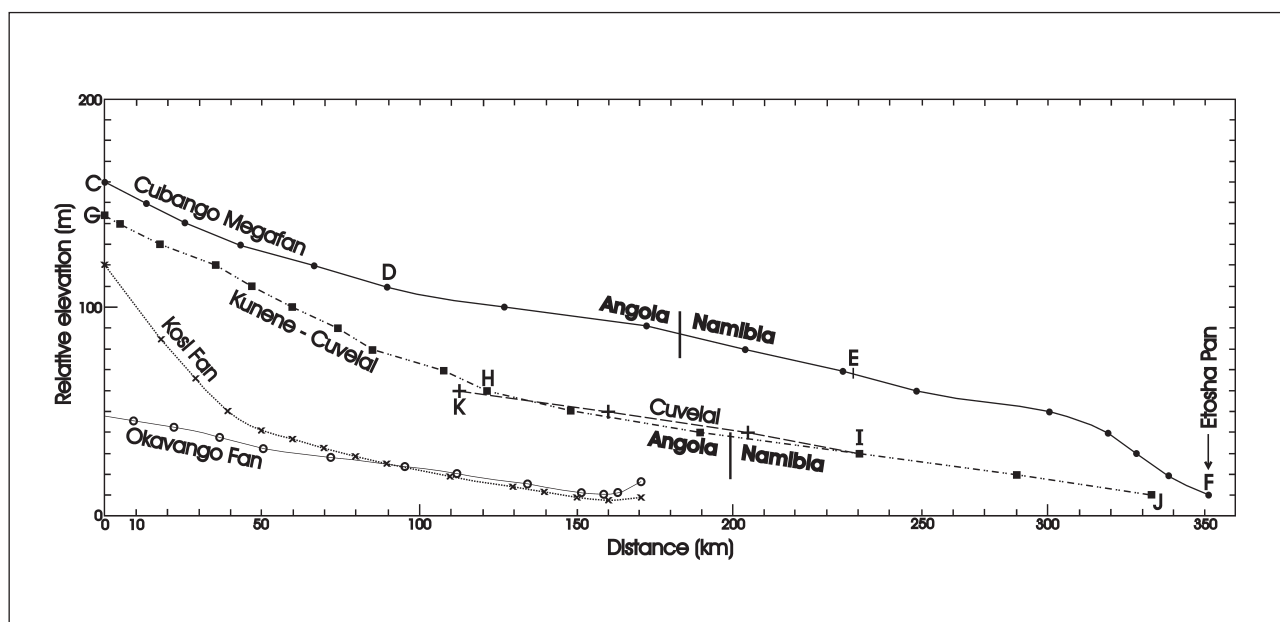


Figure 6. Comparison of the gradients of the Cubango and Kunene Megafans and the Cuvelai Drainage System with gradients of the Kosi and Okavango Fans described by Stanistreet and McCarthy (1993).

approximately at point H on the section line GHIJ in Figure 4) is heavily overgrown, swampy and with only poor development of channels (Stengel, 1963). South of 16.75°S, the slightly braided channel system gradually evolves into a system of subparallel channels with interlinking arms (*iishana*; in the Oshiwambo language, *oshana* is singular, *iishana* plural). These channels gradually assume a south-southeasterly flow direction. The parallelism and almost total lack of sinuosity of the

channels is striking. Linking of channels by means of side arms is fairly common. The greatest width is approximately 140 km just south of the Angolan-Namibian border. From this point, the system begins to converge downstream (Figure 7) becoming progressively narrower down to Lake Oponono (also referred to as Lake Oussouk by Stengel, 1963, and the Omadhiya Lakes by Mendelsohn and el Obeid, 2004). The lake is a series of extensive, shallow, grassy pans

Table 1. Record of 20 years of summer rainfall and flood data within the Cuvelai Drainage System (from Stengel, 1963). The data record one year with an exceptional flood, 12 years with floods of varying intensity and seven years without floods.

Year	Flood events	Rainfall in mm at Okatana and (Ondangua)	Comments
1941/42	Normal flood (\pm 5 million cbm)	507 (158)	
1942/43	Normal flood	800 (530)	
1943/44	Normal flood	893 (766)	Abundant fish
1944/45	Weak to normal	507 (469)	
1945/46	No flood	261 (223)	
1946/47	Normal flood	837 (670)	
1947/48	Very weak flood	544 (441)	Very localised
1948/49	No flood	590 (522)	
1949/50	Normal flood	972 (789)	
1950/51	Weak to normal	686 (753)	
1951/52	No flood	399 (246)	
1952/53	No flood	327 (316)	
1953/54	Abnormally high flood (\pm 110 million cbm)	600 (679)	Highest known flood
1954/55	No flood	254 (276)	
1955/56	Good flood (15 million cbm)	533 (457)	
1956/57	Very good flood (50 million cbm)	422 (343)	
1957/58	No flood	507 (451)	Only local flood
1958/59	Very weak flood (\pm 2 million cbm)	293 (370)	Only local flood
1959/60	No flood	418 (272)	
1960/61	Good flood (\pm 16 million cbm)	378 (341)	

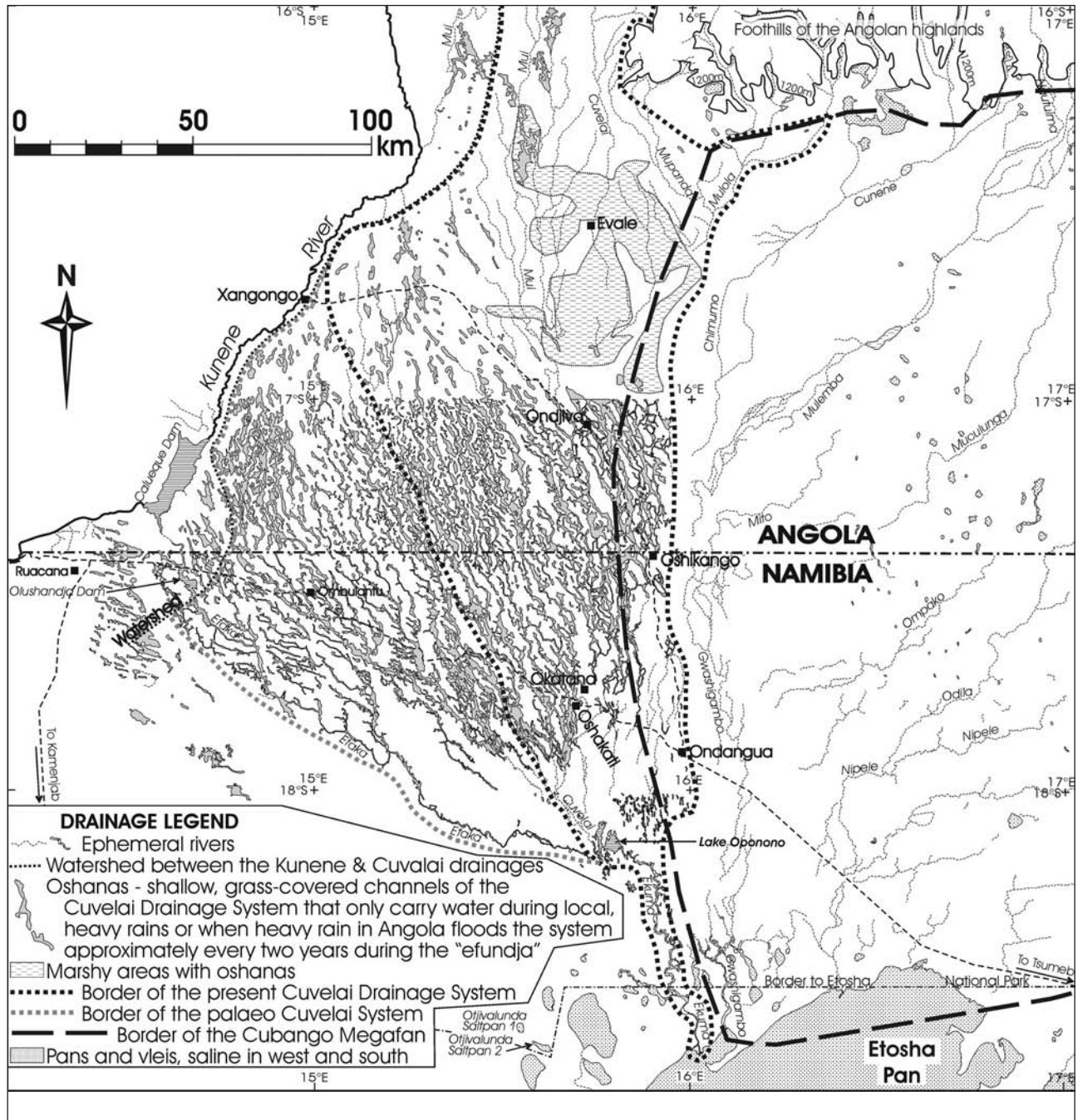


Figure 7. The present and palaeo Cuvelai Drainage Systems. The palaeo system appears to have been initiated by flooding of the Kunene River. This is no longer the case due to later incision of the Kunene River. A watershed separates both present and palaeo Cuvelai systems from the present Kunene drainage. Stengel (1963) included the Chimumo/Cunene River in the Cuvelai system but since this terminates approximately 7 km east of the nearest Cuvelai iishana (shallow channels), it is excluded from the Cuvelai system in this figure. Compiled and modified from Stengel (1963); the 1:1 m topographic map of southwestern Angola and northwestern Namibia (1982 – in Russian); 1:500 000 topographic maps of southern Angola, sheet 28, Ondjiva (1985), sheets 26 and 27, Xangongo (2000); 1:250 000 topographic maps of Namibia. Figure includes some Shuttle image interpretation where iishana detail is missing on the Angolan maps.

that coalesce into a single body of water during significant flooding. Overflow from the lake feeds primarily into the Ekuma River and thence into Etosha Pan. Some of the overflow fills a series of channels just east of the Ekuma River. This water is joined by limited amounts of water from the easternmost channels of the Cuvelai system and

finally feeds into the southern end of the Gwashigambo River. The bed of the Ekuma River is muddy but that of the Gwashigambo River is sandy. The Cuvelai River coming off the Angolan highlands (Figure 7) is perennial but the water disappears underground in the Evale region (personal communication, J. Mendelsohn, 2010).



Figure 8. Oblique aerial photograph taken during the 2009 Cuvelai system flood: view southwards across the completely flooded Ekuma River near its southern termination just before it enters the Etosha Pan. The flooded fence in the near middle distance is the northern border to the Etosha National Park (Photograph: D. Heinrich, reproduced with his kind permission).

In the present Cuvelai System, the gradient in the upper 120 km, between points G and H on the section line GHIJ in Figure 4 is 0.00069. From H to the end point of the system at the mouth of the Ekuma River at J, a distance of 212 km, the gradient is 0.00024 (0.014°). The 130 km long subfan on the section line KI has a slope of 0.00023. These gradients are identical to that of the losimean Okavango Fan of Stanistreet and McCarthy (1993). Thus, with the marked parallelism of the Cuvelai channels, the almost total lack of sinuosity and the extremely low gradient, the Cuvelai Drainage System has the key characteristics of a losimean fan as defined by Stanistreet and McCarthy (1993).

Width/depth ratios of channels in the Cuvelai system differ greatly from figures given by Stanistreet and McCarthy (1993) for the Okavango Fan. The average width/depth ratio for 92 measurements taken from the 1:50 000 topographic sheets is 448 with individual readings ranging from 86 to 1300 and widths varying from 200 m to as much as 1750 m, even along one and the same channel. Commonly, channels are only one or

two metres deep but tracts as much as 4 m deep also occur. Consequently, once flow stops, open stretches of water will last for months in parts of the system. Downstream, the number of channels decreases, several become significantly narrower and finally become very shallow rivers, some of which terminate before reaching Lake Oponono. Nevertheless, there is no significant change in the width/depth ratio of channels downstream. Many of the channels in the palaeo Cuvelai system are significantly narrower than those in the presently active system (Figure 7). The few longer rivers in the system are significantly narrower than the other channels but not deeper. Thus, the Cuvelai River just upstream of Lake Oponono has a lower average width/depth ratio of 285 (range 150 to 425), the Etaka 158 (range 90 to 340) and the Ekuma 187 (range 150 to 210).

The Cuvelai system floods between four and six times in every ten years (Stengel, 1963; Mendelsohn and el Obeid, 2004) but flood waters reach Etosha Pan far less regularly and usually in very limited volumes (Lindeque and Archibald, 1991). During the 2009 flood, the overflow from Lake Oponono completely filled the Ekuma River (Figure 8) but the flooding of the pan was not as extensive as it was in the 2008 flood. Flooding of the Cuvelai system depends very much on where heavy rains have fallen. Local heavy rains only cause local flooding. Only when heavy rains are widespread, particularly in Angola (Stengel, 1963), does large-scale flooding take place. This is illustrated in Table 1. During the 1953/54 season there were widespread rains in Angola and extensive flooding in Namibia but only 600 mm fell at Okatana (location in Figure 7). The 1956/57 and 1960/61 seasons were similar. Flow rate in the channels is not much more than 0.5 m/s and local back-flooding in a northerly direction can take place as channels and pans gradually fill up (Stengel, 1963).

Shallow, hand-dug wells that are replenished by the floods support a dense local population in the north half of the Namibian part of the system but water quality in the wells deteriorates towards the end of the dry season due to the widespread occurrence of a saline calcrete throughout the system and which commonly outcrops just above the floors of the iishana. Most iishana are

Table 2. Nomenclature of the Kalahari Group in the Owambo Basin (modified after Miller, 1997; 2008).

Central basin			Southern and western margins
Formation	Member	Lithology	Formation
Andoni	Upper	Poachers' Point Carbonate	Etosha Calcrete
		Ekuma Delta Sandstone	
		Etosha Pan Clay; interbedded Oshigambo Sandstone Member near top	
	Lower	Sand, clay; calcrete and dolocrete nodules	
Olukonda		Red sand and clay	
Beiseb		Red and white gravel	
Ombalantu (possibly pre-Kalahari)		Red sand, silt, clay, locally partially cemented by intergranular silcrete; local basal gravel	

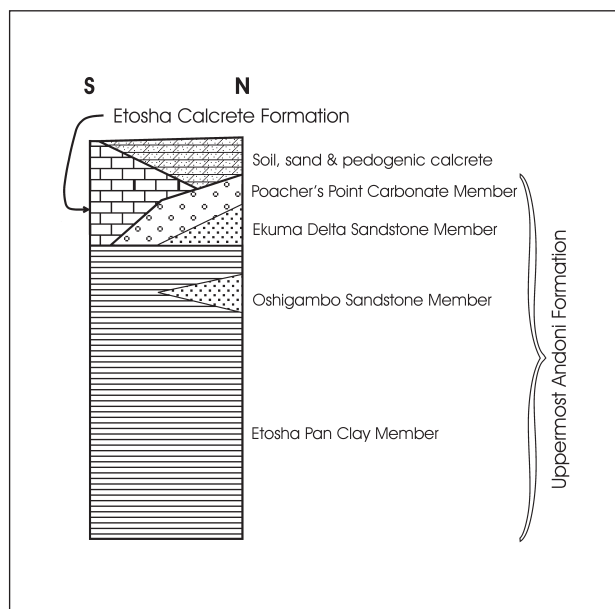


Figure 9. Schematic section of the top of the Andoni Formation in the Etosha pan area. The section shows the relationships of the formation to the Etosha Calcrete Formation and to the younger soil and sand cover which commonly contains pedogenic calcrete.

flooded by dark grey, highly saline mud but the channels that have the strongest flow during flooding commonly have a thin bed of fine-grained sand. Groundwater below the channels is highly saline. The sands above the calcrete are characteristically highly silty, less saline and support millet, sorghum and vegetable fields (Mendelsohn et al., 2000). The central areas of the bevelled high ground between the *iishana* are only 1 to 3 m above the floors of the channels and generally have a thin cover of white aeolian sheet sand. In these regions, the depth to the calcrete varies between 30 cm and 4 m. The northern half of the Namibian part of the system supports open to rather dense stands of woody vegetation, mainly *Colophospermum mopane*, but the southern, more saline part of the system is characterised by extensive, treeless grassy plains with a low population density.

The Kalahari Group in the Owambo Basin

The Kalahari Group fill of the Owambo Basin is made up, from the base upwards, of the siliciclastic Ombalantu, Beiseb, Olukonda and Andoni Formations (Table 2). The thickest intersection of Kalahari sediments in the basin was 471 m in the Olukonda borehole (number 9124, Figure 2). Coeval with and interfingering with all the above Kalahari siliciclastic units is a gigantic, basin-margin groundwater calcrete, the Etosha Calcrete Formation (Figure 2, Table 2). The stratigraphic units in the Etosha Pan area are the uppermost parts of the Andoni Formation and the Etosha Calcrete Formation together with younger soils, pedogenic calcretes and aeolian sands. Only these are of relevance to this paper.

Andoni Formation – the uppermost members in the Etosha Pan area

These are the Etosha Pan Clay Member, the Oshigambo Sandstone Member, the Ekuma Delta Sandstone Member and the Poacher's Point Carbonate Member (Table 2, Figure 9). Good exposures occur within and along the margins of Etosha Pan.

Etosha Pan Clay Member

This unit has been studied by Rust (1984; 1985), Buch et al. (1992), Buch and Rose (1996), Buch and Trippner (1997), Hipondoka (2005) and Miller (2008). It has been named the Etosha Pan Clay Member by Miller (2008). Averaging 50 m in thickness (Miller, 2008), it is the oldest of the four named members and is the main stratigraphic unit that formed the bed of the Palaeolake Etosha. It now forms the present floor of Etosha Pan but this is several metres below the original top of the member. In scarp exposures, the top is at 1084.2 m at Pelican Island and 1086 m on the eastern side of Poacher's Point. At the western margin of the pan, the top occurs at an elevation of 1085 m. Lithological logs of water boreholes show that the distribution of the clay beyond and beneath the present margins of the pan is limited (Figure 3).

Although generally pale olive green in colour, parts of the clay section in some pan-edge boreholes are either grey or pale brown. Between one and four sand layers of varying thickness are interbedded with the clay.

The only preserved borehole samples through the Etosha Pan Clay Member are from a borehole approximately 400 m east of the western edge of the Stinkwater Peninsula (200746, van Vuuren, 2008; Figures 2, 3 and 10). Below a 1 m sand cover, this encountered 5 m of the Etosha Calcrete Formation. The lower 4 m of this calcrete contained scattered oolites weathered out of the underlying 1 m of the Poacher's Point Carbonate Member. The latter was, in turn, underlain by two sections of pale olive green clay of the Etosha Pan Clay Member, respectively 9 m and 12 m thick, separated by 36 m of a highly permeable, pale green, clayey sand aquifer patchily cemented by nodules of calcrete and dolocrete (Figure 10). This borehole is located beyond the present limits of the floor of the pan and the clays will, therefore, not have been affected by recent processes that have modified the upper clays within the pan.

The results of analysis of samples from borehole 200746 are given in Figure 10 (see Appendix for methodology). All samples were saline with NaCl contents varying from 1.25 and 8.74%. KCl contents ranged from 0.17 to 0.71%. The sand and silt fractions consisted entirely of quartz. An occasional chert grain was present. No detrital K-feldspar was detected. This accords with the observation by Giresse (2005) that the upper Kalahari of the Congo Basin lacks detrital feldspar. The slow-settling, fine-grained fractions consisted of varying proportions of authigenic analcime

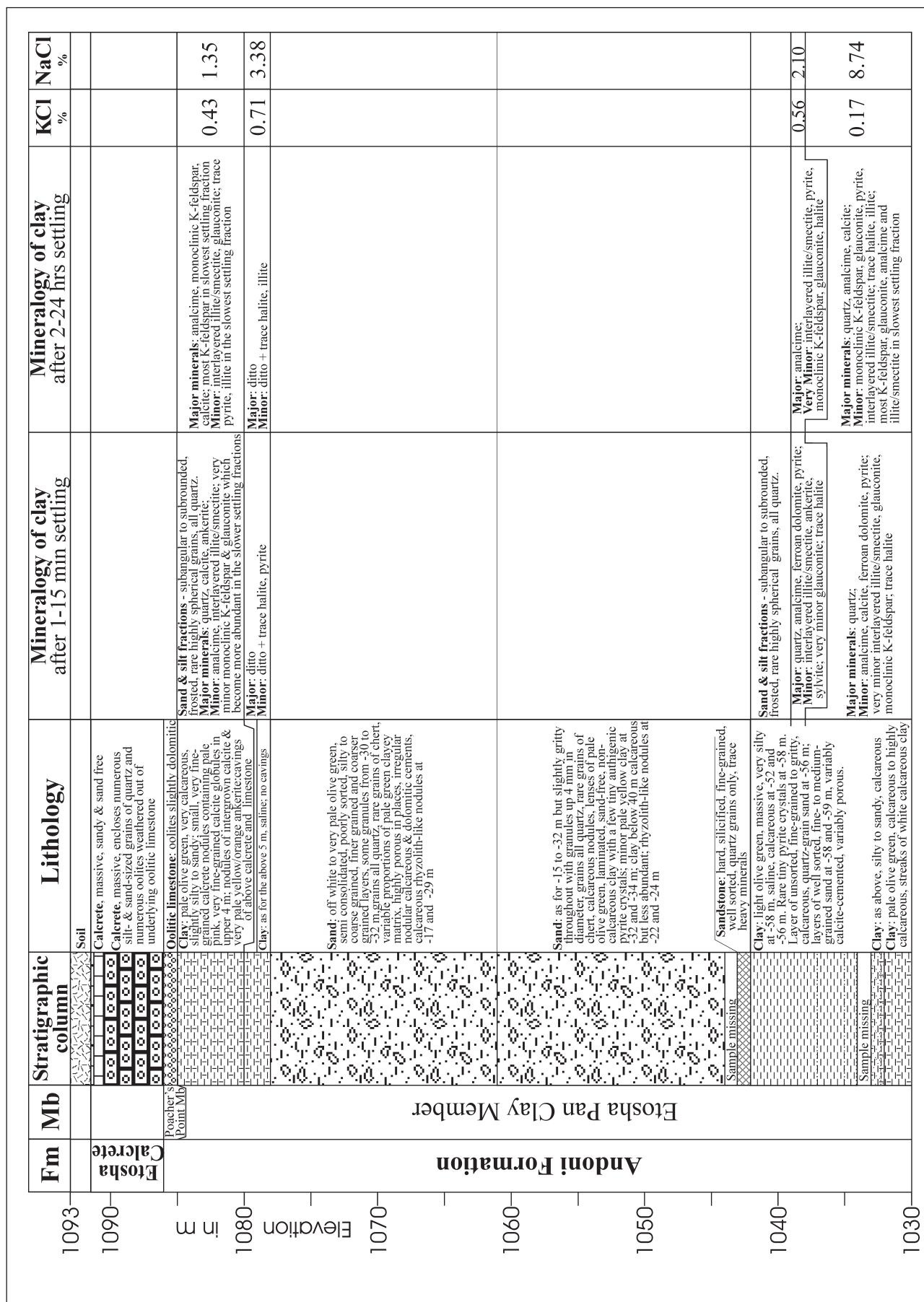


Figure 10. Geological log of borehole 200746 together with the mineralogy of the fine-grained fraction of the clays.

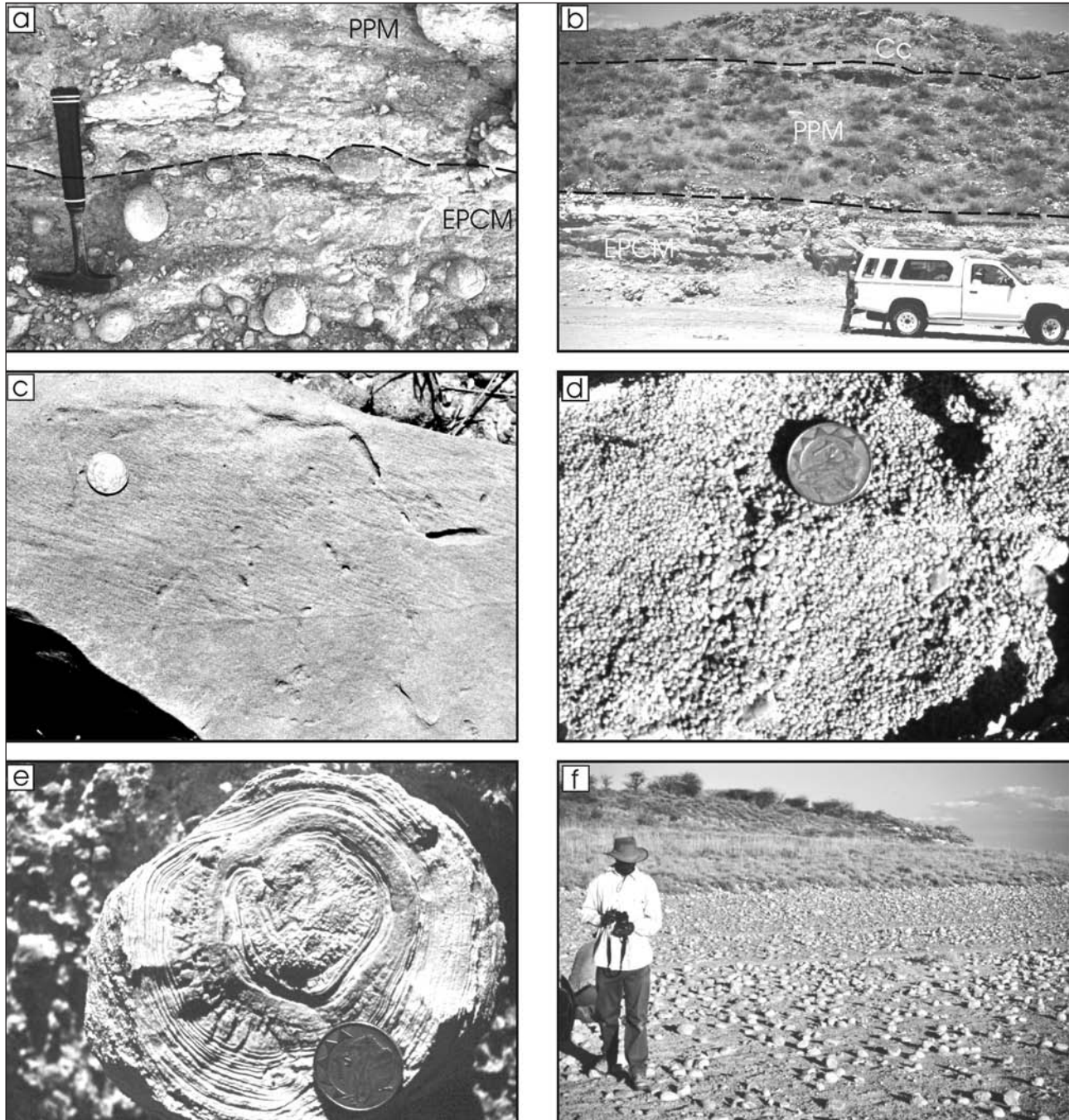


Figure 11. (a) The uppermost 50 cm of the Etosha Pan Clay Member where it lies directly under the oolitic carbonate of the Poacher's Point Member at the southern tip of Pelican Island. Note the scattered oncolites, many of which are broken, in the clay. The oncolites have been derived from coeval, pan-margin exposures of the Poacher's Point Member; (b) Exposure of the Etosha Pan Clay Member, the Poacher's Point Member and a capping of pedogenic calcrete forming the western flank of the Poacher's Point Peninsula. The top of the clay lacks oncolites and oolites entirely at this locality; (c) Cross-bedded layer in white-weathering, highly silicified, Ekuma Delta Sandstone Member near the termination of the Ekuma River; diameter of coin 22 mm; (d) Oolitic Poacher's Point Member, Pelican Island; (e) Several growth zones in an oncolite from the Poacher's Point Member, eastern edge of Etosha Pan; (f) Unusual concentration of oncolites lying on the pan floor; these have weathered out of exposures of the Poacher's Point Member along the eastern edge of the pan. Modified after Miller (2008).

and monoclinic K-feldspar (sanidine?), interlayered illite/smectite, glauconite and, in places, glomeroporphyroblastic clusters of tiny pyrite crystals. Traces of halite and sylvite crystallised out during air drying of the samples. In most samples, the greatest concentrations of the authigenic analcime and K-feldspar occurred in the

finest grained samples collected after 24 hours of settling. The sample from the depth of 55 m contained angular green and white fragments of consolidated layers of very fine-grained analcime, the green with minor very fine-grained glauconite, the white with minor, very fine-grained, monoclinic K-feldspar. Sections

of the clay contained small, irregularly shaped nodules of microcrystalline, calcrite-like calcite, intergrown calcite and very pale yellow ankerite with crystals up to 1 mm in size, and ferroan dolomite. Between the depths of 10 and 15 m, the calcrite-like nodules contained scattered quartz grains as well as distinct, spherical to ovoid globules up to 5 mm in diameter of very pale pink, quartz-free calcite. These globules were massive and structureless and appear to have formed authigenically in the host calcrite.

The uppermost 3 m of the pan present floor are saline and have been variously weathered, altered and redeposited (Buch and Rose, 1996). NaCl contents vary between 2 and 13.5% with the highest values in the deepest parts. White halite encrustations are rare on the floor of the pan but are rather common on the gentle slope that is marginal to and rises away from the pan floor (Jaeger, 1926; Buch and Rose, 1996). The pH is frequently between 9 and 10.5 (Buch and Rose, 1996). Pale olive green clays and green claystones of this upper 3-m section consist of 85% clay-sized material, 10% silt and 5% sand-sized grains (Hipondoka, 2005). The mineralogy of the <2 μ fraction of the claystones is largely authigenic analcime, with or without authigenic monoclinic K-feldspar (almost a high sanidine), an unidentified green clay mineral and, locally, sepiolite and loughlinitite (Buch and Rose, 1996). Authigenic K-feldspar is most abundant in the eastern parts of the pan (Buch and Rose, 1996).

Buch and Rose (1996) also report a thin cover of allochthonous sediments derived from outside the pan along the southern and southeastern parts of the pan. These have two distinct mineral associations in the 2 μ fraction, one dominated by Al₂O₃-poor saponite and/or stevensite, the other consisting primarily of calcite and dolomite, probably sourced from the pan-margin springs. Green sandstones reported to form part of this unit along the northwestern margin of the pan (Buch and Rose, 1996) are actually the Oshigambo Sandstone Member. Unusually high concentrations of Sr of up to 3100 ppm in these upper sediments in places and 7000 ppm in Fisher's Pan suggest the presence of strontianite or emmonite. As much as 8000 ppm F is also present locally in this unit (Buch and Rose, 1996).

The top 50 cm of the clay that directly underlies the oolitic and oncolite-bearing carbonate of the Poacher's Point Member at Pelican Island (Figure 11a) are made up almost entirely of green clay layers containing a debris of oolites and round and broken oncolites derived from laterally coeval, pan-margin exposures of the Poacher's Point Member. In the Poacher's Point Peninsula, the top of the Etosha Pan Clay Member in direct contact with the overlying Poacher's Point Member lacks oolites and oncolites entirely (Figure 11b). In this latter area, the clay in the low cliff section above the pan floor (Figure 11b) consists of halite-rich and halite-poor laminae (Smith, 1980; Smith and Mason, 1991).

Oshigambo Sandstone Member

The pale green, fossiliferous, extensively cross-bedded, fluvial Oshigambo Sandstone Member forms intermittent, flat-lying outcrops on the pan floor between 100 m and 1 km into the pan from the southeastern margin of the Oshigambo Peninsula and possibly as much as 5 km into the pan (Figure 3). It is interbedded and interfingers with the Etosha Pan Clay Member and occurs approximately 4 m below the projected top of the clay in the peninsula. Due to the flat bedding, the flat floor of the pan and a thickness of only 10 cm to 30 cm of the sandstone in the individual outcrops, it has been impossible so far to determine whether the outcrops are due to a single thin, slightly undulating layer, whether they represent more than one layer or whether they include channel deposits. Fossils in the sandstone are listed in Table 3.

Ekuma Delta Sandstone Member

This overlies the Etosha Pan Clay Member, probably paraconformably. It is a hard, white sandstone cemented by silcrete and carbonate. It crops out intermittently for a considerable distance along the Ekuma River and also forms a small delta that fans out onto the Etosha Pan at the mouth of the river. The sands are highly fossiliferous (Hipondoka, 2005; Table 3). Some layers are fluviially cross-bedded and grit layers with occasional small pebbles and fossil fragments also occur (Figure 11c). Small, intermittent outcrops of an identical sandstone with the same fossil suite occur along the margins of the Gwashigambo River and along the western edge of the Oshigambo Peninsular. The nature of these latter outcrops suggests that they were deposited in a meandering channel. The present, less sinuous channel of the Gwashigambo River now cuts across these meanders. A smaller delta occurs at the mouth of the Gwashigambo River. This may be of a similar age to the Ekuma sandstone.

Although the Nanzi borehole (no. 9074, Figure 2) is located south of the rivers forming the palaeo Cuvelai system, a continuous core of fine- to medium-grained, yellowish-white sandstone with manganiferous and ferruginous patches was recovered from the borehole between the depths of 5.5 and 12 m (Hugo, 1969; core stored at the Geological Survey in Windhoek). Consolidation or cementing of the Andoni sands to sandstone at this shallow depth is unusual and this sandstone may be equivalent to the Ekuma Delta Sandstone Member.

Poacher's Point Carbonate Member

The oolitic Poacher's Point Carbonate Member (name modifying slightly from Smith and Mason, 1991) is up to 5 m thick and overlies the Ekuma-equivalent sandstones in the Oshigambo Peninsula and the green Etosha Pan clays directly where these sandstone units are absent (Figure 12). This is a shoreface and nearshore, photic zone facies consisting of oomicrite and oosparite (Figures 11d; 12). The member contains

Table 3. Summary of the distribution of the palaeontological remains and age estimates of the various stratigraphic units exposed in the floor and northern flanks of Etosha Pan. Environmental indicators: * - shallow alkaline water; # - fresh water (usually well oxygenated, but Planorbids can survive in stagnant water, and some hippopotami and turtles can tolerate slightly alkaline conditions); \$ - fresh or alkaline, depending on species; & - terrestrial.

Stratigraphic unit	Oshigambo Sandstone and Etosha Pan Members	Ekuma Delta Member	Poacher's Point Carbonate Member	Pedogenic calcrete	Etosha Raised Beach	Loess on the Oshigambo Peninsula
Correlation/Age	Late Miocene ca 6 Ma	Mid Pliocene ca 4 Ma	Late Pliocene - Early Pleistocene	Early to Mid Pleistocene	Late Pleistocene	Latest Pleistocene ca 12,000 yrs
Oncolites*	Algae	X	X	X (reworked)		
<i>Chlorellopsis coloniata</i> \$	Algae		X			
Palm &	Tree	X				
Dicotyledon &	Tree	X				
Ostracoda \$	Ostracod	X	X			
<i>Hodotermes</i> sp. &	Termite			X		X
<i>Succinea striata</i> &	Land snail		X	X		
<i>Xerocerastus burchelli</i> &	Land snail		X	X		X
<i>Achatina dammarensis</i> &	Land snail		X	X		X
<i>Bellamya unicolor</i> #	Freshwater snail	X			X	
<i>Melanoides tuberculata</i> #	Freshwater snail				X	
Planorbidae sp. #	Freshwater snail			X		
<i>Mutela</i> sp. #	Freshwater clam	X	X		X	
? <i>Egeria</i> sp. #	Freshwater clam		?			
<i>Clarias</i> sp. #	Catfish	X	X			
Cyclanorbinae sp. #	Freshwater turtle	X	X			
<i>Pelusios</i> sp. #	Freshwater turtle	X	X			
<i>Crocodylus</i> sp. #	Crocodile	X	X			
<i>Phoenicopterus ruber</i> *	Flamingo	X	X			
<i>Struthio daberasensis</i> &	Ostrich		X			
<i>Struthio camelus</i> &	Ostrich			X	X	X
<i>Propedetes</i> sp. &	Spring hare		X			
<i>Panthera</i> cf <i>leo</i> &	Lion	X				
<i>Loxodonta</i> sp. cf <i>cookie</i> &	Elephant	X	X			
<i>Ceratotherium praecox</i> &	Rhinoceros		X			
<i>Hipparion</i> sp. &	Zebra		X			
<i>Equus quagga</i> &	Quagga					X
<i>Hippopotamus</i> sp. #	Hippopotamus		X			
Suidae (large sp.) &	Wild pig	X				
<i>Damalacra acalla</i> &	Hartebeest	X				
<i>Redunca darti</i> &	Reedbuck		X			
<i>Damaliscus lunatus</i> &	Tsessebe					X
<i>Antidorcas marsupialis</i> &	Springbuck					X
<i>Aepyceros melampus</i> &	Impala					X
<i>Tragelaphus spekei</i> &	Sitatunga					X
<i>Taurotragus oryx</i> &	Eland					X
Bovidae 2-3 spp. &	Antelopes	X	X			

laminated oolites, elongate silica concretions up to 0.5 m long, and laminar, columnar and oncolid stromatolites which have complex internal growth structures, replacements and overgrowths (Rust, 1985; Smith and Mason, 1991; Martin and Wilczewski, 1972; Figures 11 c to f). Oncolites often show evidence of abrasion and fragmentation, possibly the result of short-distance transport during storms, and/or overturning and

dissolution during periods of partial desiccation and shoreline retreat, followed by periods of renewed growth of laminated carbonate around the oncolites. Up to six such growth zones occur in some oncolites. Locally, oncolites grow upwards from millimetre-thin travertine layers within the oolite. A marker layer of scattered, black, slightly zoned, manganiferous concretions up to about 20 cm in diameter occurs just

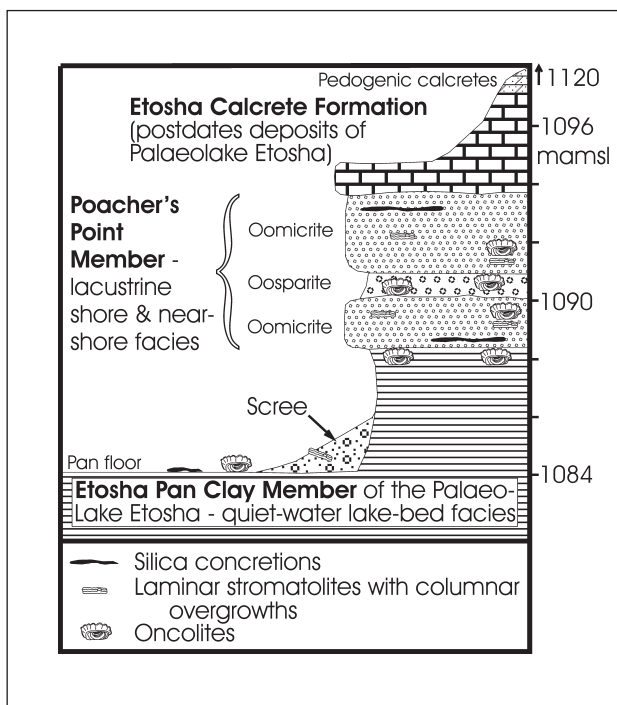


Figure 12. Illustrative section of exposures on Pelican Island showing the stratigraphic relationships between the Etosha Pan Clay Member and the Poacher's Point Carbonate Member (oolitic limestones) of the Palaeolake Etosha. These members form the uppermost units of the Andoni Formation in the island. Concretions, stromatolites and oncolites not to scale; thickness of Etosha Calcrete Formation not to scale and thicker than shown (after Smith and Mason, 1991; Martin and Wilczewski, 1972; modified from Miller, 2008).

above a pisolite-bearing zone near the top of the member on Pelican Island.

The oolites consist of slightly dolomitic calcite. One or more micron-thick layers of silica form laminae in many of the oolites.

The distribution of the Poacher's Point Carbonate Member has not yet been fully mapped out but it is significant that it does not occur along the southern margin of the pan.

Etosha Calcrete Formation

Although initially named the Etosha Limestone Formation by Buch et al. (1992), Miller (2008) considered this to be a gigantic groundwater calcrete and renamed it the Etosha Calcrete Formation (for calcrete read calcrete and dolocrete). Extensive outcrops of the top of the formation form an 80-km wide border to the Otavi Group carbonates that rim of the Owambo Basin (Figure 2). It is coeval with and is the lateral equivalent of all four Kalahari siliciclastic formations in the Owambo Basin. It bears a complex, polycyclical relationship to these siliciclastic formations which it cements, interfingers with and eventually overlies. The northern limit of outcrops is located approximately along the Omuramba Owambo east of Fischer's Pan.

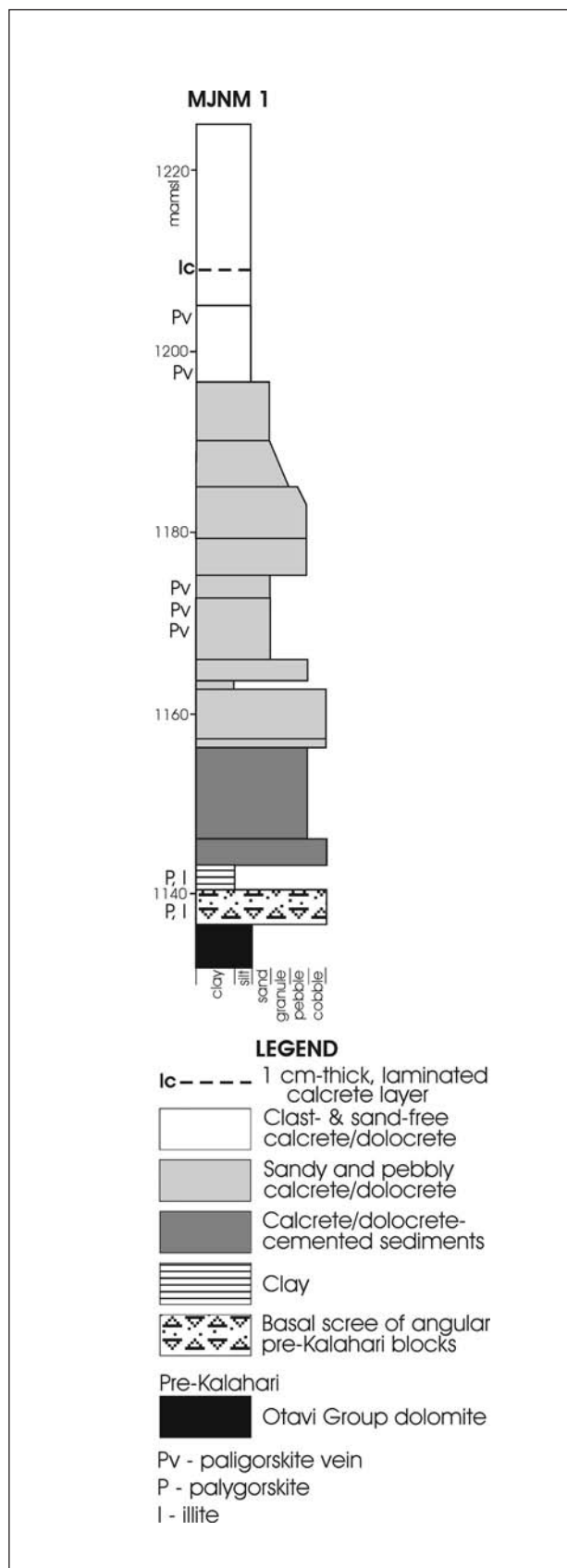


Figure 13. Simplified log of borehole MJNM 1 (JICA, 1997) illustrating the basic stratigraphic units of the Etosha Calcrete Formation. Borehole located on the farm Guinab 277 north-northeast of Grootfontein (simplified after Miller, 2008).

Remnants of the formation indicate that it once almost completely covered the present Etosha Pan (Figure 3). Its deposition began at about 70 Ma during earliest Kalahari times and continues today at the water table (Miller, 2008). Palygorskite with minor sepiolite occurs in the Etosha Calcrete Formation in the Etosha National Park (Buch, 1996) and palygorskite veins were recorded in drill cores of the calcrete north of Grootfontein (Miller, 2008). The Etosha Calcrete Formation is covered by thin sand in places or by thin patches of dark brown to blackish soils.

The formation reaches 120 m in thickness and consists of five distinct stratigraphic subdivisions (Figure 13), namely, from the base up: basal sedimentary breccia, basal clays, red Kalahari siliciclastic sediments cemented by calcrete and dolocrete, sandy and pebbly calcrete and dolocrete, and calcrete and dolocrete lacking siliciclastic inclusions (Miller, 2008). The two basal units are usually thin and discontinuous. Being a carbonate, the calcretes and dolocretes in the formation record several phases of dissolution, karsting and deposition. Dedolomitisation has generated disseminated palygorskite in the calcrete as well as thin, cross-cutting veins up to 1 cm wide of pure, white palygorskite or palygorskite and calcite (Miller, 2008).

Features and lithological units post-dating the Etosha Calcrete Formation

Regional dunes and associated sheet sands

White aeolian sheet sands overlain by pale red longitudinal dunes cover the surface of the Cubango

Megafan. These aeolian sands have been derived from the immediately underlying Kalahari sands (Lancaster, 1981; 1987; Thomas, 1986). The sheet sands choke the most of the palaeo river courses of the megafan. Consequently, these rivers rarely flow. The white sands are extensively cemented by pedogenic calcrete in the southern part of the megafan and support a dense but high shrubby vegetation with scattered tall trees such as the marula (*Sclerocarya birrea*). The sands are deep and less obviously calcretised towards the Angolan border and are covered by dense, taller vegetation that includes kiaat/mukwa (*Pterocarpus angolensis*). In this region, it is common for water from heavy downpours to drain away completely into the sand within hours (Stengel, 1963; H. Zauter, personal communication, 2009).

The east-west longitudinal dunes occur over the full length of the Cubango Megafan. They reach heights of approximately 6 m in the southern part of the fan (Figure 5) but they decrease progressively in size northwards and are less than 0.5 m in height near the Angolan border. In the latter region, groups of three to four dunes are separated by several kilometres of dune-free white sand. The longitudinal dunes die out westwards and the last remnants reach the Gwashigambo River but do not extend west thereof into the Cuvélai system.

A few, rare, pale red sand dunes in the form of isolated barchans or broad hummocks occur on the Cuvélai system between the iishana. These are extensively excavated for building sand and brick making.

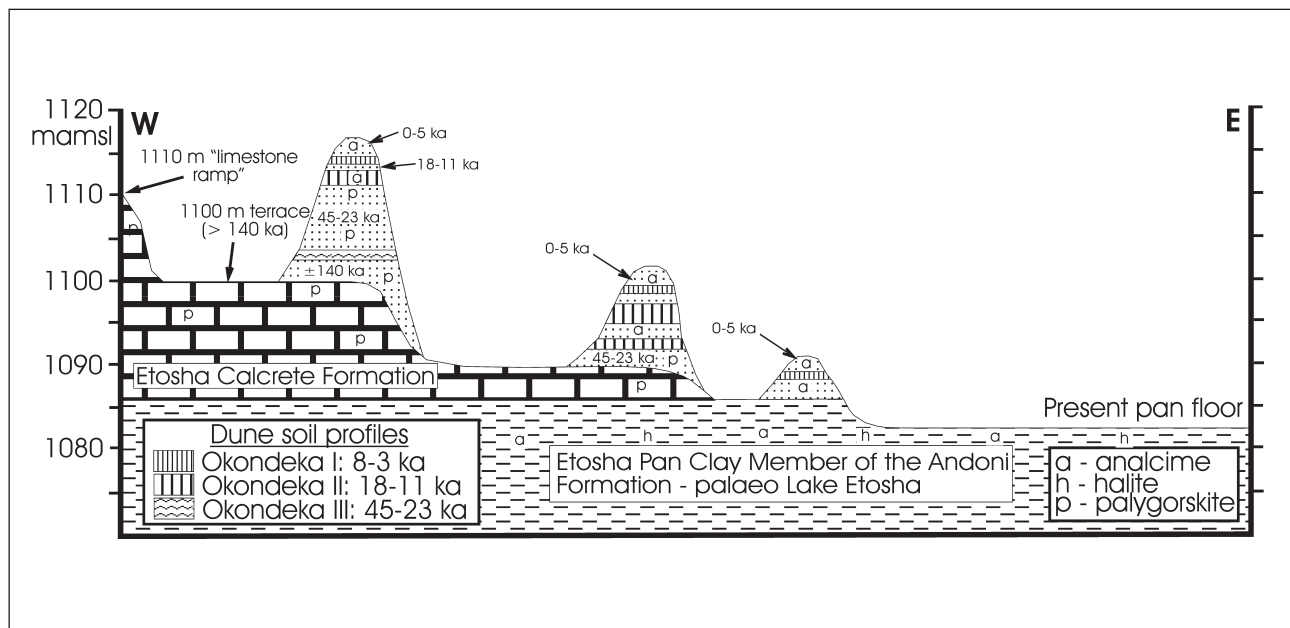


Figure 14. Schematic eastwest section across the western margin of the Etosha Pan in the Okondeka area showing (i) the 1100, 1090 and 1085 m terraces developed in the Etosha Calcrete Formation, (ii) the lunette dunes developed on the inner edge of the terraces and the internal structure of the dunes with TL ages of the aeolian sands and, (iii) the pedogenic horizons Okondeka III to I (modified after Buch, 1996; Buch and Rose, 1996; Buch et al., 1992; Miller, 2008).

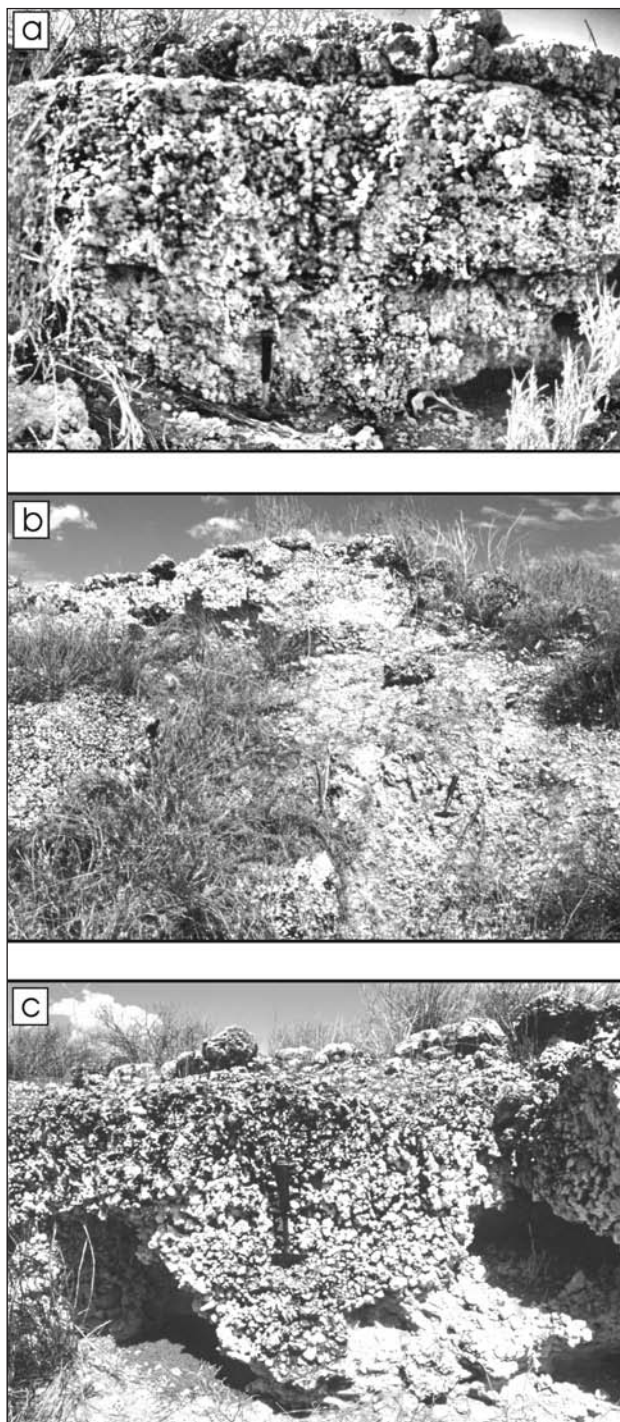


Figure 15. (a) Pedogenic calcrete on the northern bank of the Ekuma River consisting almost entirely of the nodular zone with very little fusing of the nodules towards the top; (b) Two pedogenic calcretes cover the Poacher's Point Peninsula. The photograph shows the upper calcrete which is 4 m thick. The hammer is centred on the thick nodular zone. The honeycomb zone is well developed, more weather resistant and forms the steep upper faces. A thin hardpan forms an overhanging cap; (c) The weather-resistant honeycomb zone and the thin hard pan cap of photograph 'b' (modified from Miller, 2008).

Terraces marginal to the pan

Pan-margin terraces occur at elevations of 1108 to 1110, 1097 to 1100, 1092, 1086 and 1082 m, the highest furthest from the pan and the lowest along the pan edge (Jaeger, 1926; Buch, 1996; 1997; Buch et al., 1992). Those at 1100, 1092 and 1086 m are the best developed and occur in the Etosha Calcrete Formation. The top of the "limestone ramp" of Buch (1996; 1997) and Buch et al. (1992) is at an elevation of 1110 m. However, they suggest that the 1130 m contour east, south and west of the pan marks the initial limits of Etosha Pan at approximately 2 Ma (Figures 2 and 3).

Lunette dunes, pan-margin dunes with associated loess, and dunes just west of Ruacana town

Aeolian sands blanket the elevated east-facing shorelines of the western and northern parts of the pan (Figure 3). Lunette dunes with internal pedogenic horizons at different levels (Figure 14) rest on the inside edges of the 1100, 1090 and 1085 m terraces along the western margin of the pan (Buch, 1996; 1997; Buch et al., 1992). Sand at the base of the dune on the 1100 m terrace has a poorly constrained TL age of 140 ka whereas that at the base of the dune on the 1085 m terrace is only about 10 ka. Pedogenic horizons, representing breaks in aeolian deposition, formed in the intervals 45 to 23 ka (Okondeka III), about 18 to possibly 11 ka (Okondeka II), and about 8 to 3 ka (Okondeka I) (Buch et al., 1992). Younger sands overlie the Okondeka I horizon. Only sands and pedogenic profiles younger than the Okondeka III horizon occur on the 1090 m terrace and only sands younger than the Okondeka II horizon occur on the 1085 m terrace (Figure 14). Palygorskite is abundant in the clay fraction below the Okondeka II horizon but minor at higher levels. Analcime occurs in the Okondeka II horizon and at higher levels but is absent from older horizons (Buch, 1996; 1997; Buch et al., 1992).

Low, west-facing dunes with wavelengths of 1.5 to 2 km, amplitudes 3 to 5 m and somewhat sinuous, north-northeasterly trending dune crests form the northernmost margin of the pan outside the boundary fence of the Etosha National Park. On the pan itself, only the thickest parts of these dunes, i.e. the dune crests, are preserved (Figure 3). Beyond the limits of the pan, the dunes have long, gentle stoss slopes 0.5 to 1 km in width and abrupt lee slopes only a few metres wide. The dunes are composed of pale grey, weakly cemented to unconsolidated, highly silty sands. An abundant grass cover traps considerable volumes of pale grey, silt-rich loess and it is not always easy to distinguish dune material from loess. The interdune areas are flooded during local heavy rains and have become flooded by dark grey, sandy clays. The grasses on these clays are different from those on the dunes and help to delineate which is which.

Low, west-facing dunes with similar wavelengths, amplitudes, stoss and lee faces and north-northeasterly trends occur over a distance of 4 to 5 km just west of the

town of Ruacana. These consist of very pale red sands derived from the underlying pink Nosib Group arkoses.

Pedogenic calcretes

Pedogenic calcretes have formed in places in soil profiles that cap Andoni Formation sands, the Poacher's Point Carbonate Member and the Etosha Calcrete Formation. The pedogenic calcretes are between 1 and 4 m thick and are best distinguished from the massive Etosha Calcrete Formation by being largely nodular (Figure 15). The honeycomb zone can be well developed but this is not always the case. The soil profile that covered the oncolite-bearing oolite layers of the Poacher's Point Member at the northern end of the Stinkwater Peninsula contained fragments of the oolite and weathered-out oncolites. These remnants of the Poacher's Point Member are now enclosed in the pedogenic calcrete that caps this peninsula.

Soft, olive green, slaked clay deposits

During heavy flooding of the Cuvelai system, the Ekuma River flows strongly into the Etosha Pan and can, on occasion, fill it completely. A deposit of dry, slaked, olive green clay so soft that one can push one's finger deep into it occurs almost 10 m above the floor of the pan on the margins of small gulleys on eastern edge of Pelican Island (Figure 16a). These deposits are so soft that they are preserved only in the wind shadow of the gulleys. Repeated drying out and wetting by rain has resulted in extensive slaking of the deposits and development of their soft, highly porous nature. These deposits may be more extensive than we have found.

Raised beaches and pan-margin loess

Thin raised beaches with fossils have been identified between one and two metres above the pan floor on the gentle eastern slopes of the Oshigambo and Poacher's Point Peninsulas. Above these, grass has trapped

loess of silt and sand blown by easterly winds. This loess, overlying the dunes in Figure 3, is more extensive than the raised beach deposits.

Magadi-type chert

Scattered, irregularly shaped, knobby nodules between 0.5 and 15 cm in size of white to pale grey, translucent magadi-type chert occur on the pan surface in patches up to 50 m across between the eastern pan margin and Pelican Island and in lower concentrations immediately south of the Oshigambo Peninsula (Figure 16b). The nodules have a soft, opaque white, siliceous coating. Based on the limited spread of the 2009 flood waters in the pan, that area in which one might find such chert nodules on the pan floor is outlined in Figure 3. Given the saline nature of much of the Etosha Pan Clay Member, such nodules could also be present deeper down in this clay succession.

Palaeontology and ages of the fossils in the Etosha Pan Clay, Oshigambo Sandstone and Ekuma Delta Sandstone Members and the more recent deposits

The fauna from the Oshigambo Sandstone Member and the Etosha Pan Clay Member (Table 3) is close in age to the Langebaanweg site, South Africa (Hendey, 1976). Originally correlated to the Pliocene, the Langebaanweg fauna is more likely latest Miocene in age (ca 6 Ma) on the basis of biochronological correlations to East African sites of this age (Lukeino Formation, Kenya, ca 6 Ma: Sawada et al., 2002). Particular resemblances between the Oshigambo Sandstone Member and the Etosha Pan Clay Member fossils and those from Langebaanweg are the extinct hartebeest *Damalacra acalla* (Figure 17a) and the primitive loxodont elephant *Loxodonta cookei*.

The Ekuma Delta Sandstone Member is middle Pliocene on the basis of the bovid *Redunca darti*, typically present in sites in southern and eastern Africa aged ca 4 Ma (Makapansgat, South Africa; Kanam

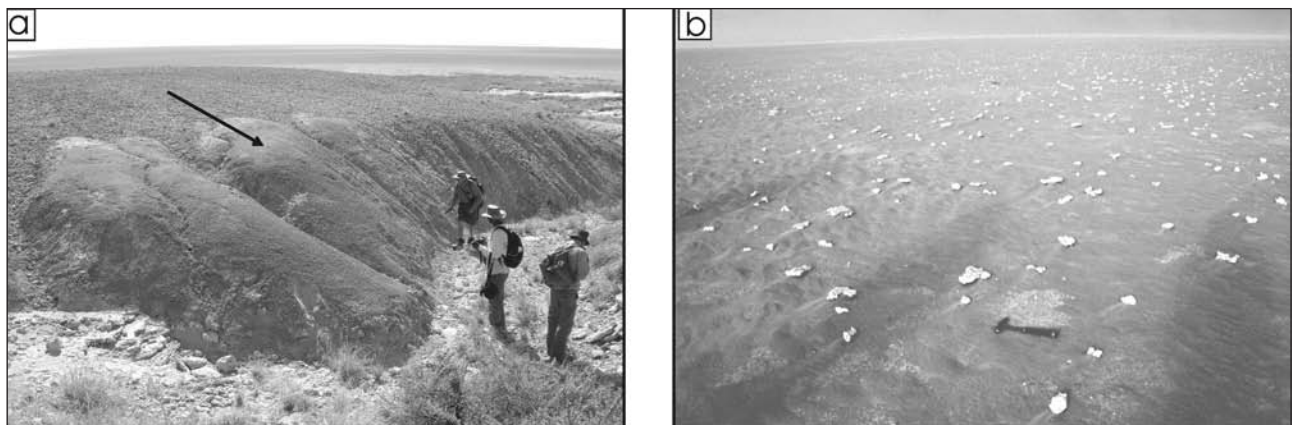


Figure 16. (a) Isolated deposit (arrowed) of soft, green, slaked clay on the flank of a narrow gully on the eastern edge of Pelican Island; (b) White, irregularly shaped, Magadi-type chert nodules scattered on the floor of Etosha Pan between Pelican Island and the eastern margin of the pan (modified after Miller, 2008).

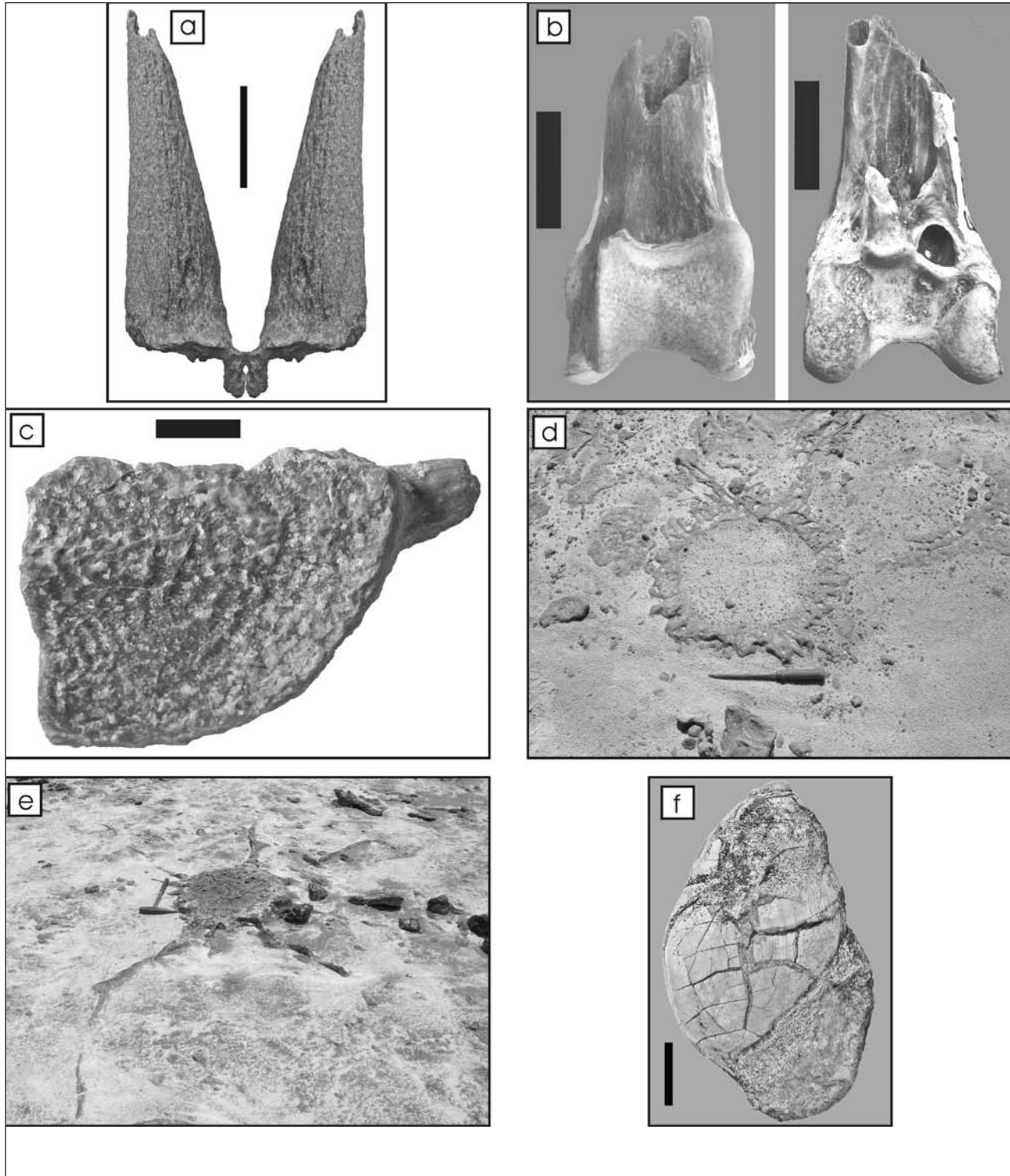


Figure 17. (a) *Damalaca acalla* horn cores from the Etosha Pan Clay Member. This is a primitive alcelaphine antelope found primarily at Langebaanweg, South Africa, but also in Kenya in Late Miocene deposits (scale bar 10 cm); (b) *Flamingo tibiotarsus*, anterior view on left, posterior view on right. Flamingo leg bones from the Etosha Pan Clay Member, aged ca 6 Ma, indicate the likely presence of alkaline surface waters in the area (scale bar 10 mm); (c) *Cyclanorbis* scute from the Etosha Pan Clay Member, ca 6 Ma. Freshwater turtle scutes are common in the Etosha Pan Clay Member and the Ekuma Delta Sandstone Member, and indicate the presence of fresh or only weakly alkaline surface waters (scale bar 10 mm); (d) Palm tree bole in the Ekuma Delta Sandstone Member. Note the abundant small roots surrounding the trunk (pen for scale); (e) Dicotyledon tree bole in the Ekuma Delta Sandstone Member. Note the few large diameter roots radiating away from the bole (hammer for scale). There are many similar occurrences in the Ekuma Delta Sandstone Member which indicate that groundwater was fresh at the time that they grew; (f) *Achatina* from the Poacher's Point Carbonate Member. *Achatina* is a land snail common in steppe and savannah (scale bar 10 mm).

Museum Cliff, Kenya). The extinct spring hare from Ekuma Delta Member is comparable to that from Laetoli, Tanzania (ca 3.7 Ma) and the ostrich eggs from this level belong to *Struthio daberasensis*, a middle Pliocene species, known from Namibia and East Africa (Pickford et al., 1995). The large white rhino from the Ekuma Delta Member is the extinct species *Ceratotherium praecox* which ranges in age from ca 7 to ca 4 Ma.

The age of the Poacher's Point Carbonate Member is not well constrained by the fossils, but the presence of eggshells of the extant species *Struthio camelus* indicates an age younger than mid-Pliocene.

The pedogenic calcrete on the northern edge of the pan is correlated to the Pleistocene as it underlies sediments with a latest Pleistocene age at the Oshigambo Peninsula, and it overlies the Poacher's Point Carbonate Member. The fossil eggshells found in this deposit (*Struthio camelus*) are concordant with such a correlation.

Raised beach deposits which underlie loess on the Oshigambo Peninsula have yielded remains of freshwater molluscs, *Melanoides tuberculata*, *Bellamyia unicolor* and *Mutela* sp.

The loess on the Oshigambo Peninsula contains a diverse and rich extant fauna which has been dated by the ^{14}C method to ca 12 Ka.

The palaeoenvironment of the Etosha succession varied through time. Early stages (Etosha Pan Clay Member, Ekuma Delta Member) were dominated by freshwater deposition, but with a hint of salinity (flamingos in the Etosha Pan Clay Member, Figure 17b), becoming saline/alkaline during the late Pliocene and Pleistocene (oncolites, magadi-type chert). The freshwater molluscs from raised beach deposits that underlie the loess on the Oshigambo Peninsula, indicate a palaeoclimate that is somewhat more humid than that of today. The fauna in the loess on the Oshigambo Peninsula suggest a short-lived period of paludal or lacustrine conditions at the end of the Pleistocene.

Discussion: the occurrences of monoclinic authigenic K-feldspar and magadi-type chert

Monoclinic authigenic K-feldspar

The uppermost layers of the Etosha Pan Clay Member contain abundant authigenic analcime and monoclinic K-feldspar (Figure 10). McCarthy et al. (1991) also record the presence of authigenic K-feldspar in the saline sediments of the islands of the Okavango Delta. Buch and Rose (1996) were the first to record the presence of abundant, very fine-grained, authigenic K-feldspar in the clay of the Etosha Pan Clay Member. Their structural analysis of the (060) and ($\bar{2}04$) reflections (Wright, 1968) showed it to fall in the sanidine field, i.e. monoclinic K-feldspar. In the present study, all samples fell within or very close to the field delineated by Buch and Rose (1996).

Such authigenic K-feldspar is referred to as monoclinic K-feldspar in some publications (Hearn et al.,

1982; Buch and Rose, 1996) and sanidine or low sanidine in others (Hansley, 1982; Ali and Turner, 1982; Stamatakis, 1989). It is recorded as overgrowths in tuffs (Sheppard and Gude, 1968; 1973; Surdam and Parker, 1972), on detrital feldspar in arkosic sandstones (Kastner, 1971; Desborough, 1975; Ali and Turner, 1982; Hearn et al., 1982; Bjørlykke, 1984; Kaiser, 1984; Helmold and van de Kamp, 1984; Loucks et al., 1984; Siebert et al., 1984; Surdam et al., 1984) or as a primary authigenic mineral in sedimentary rocks lacking detrital feldspar (Kastner, 1971; Hearn et al., 1982; Buch and Rose, 1996). Such overgrowths are in optical continuity with detrital host grains of volcanically sourced sanidine (Hansley, 1982; Ali and Turner, 1982; Stamatakis, 1989) but are optically discontinuous with grains of triclinic K-feldspar (Ali and Turner, 1982; Bjørlykke, 1984). The composition of such authigenic K-feldspar is commonly $\text{Ab} < 1\text{An}0.0\text{Or} > 99$, i.e. almost pure K-feldspar (e.g. Helmold and van de Kamp, 1984). Distinctive for such authigenic monoclinic K-feldspar are low contents of Rb and Ba relative to igneous sanidine (Hearn et al., 1982). Saline conditions appear to be essential for it to form (Hardie, 1968; Bradley and Eugster, 1969; Buyce and Friedman, 1975; Eugster and Hardie, 1978; Smith, 1979; Stamatakis, 1989).

The rate of formation of the authigenic analcime and K-feldspar in the Etosha Pan Clay Member must have been relatively fast and dependant on only a very limited amount of burial. The maximum thickness of cover above the top of the Etosha Pan Clay Member was about 41.5 m (5.5 m of the Poacher's Point oolite and 36 m of the Etosha Calcrete Formation). The age of the Ekuma Delta Sandstone Member is 4 ± 1 Ma. Thus, the analcime and K-feldspar in the top 5 m of the Etosha Pan Clay Member formed within 4 my.

Magadi-type chert

Chert nodules that occur on the floor of saline lakes are considered by Eugster (1967, 1969), Hay (1968), Jones et al. (1967), Sheppard and Gude (1982, 1986) and (Carozzi, 1993) to have been derived from the precursors magadiite ($\text{NaSi}_7\text{O}_{13}(\text{OH})_3 \cdot 3\text{H}_2\text{O}$) and/or kenyaite ($\text{NaSi}_{11}\text{O}_{20.5}(\text{OH})_4 \cdot 3\text{H}_2\text{O}$). Evaporitic brines with a pH of 11 or higher, as found in East Africa and elsewhere, can dissolve up to 2700 ppm SiO_2 (Jones et al., 1967). Lowering of the pH by the influx of fresh water, by percolation of biogenically produced CO_2 or by simple evaporative concentration of the brine (such as in animal footprints) results in almost instantaneous precipitation of soft, pliable magadiite (Eugster, 1967, 1969; Hay, 1968). Dehydration and the loss of sodium lead to relatively rapid conversion of magadiite to kenyaite and thence to chert (Sheppard and Gude, 1982, 1986; Carozzi, 1993). The scattered distribution of the nodules on the floor of the pan suggests that the Etosha magadi-type chert nodules could be the result of the present-day concentration and final evaporation of brines in animal footprints on the pan floor.

Interpretation: evolution from Palaeolake Etosha to Etosha Pan

The shear size of the Cubango Megafan, approximately 350 km long and at least 300 km wide, relative to the other fans indicates that it was the main contributor to the sediment fill of the Owambo Basin, at least during Andoni times. It is confined to the Owambo Basin and its symmetrical shape could only have been produced by a palaeo Cubango/Okavango river system depositing sediment lobes down the fan axis and almost symmetrically on either side of it. Deposition took place from north to south with the topographically lowest region in the Owambo Basin, the Palaeolake Etosha, being the end point of the fan. The 50 m-thick Etosha Pan Clay Member represents the end-point fines of this system. Very fine-grained, authigenic analcime, monoclinic K-feldspar (sanidine) and glauconite are abundant in the clays. With slopes of 0.00056 over a distance of the 95 km in the proximal parts of the fan and of 0.00028 over the final 255 km, the Cubango Megafan falls into the category of low-inclination braided fluvial fans of Stanistreet and McCarthy (1993).

The distal clay-rich regions of the Kunene Megafan suggest that Palaeolake Etosha may also have been its end point (Figures 2 and 4). However, because of the significantly smaller size of this megafan, the proportion of clay that it contributed to the Etosha Pan Clay Member was probably much less than that contributed by the Cubango Megafan. Thus, the influence of the palaeo Kunene River does not appear to have had as much of an influence on flow into and deposition in Etosha Pan (i.e. Palaeolake Etosha) as previously suggested by Wellington (1938).

Clay accumulation in Palaeolake Etosha was intermittent with clay deposition periodically interrupted by accumulation of sands and periods of non deposition. Flooding was seasonal and often limited in volume permitting regular and complete desiccation. Varying halite contents through the clay succession and minor sylvite in places suggest longer and shorter periods of regular flooding and desiccation. Palaeolake Etosha was apparently not a permanent body of fresh water although it may have held significant depths of fresh water at times after particularly heavy flooding.

In the Beiseb Pan borehole, number 9296 (Figure 2), which is only about 6 km away from the nearest suboutcrop of the Etosha Pan Clay Member, remnants of shells of the extinct ostrich-like bird *Namornis osbanai* were recovered from a depth of 37 m below surface, i.e. at an elevation of 1075 m. Senut and Pickford (1995) and Pickford and Senut (2000) ascribe a Miocene age bracket of 16 and 15 Ma to such shells. The last 13 m of the Etosha Pan Clay Member between the elevations of 1075 m and the top of the member at 1088 m (Figure 12) were deposited over a period of about 12 Ma since fossils in the Ekuma Delta Sandstone Member, which is the capping to the clay, have an age of

4 ± 1 Ma. This suggests that the base of the 50-m thick Etosha Pan Clay Member may have an Oligocene age.

Sedimentation in the Cubango and Kunene Megafans and Palaeolake Etosha coincided with a period of oceanic warming and wetter conditions onshore during the Oligocene and Early Miocene (Tyson and Partridge, 2000; Haddon and McCarthy, 2005). Floral and faunal remains attest to wetter conditions between 19 Ma and 17 Ma (Bamford, 2000; Vrba, 2000). High onshore-derived organic matter in offshore cores of equivalent age also reflect this wetter period (Udeze and Obok-Ikuenobe, 2005). However, the lower aeolianites of the Namib-wide Tsondab Sandstone Formation (Ward, 1987 a and b) were being deposited at the same time (Senut and Pickford, 1995; Pickford and Senut, 2000) indicating concomitant arid conditions along the western continental margin.

Much of the Kunene Megafan is obscured by the younger and scarcely aggradational Cuvelai Drainage System. It is possible that the early Cuvelai system may have evolved gradually from the Kunene Megafan. The shallow channels of the Cuvelai system are wide and well developed right up to the watershed between the Kunene River valley and the present Cuvelai system. The authors interpret this to indicate that the channels of the Cuvelai system, or at least the westernmost channels, were fed and cut by overflow from the Kunene River as far downstream as the Ruacana Falls. As the Kunene River began to incise its present valley, it supplied progressively less and less water to the Cuvelai drainage. Gradually, the western part of this system became a fossil system (Figure 7) fed only by local rains. The feed for the rest of the Cuvelai system would then have been from the northern Cuvelai River and southwestern parts of the Angolan highlands (Figure 7) very much as it is today. With a slope of 0.00069, the upper 120 km of the Cuvelai Drainage System can be classified as a low-inclination, braided river fan but the remaining section of 212 km down to the edge of Etosha Pan has a slope of 0.00024 (Figure 6) which is identical to that of the Okavango Fan and falls into the category of losimean fans of Stanistreet and McCarthy (1993). The sub-parallel orientation of the channels in the system and the low sinuosity thereof are typical characteristics of such fans. Nevertheless, the high average width/depth ratio of 448 of the channels in the Cuvelai system and the lack of an obvious downstream change in this ratio differs markedly from the Okavango Fan. Another significant difference is the lack of reed and papyrus beds marginal to the Cuvelai channels downstream of the Evale swamps. Without the filtering effect of such vegetation, the flood waters of the Cuvelai system are muddy all the way down to the mouth of the Ekuma River.

The paucity of curvilinear elevation contours in the Cuvelai system (Figure 4) strongly suggests that it was and is a sediment-starved system. Only between 5 and 10 m of sediment has accumulated in the Cuvelai system at its eastern edge where it overlaps onto the western edge of the Cubango Megafan (Figures 4; 6). The first

evidence of a contribution of the Cuvelai system to the sedimentary pile in Palaeolake Etosha is provided by the pale green Oshigambo Sandstone Member. This is not far from the mouths of the Ekuma and Gwashigambo Rivers. This sandstone was deposited before Palaeolake Etosha dried up as it occurs approximately 4 m below the top of the Etosha Pan Clay Member. Whether or not this was a Cuvelai system that included feed from the Kunene River is unknown. The fossil remains in the sandstone have an age of 6 ± 1 Ma. Thus, the accumulation of sediments in the Kunene Megafan must have ceased somewhat prior to 6 Ma. Green sandstones at the northeastern margin of the pan (Buch and Rose, 1996) may be an equivalent that could be attributed to the Cubango Megafan.

There is evidence for both fresh and saline water. The flamingo remains and the oncolites are indicative of saline conditions. The salinity of the clays (Buch, 1993) points to regular desiccation. In contrast, tree boles in the Oshigambo and Ekuma Delta Sandstone Members and the hippopotamus and turtle remains provide evidence of fresh water at times in the shrinking palaeolake. Figure 10 shows the uppermost clays in borehole 200746 to have been the least saline.

The initiation of the degradation of the Kunene River Valley that eventually cut it off from the Cuvelai system may also have been the stage at which the Okavango River began to cut its present, southeast-trending valley into the Cubango Megafan. This would then also have been the stage at which more and more water was diverted down the incising valleys and away from Palaeolake Etosha. It was the beginning of the end of the palaeolake.

The upper 4 m of the Etosha Pan Clay Member between the Oshigambo and Ekuma sandstones were deposited over a period of approximately 2 my. Deposition of the Etosha Pan Clay Member ceased with the deposition of the Ekuma Delta Sandstone Member and the latter was the last stratigraphic unit to have been deposited by any of the fluvial systems feeding into Palaeolake Etosha. This implies that both the Kunene and Okavango Rivers had already incised their present valleys deeply enough to contain even the highest floods. A further implication is that the Ekuma Delta Sandstone Member was deposited solely by that part of the Cuvelai system that is shown as the presently active system in Figure 7. This stage was reached at approximately 4 Ma. However, incision of the Kunene and Okavango valleys could have been initiated well before this and possibly as far back as 6 Ma. This incision appears to mark the end of large-scale, regional aggradational fluvial accumulation of the Kalahari Group in the Owambo Basin and it may also mark the end of regional fluvial accumulation of the Kalahari Group throughout the southern Kalahari Basin as a whole.

By implication, the development of the Okavango Swamps may have started in earnest at about 6 Ma and must have been well established by about 4 Ma.

The sediments deposited subsequently in the Okavango Fan are located in a tectonic depression and overlie possibly as much as 300 m of older sediments sourced from several rivers that terminated in this depression (Haddon and McCarthy, 2005).

The areal extent of Palaeolake Etosha shrank as flow in the Cuvelai system progressively decreased. The oolites of the Poacher's Point Carbonate Member (Figure 12) mark a major change in sediment composition and record the final demise of Palaeolake Etosha. The Poacher's Point Carbonate Member is the uppermost unit of the Andoni Formation. It is chemical and not fluvial in origin and is interpreted as a lake-margin deposit. Its restriction to the northern part of Etosha Pan over an area far smaller than that of the underlying Etosha Pan Clay Member indicates that the areal extent of Palaeolake Etosha had shrunk significantly (Figure 3). Two further lines of evidence support this.

The first is the occurrence of oncolites in layers in the top 50 cm of the green clays of the Etosha Pan Clay Member at the southern end of Pelican Island (Figure 11a). Many of these oncolites are broken. They are not primary to these green clays. They have their origin in the pan-margin Poacher's Point Carbonate Member and were washed into the deepest part of what remained of Palaeolake Etosha where the final deposition of clay, possibly reworked, was taking place. This implies that the top of the Etosha Pan Clay Member is diachronous as is Poacher's Point Member and that accumulation of the final layers of clay in the deeper, central parts of the lake was coeval with oolite deposition at the lake margins. Furthermore, the disrupted layering in the oncolites (Figure 11e) is interpreted as having been caused by periods of exposure during which no growth and even dissolution and fragmentation took place followed by renewed inundation and resumption of algal growth on the oncolites. Up to six periods of such interrupted growth are evident in some oncolites.

The second point is the source of carbonates for the oolites of the Poacher's Point Carbonate Member. The lack of lacustrine carbonate layers in the Etosha Pan Clay Member suggests that the waters of the palaeo Cubango, Kunene and Cuvelai systems had very low carbonate contents. It is also rare indeed to find borehole logs that record pan carbonate layers interbedded with clays in the Andoni Formation. A shrinking Palaeolake Etosha would have permitted the Etosha Calcrete Formation to start advancing northwards over the Etosha Pan Clay Member (there is no record of the Poacher's Point Carbonate Member along the southern margin of Etosha Pan). The source of the water for this calcrete was and still is the karsted dolomites and limestones of the Otavi Group that form the elevated margins of the Owambo Basin (Figure 2). This water is hard and abundant. Today, numerous seepage springs occur at the edge of this calcrete along the southern margin of Etosha Pan (Figure 3). Any water flowing from

the calcrete into Palaeolake Etosha would have brought dissolved carbonate with it. Warm water loses CO₂ and leads to deposition of excess calcium as CaCO₃. Daily solar heating of the shallow waters along the lake margins would have resulted in the accumulation of lake-margin carbonate. In such a shallow environment, oscillatory wave action will have produced the oolites from the primary carbonate deposits.

By the time deposition of the Poacher's Point Carbonate Member ceased, probably during the mid Pliocene, Palaeolake Etosha no longer existed.

Once Palaeolake Etosha dried up completely, the uppermost parts of the palygorskite-bearing Etosha Calcrete Formation were able to advance over the Poacher's Point Carbonate Member almost to the northern margin of the present pan (Figure 3; Miller 2008). Although this formation is coeval with the whole of the Kalahari siliciclastic succession, its deposition outlasted accumulation of the Kalahari siliciclastic succession and continues today below surface at the water table (Miller, 2008). By the time this advance ceased, the palaeolake deposits were almost totally covered by as much as 36 m of the Etosha Calcrete Formation. Along the northern edge of the present pan, sands covered the palaeolake deposits.

The gradual cessation of flow into Palaeolake Etosha implies a major change in climate (Korn and Martin, 1937; 1957), probably due to a large-scale expansion of arid anticyclonic conditions (Lancaster, 1981; 1987). The change must have been rather gradual and may well have begun with the deposition of the aeolian, Namib-wide Tsondab Sandstone Formation in the Namib Desert (Ward, 1987 a and b; Ward et al., 1983). Remnants of egg shells of a large ostrich-like *Aepyornithoid* bird with an age of between 20 Ma and 16 Ma (Pickford et al., 1995; Senut and Pickford, 1995; Pickford and Senut, 2000) occur in basal dune sands of the Tsondab sandstone. The Tsondab aridity may in part be associated with Early Miocene upwelling (Summerhayes et al., 1992). Development of an east-west climatic gradient between ca 14 Ma and ca 10 Ma across Southern Africa (Tyson and Partridge, 2000) was gradually enhanced by strengthening of upwelling during the establishment of the cold Benguela Current (Siesser, 1980; Dingle et al., 1983; Diester-Haass et al., 1992; 2002; Christensen and Girdaueau, 2002; Haddon and McCarthy, 2005). Fossils in karst cavities in northern Namibia support Late Pliocene – Early Pleistocene aridity (Pickford et al., 1994).

Studies of offshore sediments also support the gradual increase in aridity across the subcontinent. Middle Miocene sediments off the coast of South Africa contain the highest proportion of land-derived organic matter ascribed to a subtropical climate (Udeze Oboh-Ikuenobe, 2005). Thereafter, a significantly reduced sediment supply and much less land-derived organic matter suggest a cool, temperate climate (Udeze Oboh-Ikuenobe, 2005) that coincided with establishment of the cold Benguela Current and the development

of hyperarid conditions in the Namib Desert (Ward et al., 1983). During the Early Pliocene, a warm, temperate climate with some dry intervals prevailed as is suggested by an increase in grass pollen in the offshore sediments (Udeze Oboh-Ikuenobe, 2005).

The "enormous volumes" of Kalahari dune sand (Korn and Martin, 1937; 1957) are the result of aeolian reworking of sands that form the uppermost layers of aggradational deposition in the Kalahari Basin (Thomas, 1986). The winds during this end-Tertiary period of extreme aridity were easterly in northern Namibia (Lancaster, 1981; 1987; Thomas, 1986). The white sheets sands that now largely block the palaeo river courses on the Cubango Megafan and the overlying pale red, linear dunes are attributable to this period of aeolian reworking. A late Pliocene to Early Pleistocene age for the onset of these extremely arid conditions is supported by observations and conclusions by Botha et al. (1986), Partridge (1993); Pickford et al. (1994), Tyson and Partridge (2000), Helgren and Brooks (1983) and Haddon and McCarthy (2005).

The Etosha fossils suggest that this extremely arid period lasted from about 4 Ma to 2 Ma, the latter date being that proposed by Buch (1996, 1997) and Buch et al. (1992) for the initiation of the evolution of the present Etosha Pan.

At ca 2 Ma, the climate once again became somewhat more humid. This is evident in southeastern Angola and eastern and southeastern Namibia where the dunes do not continue down into the valleys of the Cubango, Cuito, Kwando, Epukiro, Auob and Nossob Rivers (Mendelsohn and el Obeid, 2004; Mendelsohn et al., 2002). Instead, these rivers cut through the dunes. Furthermore, the occurrence of Early Stone Age implements on the highest river terraces in the Auob River suggests that the dunes must be significantly older than 500 ka (Volman, 1984; Korn and Martin, 1957).

Rust (1985) and Buch and Rose (1996) suggest that wind ablation and scarp retreat were the processes by which the present Etosha Pan developed. Buch (1997) suggests this was initiated at an elevation of 1130 m (Figures 2; 3) which would have been in calcrete of the Etosha Calcrete Formation (Figures 2; 3). Once periodic flooding in the Cuvelai system resumed, the Etosha area, still the lowest topographic region in the Owambo Basin, became the locus of a new terminal lake atop the Etosha Carbonate Member. However, in the hard calcrete of the lake bed, something other than wind must have played a role in enabling wind ablation to effect deepening. The agents for this were water and halite. Over the past 500 ka, cyclical wetter and drier periods have prevailed in southern Africa (Tyson, 1999; Partridge et al., 1997) but a far longer period of cyclicity is suggested if the initiation of the evolution of Etosha Pan goes back to 2 Ma. Repeated seasonal flooding and desiccation, probably much in the same way that flooding takes place today, will have facilitated salt build up. With each desiccation event, salt will have

crystallised in microfractures in the calcrete initiating fragmentation thereof. Strong easterly winds permitted removal of the calcrete fines. Such winds these days still produce huge offshore dust plumes (Mendelsohn and el Obeid, 2004). The terraces around the pan demonstrate that the post-calcrete evolution of the pan took place in several stages. Jaeger (1926) suggested that the pan-margin terraces were cut by wave action during high-water stands in the evolving Etosha Pan. However, highstand pan-margin deposits need to be found to support this contention. Alternatively, the terraces could be related to variations in the effectiveness of wind ablation corresponding to wetter and drier periods. Since the oldest lunette dunes on the western margin of Etosha Pan rest on the 1100 m terrace, some 30 m of calcrete (from 1130 to 1100 m) will have had to have been removed before accumulation of the first lunette dunes began. Thus, the lower palygorskite-bearing parts of the lunette dunes record wind ablation removal of only the remaining 6 m of the Etosha Calcrete Formation, if the original thickness of the latter had indeed been 36 m. Only thereafter did the underlying, analcime-bearing Etosha Pan Clay Member become re-exposed. Thus, the pan stratigraphy is inverted in the lunette dunes. The authors suggest that the pale grey, west-facing dunes on the northern margin of the pan and those near Ruacana have approximately the same age as the lunette dunes. Their orientations suggest that they were all produced by the same regional wind regime.

The base of the lunette dunes on the 1100 m terrace have a TL age of 140 ka (Buch, 1996; Buch et al., 1992). In excavating the present Etosha Pan over the past 2 my, wind ablation may have removed possibly as much as 46 m of stratigraphic section (Etosha Calcrete Formation, 36 m; Poacher's Point Carbonate Member, 5.5 m; Etosha Pan Clay Member – 4.5 m. See Figure 12).

Occasional flooding of Etosha Pan via the Cuvelai system still takes place today (Figure 8) and can partially or, less commonly, completely fill the pan. However, flooding in the pan can be dramatic. Almost 2000 mm of rain fell at Tsumeb in the summer season 1908/1909 (Schatz, 1997). As a result, the Omuramba Owambo ran for a whole year from east to west through Fischer's Pan into Etosha Pan and crocodiles and hippopotami were found in the river. These died once the river dried up. The flow in the Omuramba Owambo will have been due entirely to oversaturation of the carbonates of the Otavi Mountainland and the downhill flow from there through the karsted Etosha Calcrete Formation, the latter forming the bedrock to the Omuramba Owambo. The Cuvelai floods in the same year must have been tremendous (no records) with the result that the Etosha Pan is likely to have been totally flooded. Fr. Dufour, in a letter to Fr. Duparquet in 1880, pointed out that crocodiles, hippos and abundant fish occurred in the permanent bodies of water in the swampy regions in the Evale area in Angola (recorded in Stengel, 1963). In such a situation, it is probable that the crocodiles and

hippopotami migrated down the Cuvelai system. If this was the case, one can assume that it was not the only time that such an event occurred. Thus, the Cuvelai system could contain fossil records in places extending from the ca 4 Ma suite in the Ekuma Delta Sandstone to the youngest record of 1909.

There is no historical record of really deep water in the pan but the soft, olive green, slaked clay in gulleys on Pelican Island (Figure 16a) indicates that subsequent to excavation of the pan by wind ablation, flood water has filled the pan to depths of 10 m or more. These clay deposits appear to have initially resulted from the churning up of the pan floor during floods far more intense and far deeper than the recent floods in the pan. Once completely slaked and dried out, such deposits could easily be removed by relatively strong winds except where located in topographic wind shadows. Support for such large-scale flooding is provided by up to six past highstand strandlines between 140 ka and 12 ka in age and up to 26 m above the present pan floors in the Ngami and Mababe Basins in Botswana (Burrough and Thomas, 2008; Burrough et al., 2007).

However, flood waters have always evaporated leaving the pan as most people know it, dry and waterless. Nodules of white magadi-type chert have formed in places on the pan surface through the final dehydration of magadiite that crystallised from highly concentrated brines remaining in animal footprints during the final phase of desiccation in the pan.

Summary

The Kalahari siliciclastic sediments in the Owambo Basin consist of red clayey to sandy sediments of the Ombalantu, Beiseb and Olukonda Formations overlain by whitish sands, pale green clayey sands and pale olive green clays of the Andoni Formation (Table 2). Interfingering and coeval with all these siliciclastic units is the Etosha Calcrete Formation, a huge groundwater calcrete sourced from and fed by springs arising in the folded and karsted Neoproterozoic Otavi Group carbonates that form the elevated rim of the Owambo Basin.

Most of the Andoni Formation was deposited by megafans arising on the southern edge of the central Angolan highlands, the Cubango Megafan in the east and the Kunene Megafan in the west. Superimposed on the Kunene Megafan is the Iosimean Cuvelai Drainage System (Figures 4 and 7). These three fluvial systems had a common end point in the region of the lowest elevation in the Owambo Basin, Palaeolake Etosha.

Within this palaeolake, 50 m of pale olive green clays of the Etosha Pan Clay Member were deposited between about the Oligocene and 4 Ma by the Cubango and Kunene Megafans. These clays form the present floor of Etosha Pan (Table 2, Figure 3). Variable halite concentrations though the clay point to cyclical flooding and desiccation. The clays contain authigenic analcime, monoclinic K-feldspar and glauconite. The Oshigambo

Sandstone Member occurs interbedded with and 4 m from the top of the Etosha Pan Clay Member and is the first indication of the contribution of the Cuvelai Drainage System to the stratigraphic record at the top of the Andoni Formation. This sandstone contains a 6 ± 1 Ma vertebrate fossil suite similar to that from Langebaanweg, South Africa. Since the Cuvelai system overlies the Kunene Megafan, this sandstone indicates that by 6 Ma at the latest, build up of the Kunene Megafan, and probably the Cubango Megafan as well, had ceased. The Ekuma Delta Sandstone Member is the last fluvial unit to have been deposited in Palaeolake Etosha. It lies directly on top of the Etosha Pan clay and contains a 4 ± 1 Ma vertebrate fossil suite. The above-mentioned authigenic minerals in the top 4 m of the Etosha Pan Clay Member formed within the 4 Ma that followed deposition of the Ekuma Delta Sandstone Member.

Progressively decreasing flow into Palaeolake Etosha from the Cubango Megafan and the Cuvelai Drainage System is ascribed to the Okavango and Kunene Rivers starting to cut their present valleys. It was during this period that large-scale aggradational deposition in the Owambo Basin and possibly also throughout most of the southern Kalahari stopped. It was also the period during which the Okavango River assumed its southeasterly course and the Okavango Fan began to form.

The oolitic Poacher's Point Carbonate Member is a purely chemical deposit, is confined to the northern parts of the pan, is a pan-margin deposit, is the highest stratigraphic unit of the Andoni Formation and reflects a progressive decrease in the area of Palaeolake Etosha until it finally dried up completely.

During a period of extreme aridity between about 4 Ma and 2 Ma, the Etosha Calcrete Formation was able to advance northwards over the dry palaeolake deposits almost to the northern margin of the present Etosha Pan. Palygorskite is the characteristic clay mineral in the calcrete (Buch, 1996; Miller, 2008). At the same time, huge volumes of the surface sands in the Kalahari Basin were reworked, first into white aeolian sheet sands and then into pale red regional dune systems (Thomas, 1986). These aeolian deposits cover most of the Cubango Megafan. Very thin white aeolian sands cover the high ground between the Cuvelai channels.

At about 2 Ma, periodic flow in the Cuvelai system resumed. Flood waters accumulated in a new end-point ephemeral lake atop the Etosha Calcrete Formation in the Etosha Pan area. Repeated desiccation gradually led to salt build up which fragmented the calcrete. Wind ablation by the strong easterly winds of the arid anticyclonic weather system (Lancaster, 1981, 1987) removed all fines in this topographic depression (Rust, 1984; 1985). Wind ablation deepened the depression in stages leaving a series of terraces marginal to the pan and marking a progressively decreasing area at elevations of 1108 to 1110, 1097 to 1100, 1092, 1086 and 1082 m. The base of the oldest lunette dunes that

accumulated on the 1100 m terrace on the western side of Etosha Pan contains palygorskite derived from the remaining 6 m of the Etosha Calcrete Formation. The upper parts of the lunette dunes contain analcime derived from the subsequently re-exposed pan clays (Buch and Rose, 1996).

Floods in the Cuvelai system occasionally reach Etosha Pan. Complete filling of the pan is rare but soft, green, slaked clays in gulleys on Pelican Island indicate that past floods have been as deep as 10 m. During the final stages of evaporation of flood waters on the pan floor, white magadi-type chert nodules crystallised from the brines that remained in animal footprints.

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Appendix: Analytical procedures

A weighed amount of clay was placed in known amount of distilled water in a separating funnel. The clays disaggregated almost completely within minutes. After 15 minutes, the samples were thoroughly shaken in the separating column and allowed to settle. The accumulated sediment at the base of the funnels was taken off in separate samples after one minute, five minutes, 15 minutes, one to two hours and 24 hours and air dried.

Fragments of calcrete nodules were separated from quartz grains in the coarser grained samples. The calcrete nodules, the coarser fractions and the successive fines were subjected to XRD analysis. All XRD diffractograms revealed only the single peak of the (131) reflection of monoclinic K-feldspar. Not one had the (1 $\bar{3}$ 1) doublet of triclinic K-feldspar (Wright, 1968; Goldschmidt and Laves, 1954; Hearn et al., 1982).

Measured aliquots of the water remaining in the settling column were analysed for dissolved K and Cl

contents by Analytical Laboratory Services in Windhoek after remaining suspended fines were either filtering out or removed by centrifuging. K was determined by ICP-OES and a Perkin Elmer Optima 7000DV using the PlasmaCAL standard (Lot number SC 9013721) of 10000 µg/ml: relative standard deviation = 1%; relative error on synthetic samples at different concentrations = 3%; detection limit = 0.5 mg/litre. Chloride was determined by the argentometric method 4500-Cl-B described in Eaton et al. (1995): relative standard deviation = 5%; relative error on synthetic samples at different concentrations = 5%; detection limit = 1 mg/litre.

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