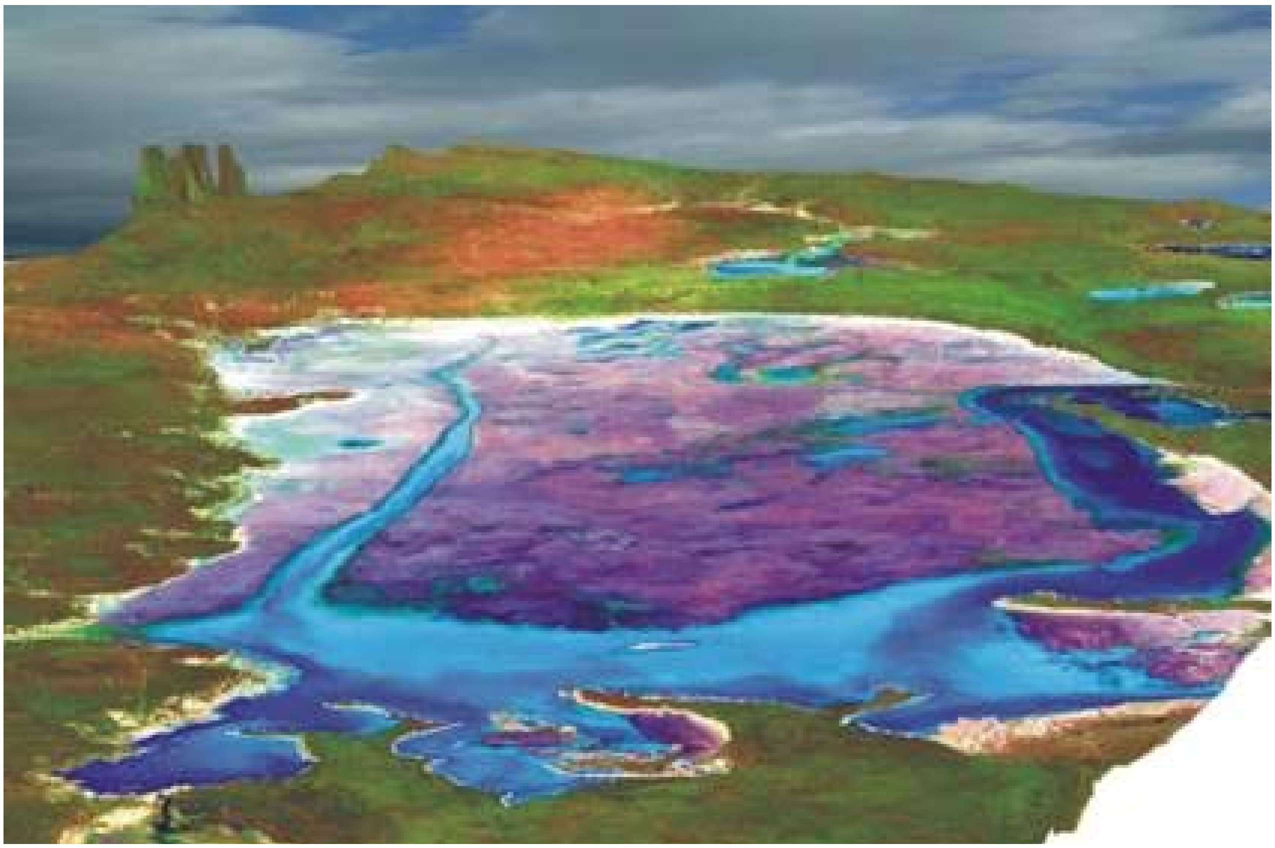
The Development and Evolution of Etosha Pan, Namibia



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To the demise of the Bantu Education System

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Abstract

This study explores and examines the geomorphology of a large endorheic basin, approximately twice the size of Luxemburg, situated in the Etosha National Park, Namibia. The main focus is directed on how and when this depression, known as Etosha Pan, came into being. Geomorphological investigation was complemented and guided primarily by the application and interpretation of satellite-derived information.

Etosha Pan has attracted scientific investigations for nearly a century. Unfortunately, their efforts resulted into two diverging and mutually exclusive views with respect to its development. The first and oldest view dates back to the 1920s. It hypothesized Etosha Pan as a desiccated palaeolake which was abandoned following the river capture of its major fluvial system, the Kunene River. The river capture was assumed to have taken place in the Pliocene/Early Pleistocene. In spite of the absence of fluvial input that the Kunene contributed, the original lake was thought to have persisted until some 35 ka ago, long after the Kunene severed its ties with the basin. The current size of the basin and its playa status was interpreted to have resulted from deteriorating climatic conditions.

The opposing view emerged in the 1980s and gained prominence in the 1990s. This view assumed that there were an innumerable number of small pans on the then surface of what later to become Etosha Pan. Since the turn of the Pliocene to Early Pleistocene, these individual pans started to experience a combined effect of fluvial erosion during the rainy season and wind deflation during the dry period. The climatic regime during that entire period was postulated to be semi-arid as today. This climatic status was used to rule out any existence of a perennial lake within the boundary of Etosha since the Quaternary. Ultimately, these denudational processes, taking place in a seasonal rhythm, caused the individual pans to deepen and widen laterally into each other and formed a super-pan that we call Etosha today. Thus the Kunene River had no role to play in the development of the Etosha Pan according to this model. However, proponents of this model acknowledged that the Kunene once fed into the Owambo Basin and assigned the end of the Tertiary to the terminal phase of that inflow.

Findings of this study included field evidence endorsing the postulation that the Kunene River had once flowed into the Owambo Basin. Its infilled valley, bounding with the contemporary valley of the Kunene near Calueque, was identified and points towards the Etosha Pan. It is deliberated that a large lake, called Lake Kunene, existed in the basin during the time. Following the deflection of the Kunene River to the coast under the influence of river incision and neo-tectonic during the Late Pliocene, new dynamics were introduced over the Owambo Basin surface.

After the basin was deprived of its major water and sediment budget that the Kunene River contributed, it was left with only smaller rivers, most notably the Cuvelai System, as the only remaining supplier. This resulted in the Cuvelai System concentrating and limiting its collective load deposition to a lobe of Lake Kunene basin floor. The accident of that lobe is unclear, but it is likely that it constituted the deepest part of the basin at the time or it was influenced by neo-tectonic that helped divert the Kunene River or both. Against the backdrop of fluvial action that was initiating the new lake, most parts of the rest of the basin, then denied of lacustrine activity, were intermittently riddled with a veneer of sediment, especially during phases of intensified aeolian activity. In the mean time, the area that was regularly receiving fluvial input started to shape up as a distinct lake with the depositions of sediments around the water-body, primarily via littoral action, serving as embankment. Gradually, a shoreline is formed and assisted in fixing and delineating the spatial extent of the new and much smaller lake, called Lake Etosha. That Lake Etosha is the predecessor of the modern day Etosha Pan.

Indicators for a perennial lake found in this study at Etosha include fossil fragments of *Clariidae* species comparable to modern species measuring some 90 cm, and those of sitatunga dated to approximately 5 ka. None of these creatures exist today at Etosha because of their ecological requirements, which among others, include permanent water. The sitatunga, in addition, is known as the only truly amphibious antelope in the world.

Since its inception, the new lake underwent a number of geomorphological modifications. A prominent character amongst these modifications is the orientation of the lake, which has its long-axis oriented in the ENE-WSW direction. It resulted from wave action affected by the prevailing dominant northeasterly wind, which is believed to have been in force since the Middle Pleistocene.

Lake Etosha has also witnessed phases of waning and waxing under the influence of the prevailing climatic regime. Over the last 150 ka, the available data intercepted about seven phases of high lake levels. These data are generally in agreement with regional palaeoclimatic data, particularly when compared with those obtained from neighbouring Makgadikgadi Pans in Botswana. The last recorded episode of the wet phase at Etosha was some 2,400 years before the present.

Ausführliche Zusammenfassung

Die vorliegende Studie untersucht Aspekte der Geomorphologie und Landschaftsgeschichte eines großen, heute endorheischen Beckens im zentralen Nord-Namibia, nämlich des Owambo-Beckens als Teil der Kalahari im weitesten Sinne. Der rezent tiefste Teil dieses Beckens umfasst die so genannte Etoscha-Pfanne im gleichnamigen Nationalpark; sie besitzt eine Fläche von etwas zweimal der Größe Luxemburgs. Dabei wurde das Hauptaugenmerk auf die Frage gelegt, wie und wann diese Depression unter Berücksichtigung der tektonischen und klimatischen Entwicklung des Großraums entstanden ist. Hierzu wurden zwischen 2002 und 2004 umfassende Geländestudien von mehreren Monaten Dauer mit unterstützender Luft- und Satellitenbild-Auswertung sowie Kartier- und Laborarbeiten unternommen.

Die Region um die Etoscha-Pfanne ist seit fast 100 Jahren Gegenstand wissenschaftlich-geographischer Forschung, wobei die jeweiligen Resultate zu teilweise sehr konträren Hypothesen geführt haben. Die älteste publizierte Theorie (1920er Jahre) geht davon aus, dass es sich bei der Etoscha-Pfanne um den ausgetrockneten Endsee des Kunene-Flusses handelt, welcher seinerseits gegen Ende des Tertiärs oder zu Beginn des Quartärs infolge Flussanzapfung von Westen her zum Atlantik umgelenkt wurde. Diese Umlenkung soll das Ende des Zuflusses und demzufolge die sukzessive Austrocknung des Etoscha-Sees bewirkt haben, der aber – so die Hypothese – noch bis vor ca. 35.000 Jahren bestanden haben soll. Sowohl die Größe der Etoscha-Depression wie auch der Playa-Formenschatz und die umlaufenden Terrassen-Galerien wurden als Indikatoren dieser insgesamt zunehmenden Austrocknung gedeutet.

Eine gegensätzliche Theorie wurde ab den 1980er Jahren entwickelt, wonach die Entwicklung des regionalen Flußsystems, insbesondere des Kunene, mit der Entstehung der Pfanne nicht in Verbindung zu bringen sei. Es wurde aber auf die Möglichkeit hingewiesen, dass im Tertiär der Kunene in das Owambo-Becken geflossen sein könnte und somit für Teile der Kalahari-Formation als Beckenfüllung verantwortlich wäre. Vielmehr sei im tektonisch prä-disponierten Beckentiefsten nach-kunenezeitlich eine Megapfanne infolge rückschreitender Hangentwicklung aus kleineren Pfannen entstanden. Offen blieb jedoch die Frage, in welcher Weise sich die wirksame Erosion in unmittelbarer Nähe des Vorflut- oder Haupterosionsniveaus in einem sehr flachen Großrelief als Prozeß etablieren und zum kilometerweiten Rückschreiten der wenige Meter hohen Hänge führen konnte. Dies hat dazu geführt, dass ab den 1990er Jahren der Wirkung des Windes diese erosiven Fähigkeiten zugesprochen wurden. Demnach habe sich etwa ab der Tertiär-Quartär-Wende unter generell gleich bleibenden semi-ariden Umweltbedingungen eine Art Kombinationseffekt aus saisonaler fluvialer Aufbereitung mit ebenso saisonalem äolischen Materialaustrag etabliert. Dieser soll zur sukzessiven Tieferlegung des Pfannenbodens seit dem Ende des Tertiärs infolge von Auswehung geführt haben. Als Indizien hierfür werden u. a. eine mangelnde bzw. sehr dünne Sedimentdecke in der Pfanne sowie die Existenz von äolischen Sandkörpern, den so genannten Lunette-Dünen am Westrand der Pfanne angeführt. Die Existenz eines perennierenden Sees ist in dieser Hypothese explizit ausgeschlossen worden, zumindest für die letzten 140.000 Jahre.

Die Ergebnisse der Gelände- und Auswertungsarbeiten vorliegender Studie können beide genannten Hypothesengebäude nur zum Teil bestätigen. Die Untersuchung der Wasserscheidenbereiche zwischen Kunene-System und dem ins Etoscha-Becken entwässernden Cuvelai-Oshana-System insbesondere am westlichsten Zweig, dem Etaka-Fluß, erbrachte das Resultat, dass ein durchgängiges, sehr breites und teilweise verfülltes Paläotal von einem nordwärts fließenden Kunene-Tributär bei Calueque, dem Olushandja, über die Wasserscheide hinweg zum Etaka und damit nach Etoscha führt. Die Rekonstruktion des direkten Wasserscheidenbereiches ist heute nur noch schwer möglich, da direkt auf der Wasserscheide ein Stausee der nordnamibischen Wasserfernversorgung angelegt wurde. Anhand der wasserbaulichen Planungs- und Bauunterlagen konnte jedoch die Lage relativ genau lokalisiert werden. Die Höhe der Wasserscheide über dem rezenten Normalwasserbett des Kunene beträgt nur 15 Meter und es ist daher anzunehmen, dass die Umlenkung des Flusses keine allzu starke tektonische Aktivität erforderte, zumal das obere Tertiär im subkontinentalen Maßstab als

Periode starker Hebung in den Randstufenbereichen gilt. Der Kunene selbst fließt heute westlich des Anzapfungsbereichs in einem von teilweise starken Längsprofil-Diskontinuitäten gekennzeichneten und tief ausgeräumten Tal, angelehnt an eine tektonische Struktur, die in der angolanischen Literatur als Ruacana-Graben bezeichnet wurde. Auch das intramontane Becken von Ruacana ist wohl als Weiterbildung einer ursächlich tektonisch bedingten Depression nahe der kontinentalen Hauptwasserscheide anzusehen.

Das Vorkommen einer noch in geomorphologisch jüngerer Zeit möglichen Durchflussverbindung lässt auf die zeitweise Existenz eines oder mehrerer perennierender Seen schließen, zumindest in Zeiten stärkerer Wasserschüttung im angolanischen Hochland. Genauere zeitliche Einordnungen können nicht gemacht werden, da derzeit noch kein adäquat datierbares Material vorliegt. Unter der Annahme, dass ein solcher Kunene-See mit oder ohne ehemaligem Ausfluss im Beckentiefsten existiert hat, trat nach der endgültigen Ablenkung des Zulaufs definitiv ein Systemwechsel hin zu einer anderen Dynamik auf der Oberfläche des Owambo-Beckens ein. Das durch die Neotektonik bedingte, deutlich verkleinerte und nicht mehr weit ins angolanische Hochland reichende Cuvelai-Oshana- und Gwashigambo-Einzugsgebiet der Beckendepression muss in deutlich geringerem Maße fremdgesteuert gewesen sein, so dass die autochthonen Niederschlagsbedingungen sich in den Formen stärker niedergeschlagen haben müssen. Dies gilt insbesondere für pleistozäne Trockenphasen, wie sie anderweitig in Namibia vor allem in den jeweiligen Hochglazialen mit Meeresspiegel-Tiefständen und intensivem Benguela-Upwelling als nachgewiesen gelten dürfen. Während dieser Trockenphasen dürfte die fluviale Aktivität sehr stark zum Erliegen gekommen und ein kaum nennenswerter fluvialer Input in die Vorflutbereiche geschüttet worden sein, so dass besonders in den extremen Trockenphasen Auswehungsprozesse dominiert haben. So sind insbesondere die Längsdünenzüge im zentralen Nordnamibia aber auch in anderen Regionen des Landes auf die Ausbildung oder Reaktivierung im letzten Hochglazial zurückzuführen. Als besonderes Indiz für die neotektonische Aktivität ist die bei Flutereignissen im Satellitenbild erkennbare ostwärtige Neigung der Pfannenoberfläche zu sehen, was auf der Ostseite zu ausgeprägter Kliffdynamik führt.

In den Übergangsphasen dürften die Verhältnisse den heutigen mehr oder weniger ähnlich gewesen sein mit saisonalem, episodischem oder periodischem Input sowohl aus der direkten Umgebung des Beckentiefsten wie auch aus dem Oshana-System. Signifikante Unterschiede bestehen jedoch zu den pluvialzeitlichen Verhältnissen, die im Verlauf des Pleistozäns mehrfach aufgetreten sind. Hier herrschten perennierend feuchte Bedingungen in weiten Teilen des Einzugsgebiets, das durch dichten Trockenwald und Feuchtsavannen zu charakterisieren ist. Als Indikatoren für pluvialzeitliche Verhältnisse wurden fossilführende Sedimente im Pfannenbecken, Strandterrassen und Flussterrassen im unteren Einzugsgebiet untersucht. Die Sedimente erbrachten Fossilfunde von großen Fischen der Gattung *Clariidae* – die größten Fische, die je entdeckt wurden – und von Land- und Frischwassermollusken in unterschiedlichen Morphopositionen. Ein jünger angeschnittener Nehrungshaken enthielt u. a. Fossilien von Sitatunga, ein klarer Hinweis auf perennierende Seeverhältnisse. Letztere konnten radiometrisch auf ca. 5000 BP datiert werden – ein Datum, welches mit der in der Literatur immer wieder beschriebenen frühholozänen Feuchtphase in Namibia ebenso gut korreliert, wie mit der jüngsten Bodenbildungsphase auf den Strandwällen.

Innerhalb der geschilderten Befunde sind die früher als Lunette-Dünen beschriebenen Akkumulationskörper auf der West- und Nordseite der Pfanne ebenso als reliktische Strandwälle anzusprechen wie die identisch strukturierten, längsgestreckten, der heutigen Küste vorgelagerten Inseln, z. B. Logans Island. Letztere repräsentieren sehr wahrscheinlich das Niveau der Strandwälle einer vergangenen Seephase, als das heutige Pfannenbecken bereits bestand aber nicht komplett mit Wasser geflutet war. Sie wurden durch die nachfolgende Seephase, die vermutlich auf 5000 BP datierte, bis auf die Reliktinseln überformt. Die genannten Strandwälle aller Reliefgenerationen enthalten Mollusken der Art *Xerocerastus burchelli* und einige andere, ebenfalls häufig wechselnde oder feuchte Bedingungen bevorzugende Arten. In den älteren Strandwällen sind diese Fossilien mit carbonatischen Ausfällungen überzogen, teilweise auch vollständig verbacken, in den jüngeren Sedimenten, wie Logans Island oder dem datierten Nehrungshaken, stecken sie in frischer Erhaltung im Sedimentkörper. Lebende Exemplare konnten jedoch nicht aufgefunden werden.

Ein weiteres Indiz für die Interpretation der Sedimentkörper als fossile Strandwälle ist darin zu sehen, dass dieselben infolge Bodenbildung fest liegen und lokal durch autochthone Niederschläge gullyartig zerschnitten werden. Indizien für eine rezente äolische Entstehung infolge der Auswehung vom Pfannenboden konnten nicht gefunden werden. Stattdessen konnte gezeigt werden, dass Auswehung allenfalls den Schwemmbereich des westlichen Hauptzuflusses, den Ekuma, betraf. Insgesamt scheint der Anteil der äolischen Formung weit hinter dem der fluvial-lakustrinen zurückzutreten.

Insgesamt konnte gezeigt werden, dass die Region des zentralen Nordnamibia mit der Etoscha-Pfanne als rezentem Hauptvorfluter in ihrer tektonisch-strukturellen Prädisposition eine wechselhafte jungkänozoische Reliefgeschichte mit alternierenden Pluvialzeiten und Trockenphasen jeweils unterschiedlicher Dimension sowie Übergangsphasen aufweist.

INTRODUCTION

1.1 Background

One of the largest and oldest wildlife parks in the world is Etosha National Park (ENP) situated in north-central Namibia. The core of this nature reserve is occupied by a large (4760 km^2, Lindeque & Archibald 1991) depression known as Etosha Pan, from which the name of the park was derived. Despite its prominent ecological importance, particularly in functioning as a wetland in semi-arid Etosha, this pan has attracted relatively little geomorphological investigations. A handful of research studies that examined the area culminated in conflicting results with respect to the time and mode of its development. This study is thus aimed at exploring the evolution of this salient topographical feature of the park. In turn, results of the study will help reconstruct the palaeoclimate and palaeo-environment of the Etosha region.

1.2 Aims and Objectives of the Study

Prudent and appropriate management of our natural environment is largely dependent on our understanding of the internal and external factors that govern it. Some of the baseline data necessary for enhancing such an understanding rest in geomorphological studies of those environments. Geomorphology is the study of landforms and landscapes, particularly from the standpoint of origin. It seeks to relate specific landforms to the formative processes, such as aeolian and glacial, that molded them. Thus it strives for entangling the origin and development of present landforms, the exploration of their relationships to underlying structures, and deciphering recent geological and climatic changes that are recorded by and reflected in these surface features (e.g. Garner 1974; Büdel 1980, 1982; Busche & Hagedorn 1980; Chorley *et al*. 1984; Tinkler 1985; Hart 1986; Bloom 1991). It is therefore the purpose of this study to subject the Etosha Pan, having its mode and time of origin under intense debate, to this arm of landscape analysis.

Specifically, pertinent and vexing questions to be addressed are summarized as follows:

- The origin of Etosha Pan is hypothesized into two opposing and mutually exclusive models. These views are that the Etosha Pan is either a desiccated palaeolake or formed by virtue of aeolian deflation. In essence, this implies that the sediment balance of this basin is either positive (aggradational) or negative (erosional). The first question to be posed is therefore whether or not sedimentation exceeded erosion in the Etosha Pan.
- The answer to be obtained from the first inquiry posed above would naturally trigger a follow-up question. This is directed to the nature of the formative process or processes that were responsible for sediment deposition in or removal from the surface of the pan to characterize the sediment balance of the Etosha basin. In other words, if for example, the Etosha Pan is indeed an erosional landform, which agent(s) of denudation was or were at work?
- A crucial component fueling the current debate on the evolution of the Etosha Pan is the role of the Kunene River, or the lack of it, in the pan's development. It is therefore imperative to investigate whether or not there was a geomorphic relation between the Kunene River and the Etosha Pan.
- The final, major aspect to be addressed in this study is practically of hindsight in character, drawing inference from geomorphological evidence gathered for answering the preceding questions. It will contextualize the

palaeoclimate and palaeo-environment of the Etosha region at the time of the pan's development.

To answer these questions, geomorphology as a stand-alone approach is rather inadequate in applying it to a large area such as Etosha region and under the limited time frame of this investigation. Conventional geomorphological approaches, such as fieldwork and laboratory analysis, are thus synergized with Remote Sensing and auxiliary data. Due to the handling of large spatial data, a Geographic Information System (GIS) is employed for data management, integration, analysis, and output.

CHAPTER 2

METHODS AND TECHNIQUES

Introduction

This Chapter deals with the conceptual methodology applied and the procedures followed in this study. Remote Sensing, fieldwork, laboratory and data integration under the GIS environment were the principal methods and techniques employed. Although these methods are distinct in many ways, they are undoubtedly complementary. Auxiliary data used in carrying out this study and the discussion of main steps taken to generate the necessary data layers are also highlighted in this section.

2.1 Remote Sensing (RS)

The benefit of a synoptic, bird's eye view in observing the terrain necessitated the acquisition of remote sensing data of various spatial, spectral and temporal resolutions. These data were taken from both air- and space-borne platforms. The former platform is represented by panchromatic aerial photographs at a scale of 1:50,000 (1970) and 1:78,000 (1997).

From space, a host of complementary images from CORONA, Landsat TM, Landsat-7 ETM+ and ASTER (Advanced Spaceborne Thermal Emission and Reflection Radiometer) programs were obtained. Collectively, these images cover the entire Cuvelai System, predominantly during the dry season. The Cuvelai is the main fluvial system that supplies water to the Etosha Pan (section 4.3). Few other images were taken around April, approximately the end of the rainy season. The oldest data set in the collection was acquired on August 29, 1963 from a second generation of the US spy satellite, CORONA. At best ground resolution of 140 m, this image has a ground coverage of approximately 480 by 480 km. The negative was scanned at a resolution of 1200 dpi, and subsequently georeferenced for integration in a GIS environment. The rest of the satellite images were taken between 1984 and 2002.

Typically, Landsat TM image, bands 7, 4, and 2 in red, green, and blue, respectively, is digitally enhanced using the linear stretch, following radiometric corrections. This color combination is generally recognized for its effectiveness in imaging arid and semi-arid terrains around the world (e.g. Sabins 1999). Under this false color composite, green vegetation is shown by green signatures, whilst different types of bedrocks or ground cover are displayed through a variety of color shades and texture.

A slight deviation from the above generic false color combination was extended to the Landsat ETM+ imagery, in favour of a multi-resolution integration. This was made possible due to the availability of the 15 m panchromatic band operating in the 0.50 to 0.90 μm range that is incorporated in the ETM+ instrument. In taking advantage of this higher spatial resolution, the "pan sharpening" procedure was implemented. Band 2 was thus substituted with the 15 m panchromatic band to achieve an enhanced image sharpening. In essence, such a process yields a color composite of a 15 m resolution (Lillesand & Kiefer 2000).

Merging of multi-temporal data was applied for western Etosha Pan in relation to lunette dunes. Figures 2.1a and c show false color composites (7-4-2 in RGB) of Landsat TM images acquired in the month of April in 1992 and 1997, respectively. Although the images were acquired at the same time of the year, their respective spectral properties were heavily influenced by climatic forcing. Specifically, 1992 was an exceptionally dry year in the recorded history. It was the driest year since 1965 and third driest since 1934 at Okaukuejo (Etosha Ecological Institute, unpublished data). Subsequently, drier conditions translated into higher reflectance in the visible and reflected bands when comparing the 1992 image with that of 1997. Conversely, the image taken in 1997 reflected conditions of an above long-term average rainfall that was received in that year. Although the lunette dunes are discernable in both image sets, they do not distinctly stand out from their surroundings. Fusing the two images into one false color composite, however, presented a distinguishing visual interpretability for the lunette dunes and their associated landforms. Figure 2.1b illustrates this enhanced benefit of a multi-temporal color composite in discerning lunette dunes.

Image interpretations followed conventional approaches detailed by van Zuidam *et al.* (1985) and Lillesand & Kiefer (2000). Guided by geomorphological principles, this procedure started with the detection, recognition, and identification of relevant surface expressions and ground features depicted in the image. It then interactively progressed to image analysis, which involved the division of the remote sensing image into constituent parts, the *"photomorphics"* regions. This division was based on quantitative and qualitative evaluation of certain objects, such as spectral properties, drainage pattern, natural vegetation cover, and the portrayal of geomorphic units and structures. The resultant division of image characteristics was then refined with knowledge-driven input gathered from other sources, such as relevant maps, field observations and laboratory results.

Figure 2.1: Illustrative multitemporal false color composite of Landsat images covering the western part of Etosha Pan, a) image acquired April 1992, b) multitemporal color composite of image a and c, and c) image acquired April 1997. Note: 1992 was exceptionally dry, whilst rainfall in 1997 was above the long-term average.

Shape, size, pattern, tone, texture, and association were the major characteristics which were systematically considered during the image interpretation and evaluation stages. The general form, configuration and outline of pans, channels and major landforms in the area were closely observed. Spatial arrangement of terrain units, in association with one another, was featured at length, particularly in studying the so-called lunette dunes lying in the north-west and western parts of Etosha Pan. Soil moisture condition was also characteristically a significant factor, both in the pan interior and outlaying areas.

Topographic data are indispensable in geomorphological studies. The best available elevation data for the central part of the Mega Kalahari was derived from the Shuttle Radar Topography Mission, SRTM. Although this product has an inherent vertical accuracy of 20 meters, close observation revealed that a part of the Owambo Basin surface over which the Etosha Pan lies has a localized, systematic error, in a form of stripes. These stripes start from the south-eastern

Etosha Pan and run in a northwesterly direction towards the Kunene River. They are more pronounced around the Omadhiya Lakes. Figure 2.2 illustrates this inaccuracy in the data. Although values included in the banding areas are within the 20 m accuracy, the stripes in themselves have a negative effect on longitudinal profiles of the area, as demonstrated in Figure 2.3. In a lowland area, such as the Cuvelai, a difference of more than 5 meters is indeed significant. Notwithstanding, original data were used in constructing longitudinal profiles, and detected errors were pointed out accordingly in the respective figures.

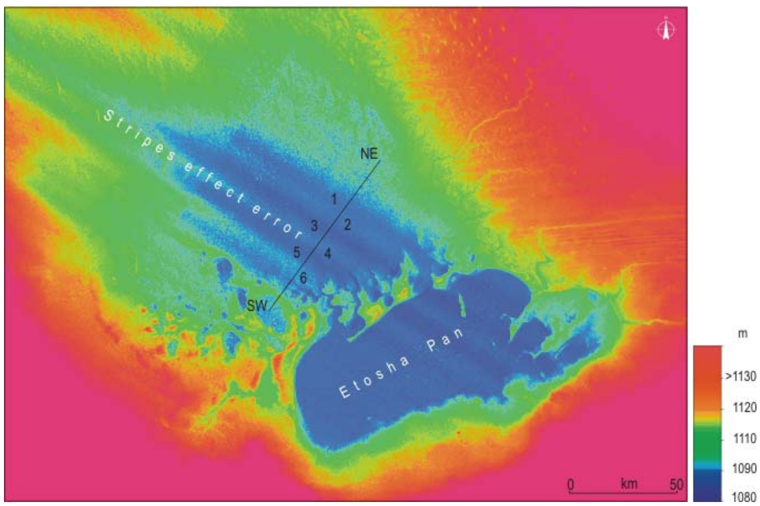

Figure 2.2: Error in SRTM Elevation data in part of the Cuvelai System

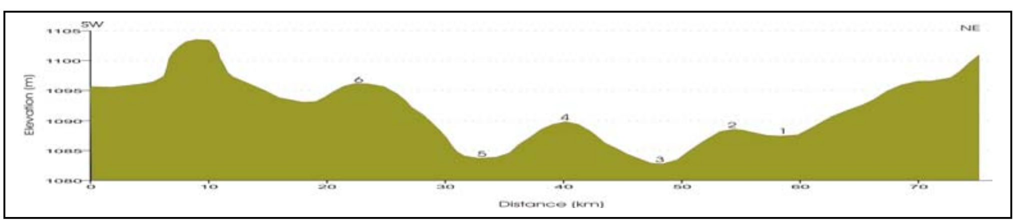

Figure 2.3: Propagation of the error into a longitudinal profile (see Figure 2.2 for location)

In comparison with the DEM of Etosha National Park, which was generated from the topographic maps (1:50 000) with a contour interval of 10 m, the SRTM data over-estimates the elevation by an average of 0.16%. This is mainly attributed to a different reference datum employed. The topographic maps use the Bessel 1841

(Namibia) ellipsoid, whereas the SRTM data is based on the WGS84 system. No attempt was made to transform either data set to a common reference datum.

ILWIS 3.2 (ITC, the Netherlands) was the main program employed for image processing. It was closely supplemented with ERDAS 8.6 (Erdas, Atlanta, GA) and ENVI 3.5 (Research System, CO) due to specialized features that they offer. For example, for the spectral library feature, the ENVI program was utilized when dealing with ASTER data.

2.2 Fieldwork

By and large, fieldwork was designed and executed based on the analysis and interpretation of remote sensing data. Two periods of fieldwork were carried out in the course of this study. The first leg of fieldwork was designed to observe the terrain right after the end of the rainy season, whilst the second and last one was carried out at the height of the dry season. Figure 2.4 presents an example of processed Landsat images used as a basis for ground truthing and field investigations.

Field data collection concentrated mainly in and around the Etosha Pan. Landscape observation and analysis, however, extended to most parts of the Etosha Pan's catchment area, located within Namibia. Coverage included the Omuramba Owambo, iishana, Ekuma and Oshigambo rivers, minor pans, Oshana Olushandja / Etaka and the surrounding. Figures 2.5 and 2.6 depict major locations where either data collection was done or detailed landscape observations were carried out.

Data collected included samples for sedimentological analysis gathered from soil pits or by using a hand-held soil auger (10 cm diameter; Eijkelkamp Agrisearch Equipment, The Netherlands), rock samples, measurements of profiles and ground-truthing of remote sensing data.

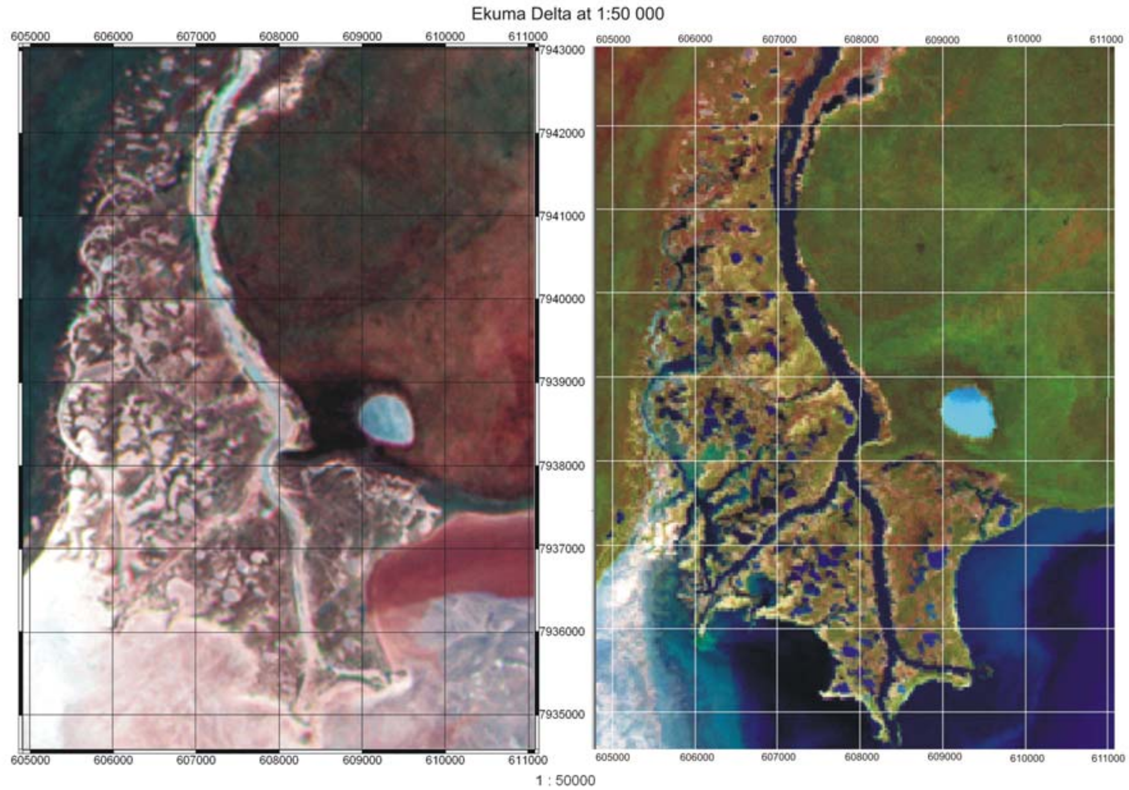

Figure 2.4: Sample from the "field-guide" book generated from satellite images for fieldwork purposes: Ekuma Delta - left image acquired in 2002 during the dry season (Landsat ETM+ Bands 7-4-8, in RGB) and right image acquired in 1997 towards the end of the rainy season (Landsat TM Bands 7-4-2 in RGB)

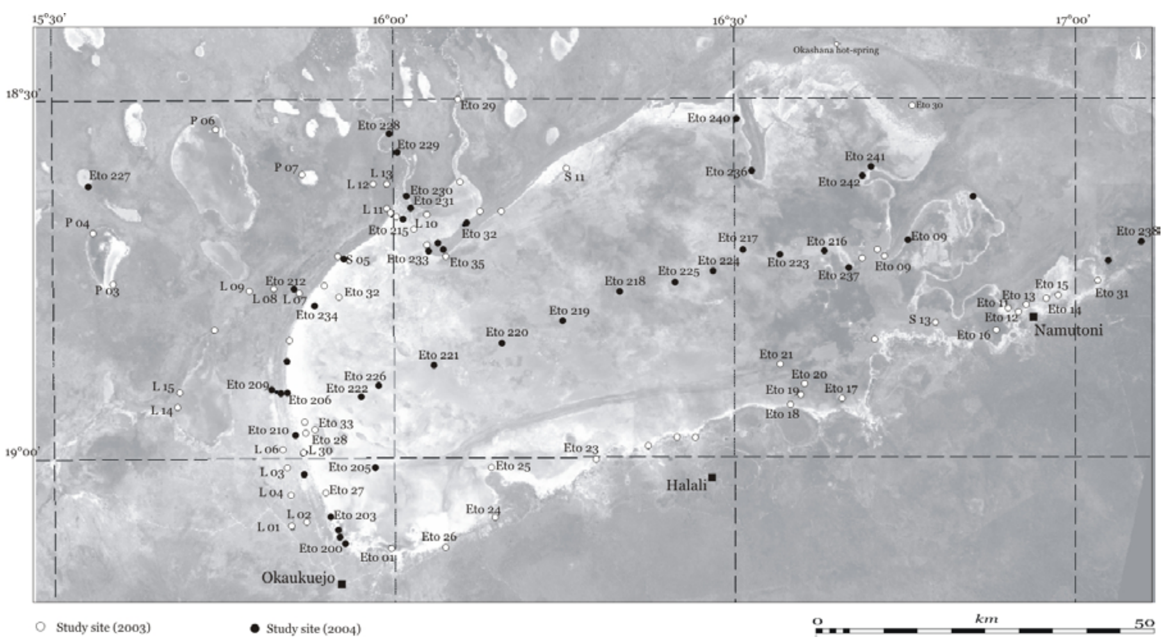

Figure 2.5: Major study sites in and around Etosha Pan

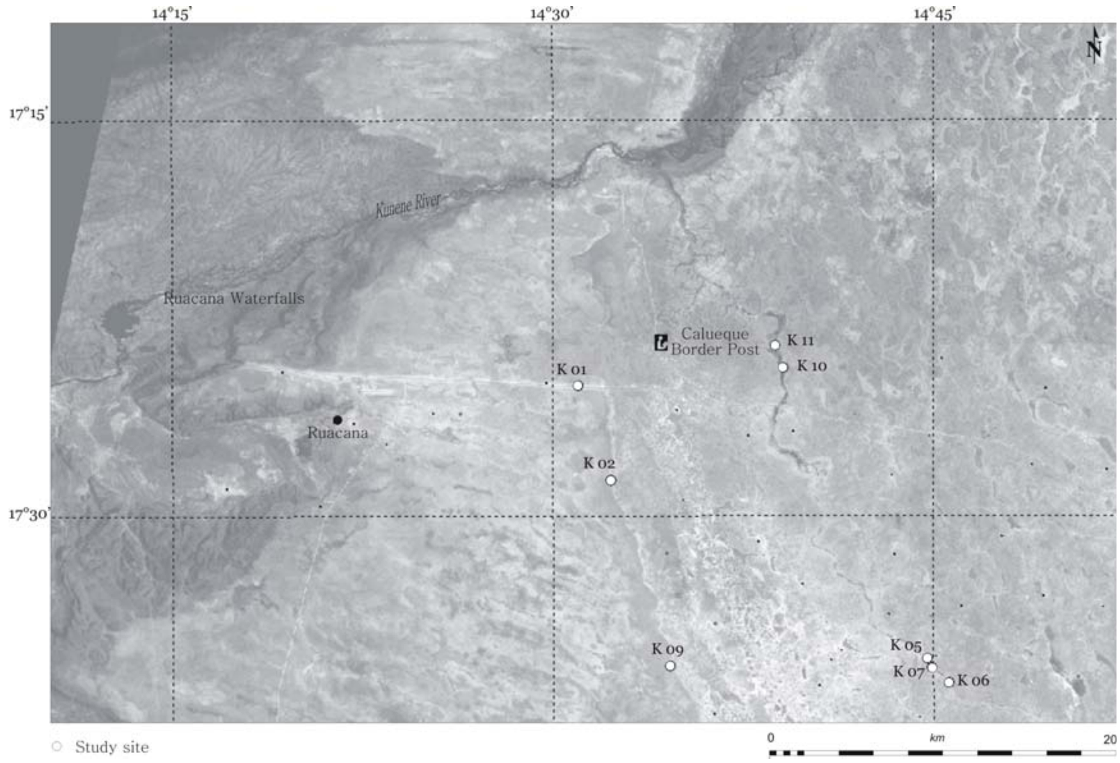

Figure 2.6: Major study sites in and around Oshana Olushandja / Etaka

2.3 Laboratory

Samples collected from fieldwork were subjected to a host of laboratory analysis. Following standard procedures (Schlichting *et. al.* 1995), grain size analysis, pH-value (in KCl), and carbonate content ($CaCO_3$) were carried out at the geomorphological laboratory of the Geography Department at the University of Wuerzburg. Briefly, for sedimentological analysis, for instance, precursory steps included the removal of carbonate by hydrochloric acid, and the application of hydrogen peroxide for digesting organic matter. Sodiumdiphosphate (0.4 n $Na_4P_2O_7 \times 10H_2O$) was used as a dispersing agent. Sand particles were then determined using a wet sieving method, while the sedimentation method was applied for mud particles.

Identification of fossils was made by several experts in their respective fields. Gastropods were positively identified by Dr Willem F. Sirgel at the University of

Stellenbosch, South Africa. Dr Hélène Jousse, at the University of München, worked on and identified fragments of a *Clariidae* species, sitatunga metatarsal, and other antelopes. Dating of fossils was done at the AMS Laboratoriet för C14-datering of the University of Lund, Sweden (Dr G Skog).

2.4 Geographical information system (GIS)

Arguably, the single most important feature of a GIS is the ability to integrate spatial data from various sources and scales. This ability was recognized from the inception of this study, and GIS was thus exploited as a tool and platform for data integration, management, visualization and output. ILWIS 3.2, a raster and vector hybrid GIS package, was used extensively. Occasionally, however, Arcview 3.2a was utilized to complement ILWIS 3.2 when necessary.

CHAPTER 3

Background Setting

Introduction

This chapter introduces the background setting of the study area. This includes the geology, drainage evolution and soils. Embedded in this coverage are relevant landforms in the region, such as the Okavango System, which is often compared to and/or contrasted with the Etosha Pan.

3.1 Location of the Study Area

Distinctly visible from space at a continental scale, the Etosha Pan is situated in the southwestern part of the African continent (Figure 3.1a), and falls within the boundary of Namibia (Figure 3.1b). The catchment area of this endorheic Etosha basin reaches the southern section of the Angolan Highlands in the north and borders respectively in the east and west with those of the Cubango/Kavango and Kunene, two of the few great rivers in the sub-continent.

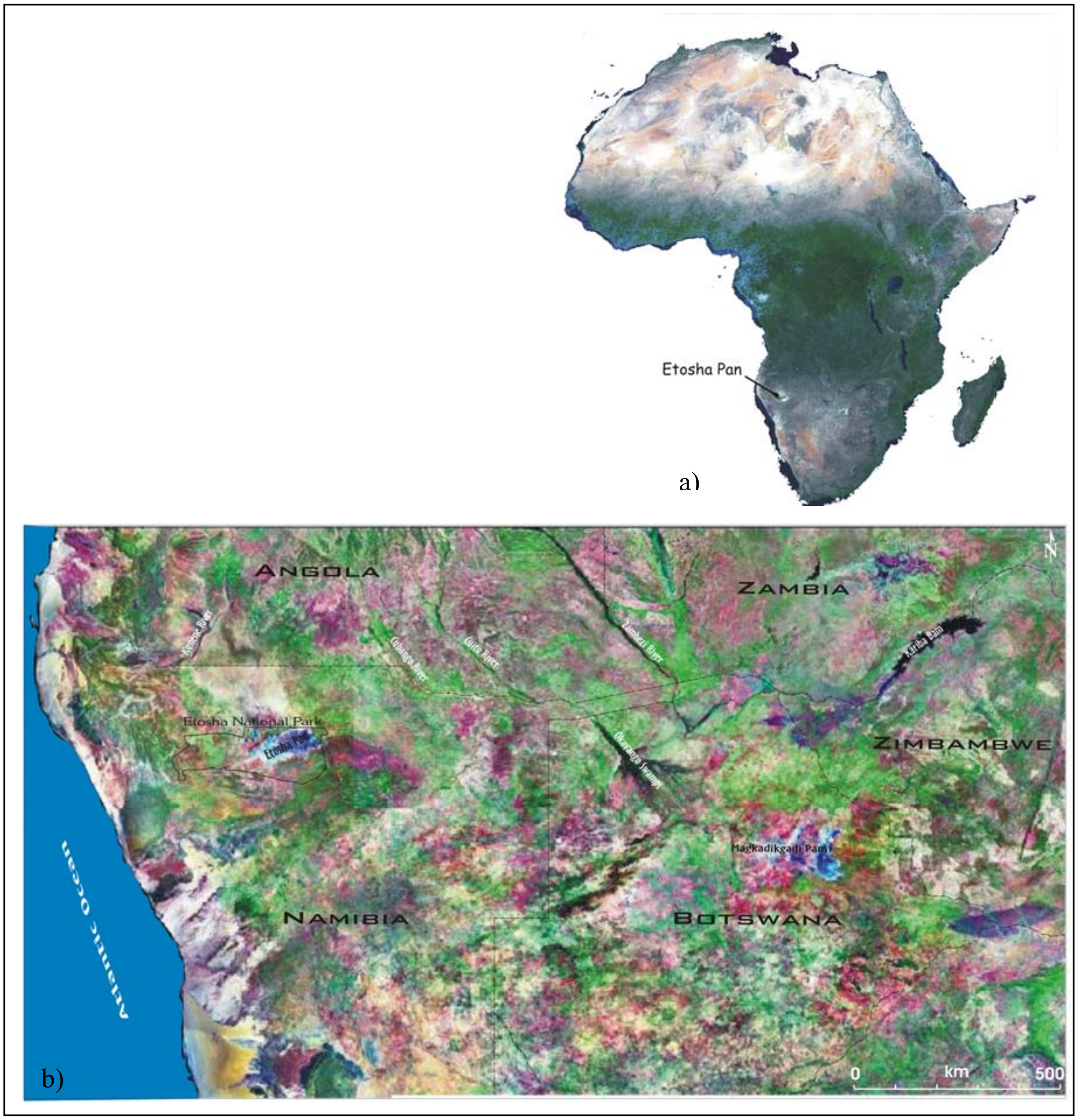

Figure 3.1: Location of the study area at a) continental scale and b) regional level

The Kunene discharges into the Atlantic Ocean, whereas the Cubango/Kavango pours its water into the Okavango swamps, the largest inland fan delta in the world (Thomas & Shaw 1991). A rim, up to 400 m above the surrounding plains, located on the southern and south-western margin of the Etosha basin separates its drainage from the adjacent catchments of four ephemeral rivers, the Ugab, Huab, Hoanib, and Hoarusib, all draining to the Atlantic Ocean.

The Etosha Pan occupies just over a fifth of the second largest national park in Namibia. At 22,700 km^2 (Mendelsohn *et al.* 2002), Etosha National Park plays a vital conservation and economic role in that nation. The park is home to 114 mammals species, 340 birds species ($^1/_3$ migratory), and a variety of trees (Van Schalkwyk, 1999). In turn, the booming tourism industry in the country depends mainly on some of these natural endowments, which helped ranking Etosha as the number one tourist destination in Namibia.

3.2 Geology

The Etosha Pan lies on the surface of the Owambo Basin, an old sea shelf dating back to the Late Precambrian. The basin owes its origin to the break-up of a super-continent called Rodinia. Today, it is one of the sub-basins that make up the greater Kalahari Basin. As one of the largest sub-basins, it covers part of southern Angola and extends into northern Namibia (Figure 3.6). The name of the wider Owambo basin is often interchanged with Etosha basin, which is practically a sub-set of the Owambo. The geology of the Owambo basin has been reconstructed based on marginal outcrops, coupled with seismic, aeromagnetic and gravity surveys and sparsely distributed wells. Most of these data have been obtained from the Namibian half of the basin, hence the Angolan half is poorly represented and meagrely known.

3.2.1 Pre-Kalahari Geology

The Damara Sequence, comprising the Nosib, Otavi, and Mulden Groups, is the oldest geological unit covering the entire basin. This sedimentary unit lies on an old continental base of granites, gneiss and volcanic rocks formed between 2,600 and 1,730 Ma (Hedberg 1979; Miller 1997). During the Late Precambrian the super-continent, Rodinia, started to break-up. Within this protracted, multi-phase period a variety of sandstones, collectively known as the Nosib Group (Figure 3.2) were deposited in the gorges and rift valleys of this fragmenting intracontinent (Hedberg 1979; Miller 1997; Mendelsohn *et al.* 2000). This deposition occurred roughly between 900 and 730 Ma ago (Miller 1983) and now has a preserved maximum thickness of about 1,400 m (Miller 1997). As the rifting continued, plates of Rodinia began to spread, forming among others a broad, shallow-water ocean on the southern edge of the Congo Craton, the northern landmass. It was in this stable platform where dolomites and limestones were deposited and accumulated to a maximum thickness of 7,615 m (Hedberg 1979; Miller 1997). Known as the Otavi Group sediments, these carbonate rocks were laid down between 730 and 700 Ma.

The spreading of the Rodinia plates, however, did not continue unchecked by the structural force of the earth's crust. Changes that occurred witnessed a reversal of the plate motions from spreading to collision, the closure of the ocean between the plates, and subduction of some plates beneath the others, which ultimately culminated in the formation of Gondwanaland (Miller 1983, 1997). The collision of plates brought about mountain building, consisted of carbonate rocks that have been previously deposited along the edge of these colliding plates. This mountain building process was particularly notable around the southern and western rims. Along with subsidence, the formation and uplifting of these mountain belts imprinted a basin behind them. That basin is what we call Owambo Basin today, and from thereon, it became a focus for sediment accumulations from topographically higher terrains in the surroundings.

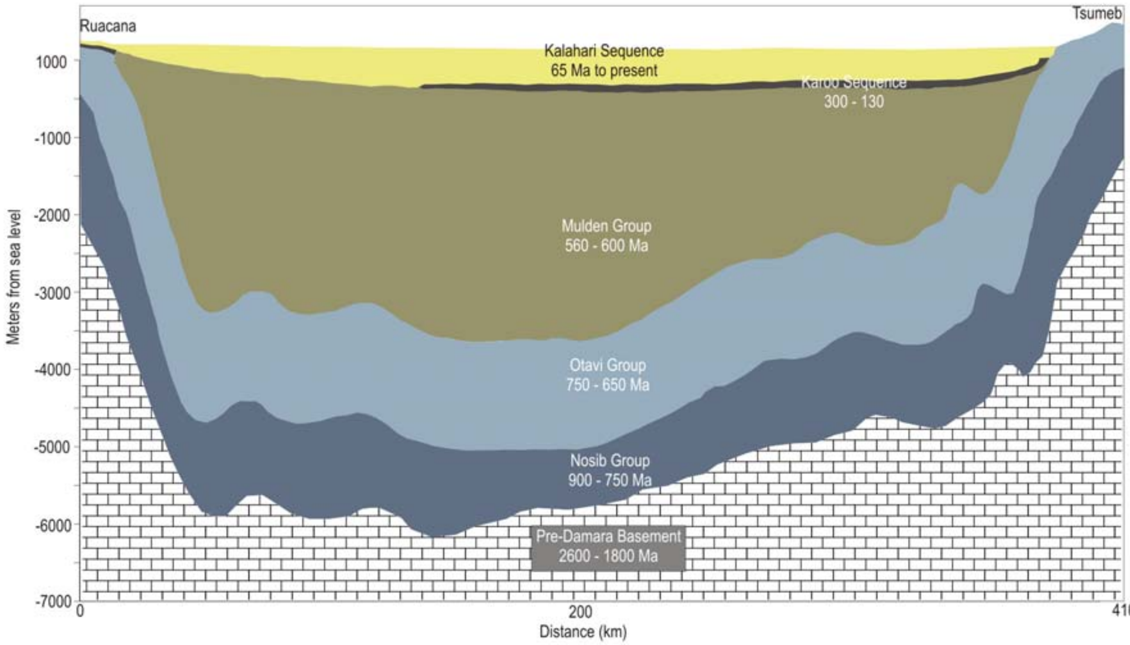

Figure 3.2 Geological cross-section across the Owambo Basin from Ruacana to Tsumeb. (Vertical exaggeration1:30; redrawn after Mendelsohn et al. 2000)

The process of sediment accumulation first started between 650 and 600 Ma with the commencement of attacking erosion of mountain belts located around the basin. The eroded materials, collectively known as the Mulden Group, were then deposited into the basin. Soon after, the Mulden together with the underlying deformed Otavi Group rocks were folded, tilted and uplifted. Much of this deformation gives the present character of elongated anticlines and synclines, trending mostly in an east-west and north-south direction along the southern and western margin, respectively, of the basin (Miller 1997).

Geochronological evidence reveals that the Damara Sequence is partially covered with a thin layer of Karoo strata, particularly in the centre and along the western rim of the basin (Figure 3.2). At around 300 Ma, there was a Gondwana-wide occurrence of glaciations, which also affected much of Namibia, including the Owambo Basin. These glaciers cut and eroded deep valleys starting from the western edge of the basin and flowed westwards to the continental margins of Gondwana (Martin 1981; Gilchrist *et al.*1994).

Thick sediments of tillite containing a wide range of clasts such as granite, quartzite, gneiss, vein quartz, siltstone, shale, dolomite, and limestone occur also in the centre of the Owambo basin. These rocks are known as the Dwyka Formation. By about 280 Ma, however, warmer climates prevailed gradually, which caused ice sheets and glaciers to melt and retreat. Sequentially, the increase of eustasy followed and caused the sea level to rise. This in turn resulted in marine transgressions impacting the basin. With seawater, carbonaceous shales, sandstones, siltstones and beds of organic material derived from plants were deposited in the basin. The Price Albert Formation refers to this sediment deposition (Miller 1997).

Between the Middle Permian to Lower Triassic, the basin lacks rocks that could be correlated with that period. Against the background of this hiatus, however, arid conditions continued to intensify. The period spanning the Late Triassic and Early Cretaceous is particularly understood to have been virtually characterised by desert-like conditions in much of Namibia. In the basin, this was deciphered from buried aeolian sandstone, which was correlated with the Etjo Formation and reaches a thickness of up to 137 m (Miller 1997).

Overlapping these latter events was a passive rifting of Gondwanaland which started in the Middle Jurassic. The ensued widespread and substantial denudation in the region was one of the major developments that exerted profound influence on the present geomorphological character of the sub-continent. The coupling of positive and negative feedbacks between tectonics and denudation set the stage for the step-wise establishment of the greater Kalahari Basin, in which the pre-existing Owambo Basin was to be incorporated. A contextual recount of these episodes is the subject of the following sub-section.

3.2.2 Structural, Sedimentary and Geomorphological Development of the Kalahari Basin

The horizontal separation of the Gondwana plates in south-western Africa started in Middle Jurassic times, as mentioned above. The rift zones and the interior of the subcontinent subsided, whilst a region sandwiched between the peri-cratonic offshore and the inter-cratonic basins was uplifted, forming a hingeline or flexural bulge (Summerfield 1985; Thomas & Shaw 1991). The presence of basins and the uplifted landmass favoured and realised substantial erosion from higher ground and deposition of the eroded materials into these basins on both sides of the hingeline. The application of Apatite Fission Track dating (AFT) in conjunction with cosmogenic isotope analyses revealed that the bulk of denudation in the southwest African margin occurred during the early post-rift period (Brown *et al.* 2000; Cockburn *et al.* 2000; Weber & Raab 2002; Raab *et al.* 2002). The denudation rates varied from as high as ~40 m/my (a total of about 5 km) for the early post-rift period (130-36 Ma) to ~5 m/my for the late post-rift stage. The total amount of denudation is generally regarded to have decreased considerable with distance from the coast to the continental interior (Van der Wateren & Dunai 2001). Even though such level of denudation was widespread across the region, it was less significant in the intercratonic basins. Over time, this transfer of materials resulted in isostatic adjustment of the crust, which acted as a catalyst for further uplift of the initial hingeline. Essentially, these events precipitated the presence and prominence of the Great Escarpment of southern Africa (Figure 3.3). Although earlier workers postulated that the Great Escarpment developed by virtue of backwearing (e.g. King 1967; Partridge & Maud 1987), recent studies concluded that the location of the Great Escarpment has at best only shifted a few kilometres inland (Brown *et al.* 2000; Cockburn *et al.* 2000; Kempf 2000; Weber & Raab 2002).

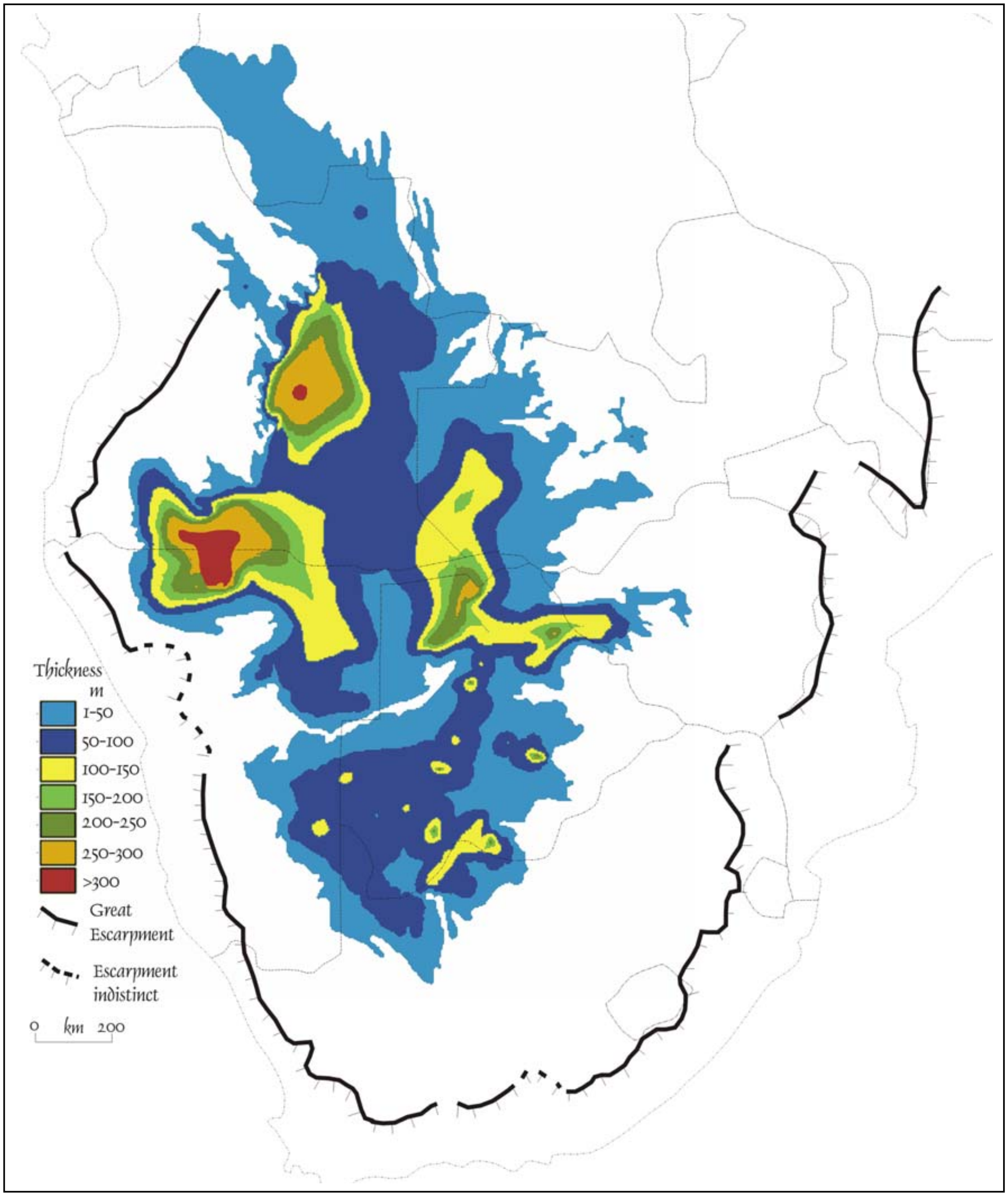

Figure 3.3: Approximate location of the Great Escarpment of Southern Africa and thickness distribution of the Kalahari Group sediments (after Thomas 1988; Thomas & Shaw 1991; Haddon 2000).

The intercratonic basin on the other hand developed steadily into the contiguous Kalahari Basin by virtue of progressive sediment deposition and redistribution. In due course, isolated basins such as the Owambo, which were already in existence, were absorbed and incorporated into what was to become the greater Kalahari Basin. Today, the Kalahari Basin as a geological entity comprises a surface area of 2.5 million km^2 (Cooke 1957), extending from western Congo and southern Gabon to the Orange River at its longest, and from north-central Namibia to western Zimbabwe at its widest (Figure 3.3). Sediments of this period, spanning the Cretaceous to Recent age and found in the entire basin are known as the Kalahari Group. However, because of its subcontinent-wide extent, the stratigraphic succession lacks temporal uniformity, and considerable local and regional variations are recognized. For these reasons, the content of the following sub-section will be limited to the sedimentological characteristics of the Owambo Basin, on the surface of which the Etosha Pan is located.

3.2.3 The Kalahari Sequence of the Owambo Basin

Four formations of the Kalahari sequence (Figure 3.2, 3.3, 3.4, 3.5) are recognised in the Owambo Basin. The Ombalantu Formation is the oldest, followed by the Beiseb, Olukonda and then the Upper Andoni Formation.

The Ombalantu Formation is made up of red, semi-consolidated but friable, variably silicified mudstones, almost entirely consisting of clay, whilst others contain an appreciable amount of silt and sand-sized grains (Miller 1997). Its upper part includes gypsum and casts of gypsum crystals. This formation is found in a broad, elongated area extending from the southeast to the northwest of the basin. Miller (1997) assessed its depositional setting to be that of a shallow, low-energy deltaic environment. The presence of gypsum suggests a restricted continental basin, with significant evaporation budget. The formation attains a maximum thickness of just over 100 m.

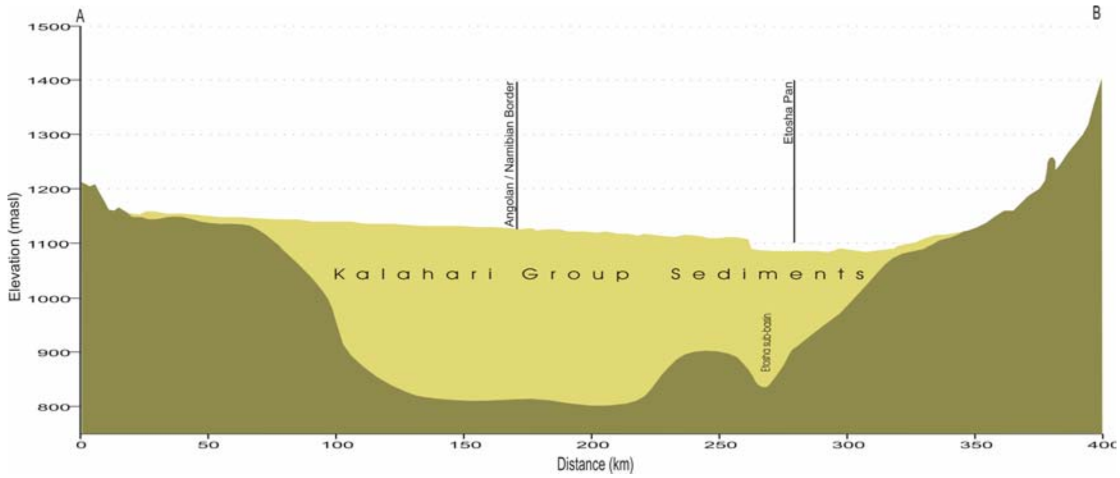

Figure 3.4: Cross-section of the Kalahari Group Sediments in the Owambo Basin along the 16°41' longitude (see Figure 3.6 for location) from the heels of the Encoco Highlands to the northern flank of the Otavi Dolomite Hills (modified from Kempf 2000).

The Beiseb Formation is the thinnest of the Kalahari sequence, with a maximum thickness of 50 m. Notwithstanding, it is the second most widespread formation across the basin. Mainly reddish in colour, but also light green to white near the present Etosha Pan, it is believed to represent a period of rapid and extensive sediment input. It consists of well-rounded sandstone and mudstone clasts as well as pebbles of various cherts up to 12 cm in diameter that are set in a matrix of fine- to medium-grained, argillaceous, calcareous to dolomitic sandstones. Silica or carbonates cement(s) these beds to variable hardness. Gypsum crystals up to 5 cm long have also been reported, occurring in the top 5 m of the formation (Hedberg 1979; Miller 1997).

The Olukonda Formation has a limited extent in the basin. It has an elongated distribution similar to that of the Ombalantu Formation discussed above. Reddish brown in colour, this stratum is friable and consists of poorly sorted and consolidated sand or sandstone. Few thin gritty and pebbly layers were also noted on its thickness of up to 175 m (Hedberg 1979; Miller 1997).

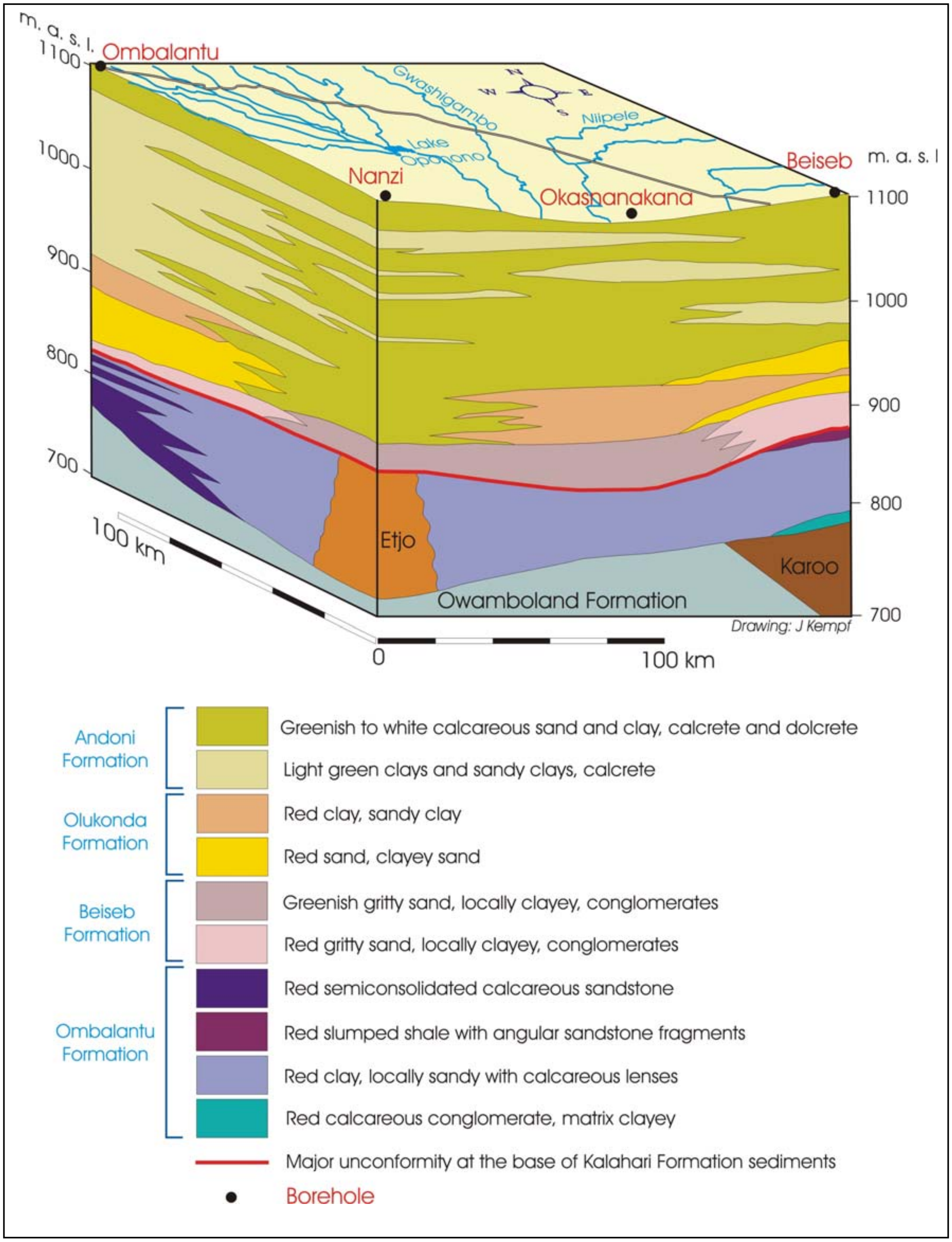

Figure 3.5: Schematic section through the Kalahari Sequence of the Owambo Basin based on the geology of the Beiseb, Okasnanakana, Nanzi and Ombalantu boreholes (modified from Miller 1997).

Within the Kalahari Sequence, the Upper Andoni Formation is the most pervasive, covering all the antecedent formations. It consists of interbedded white medium-grained sand, light greenish clayey sand and green clay. A predominantly sand layer located in a zone between 10 and 200 m thick is unconsolidated, somewhat pyritic or hematitic. The upper part of this section contains a number of small, irregularly shaped dolcrete and calcrete nodules, up to 30 cm across. Nearly 90% of the sand is quartz. These grains are polished and frosted, angular to sub-angular, and sorting improves upwards. The rest of this material consists of chalcedony, feldspar and chert (Hedberg 1979; Miller 1997).

Sand layers that are interbedded with clay range between a few centimetres to 155 m in thickness. They are also noted for being pyritic, coupled with calcareous input. Thin limestone layers up to 10 cm thick have also been reported, whilst within the environs of present-day Etosha Pan, oolitic layers of similar magnitude and fresh-water ostracod shells and their impressions were encountered (Hedberg 1979). More than half of this formation is reported to consist of olive green clay, silt or sandy clay (SACS 1980; Miller 1997; Buch & Trippner 1997); it is not clear whether the dimension mentioned refers to lateral or horizontal extent or both. The present Etosha Pan rests on a section containing more than 25% of clay (Miller 1997). On top of the Andoni Formation, there is a widespread surficial layer which is believed to be aeolian sand of Recent age (Buch & Trippner 1997; Miller 1997).

The age of the Kalahari sediments in the Owambo Basin (Figure 3.4, 3.6) is unknown. Sauer (1966), basing his estimate on fragments of *Struthio* eggshell found at 44 m depth in the Beisebvlakte, proposed the oldest possible age of these sediments to be of Upper Tertiary to Middle Pleistocene. Yet Buch (1996) assumed a Pre-Miocene age for the Andoni Formation, the youngest unit of the sequence. Moreover, Miller (1997) assigned a Cretaceous age to the Ombalantu Formation, the basal layer of this sequence. More recently, in studying the climatic geomorphology of central Namibia, Kempf (2000) deduced a middle Tertiary to Quaternary age for the Kalahari sediments.

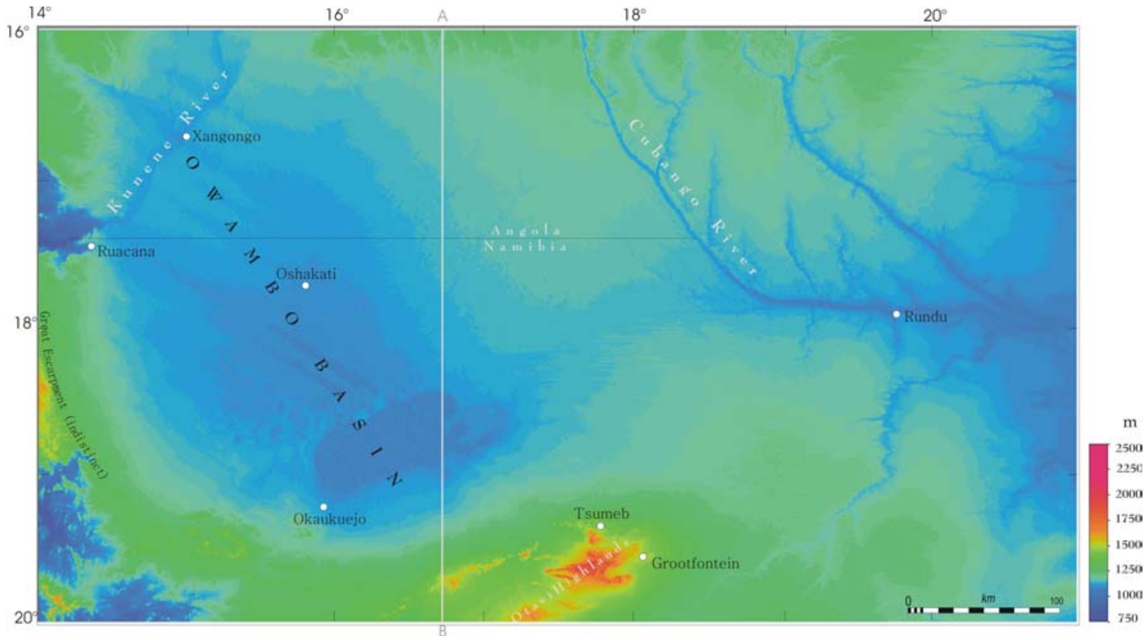

Figure 3.6 Digital Elevation Model ((DEM) 90 m ground resolution) of the Owambo Basin surface. Areas with low elevation (blue shades) at lower left are part of the gorges of the ephemeral Huab (bottom) and Hoanib (upper) rivers. The water divide between the drainage of the Etosha basin and those of the two respective ephemeral rivers is marked by a rim (in green to red shades) comprising the Otavi Group sediments. The Kunene and Cubango catchments are ill-defined with respect to that of the Etosha basin. (Data source: NASA)

Overall, this Kalahari Sequence is thought to have been deposited under an arid to semi-arid environment. Sediments were brought into the basin by seasonal waters flowing in from the north and northwest. In the process shallow, semi-permanent lakes developed on the surface of the basin. The varying spatial distribution of strata and differences in composition within strata are taken to suggest that the centre of the depression has been shifting with time from formation to formation. Thus, the first lake of the Kalahari time sediments was restricted to the central part, which is traceable through the distribution of the Ombalantu Formation. A basal conglomerate and aeolian sands of this formation define the marginal area of that lake (Hedberg 1979; SACS 1980; Miller 1997).

A moderately wetter period followed afterwards and coarse detritus from the former lake margin of the Ombalantu times was washed into the lake, resulting in the lateral widening of the previously restricted basin. These materials, mainly made up of gritty to conglomeratic sandstone, and others such as clays that were deposited in the lake interior, became the Beiseb Formation. The greenish colour of much of these materials was a result of glauconite. Typically, glauconite is suggestive of slow rates of sediment accumulation and may form by biogenetic or diagenetic alteration of minerals such as biotite in shallow water under reducing conditions (Barthelmy 2004; Gordua 2004). The red Olukonda Formation was deposited to the north, east and northwest where oxidising conditions were still in existence at that time. The pervasiveness of the subsequent green Andoni clays implies that with time, reducing conditions spread throughout the whole Owambo Basin (Miller 1997).

3.3 Drainage Evolution

The development and evolution of the drainage systems in the sub-continent are firmly rooted in the post-fragmentation of Gondwana. Dual river systems, namely, endorheic and exorheic came into being during this era. The presence of the hingeline, which later became the Great Escarpment (discussed in section 3.2.2) played a significant role in the development and modification of these river systems. Relatively short rivers with steep gradients drained from the hingeline to the coast, whilst longer but lower gradient rivers drained from the opposite side of the bulge into the interior Kalahari Basin (Thomas & Shaw 1988). It is these rivers that are generally hold responsible for sediments transfer from the highlands into the basin (Thomas & Shaw 1991). With time and neo-tectonics during the Pliocene and at the dawn of the Quaternary, however, the dynamics of most of these rivers went through significant alterations and adjustments. This is particularly so for the south-flowing rivers, such as the Okavango, proto-Upper Cunene, and Zambezi (Allanson *et al.* 1990).

The Okavango, enticingly dubbed by Kanthack (1921) as a “sister” river to the Kunene, along with its associated systems such as the Okavango Swamps, will be discussed below to illustrate these alterations and adjustments. This rationalization is accompanied by recurring comparisons and contrasts of the two major pans in the sub-continent, the Makgadikgadi and Etosha (Figure 3.1). In turn, this will help ferment a salutary and poised view of pertinent fluvial geomorphology of the subcontinent and widen the horizons in which the Etosha Pan will be examined.

3.3.1 The Okavango System

The Kavango is an endorheic, perennial river draining parts of central and eastern Angola and terminating in the Okavango Delta[i] in Botswana. Spelled Cubango in Angola, it rises on the crystalline Benguela Plateau of eastern Angola, west of the section of the watershed known as the Kalahari Col (Thomas & Shaw 1991). The Kalahari Col is positioned between higher topography of the Karroo granites in the west and the Precambrian Katanga Group rocks in the east (Dixey 1943). There, the catchment receives an average annual rainfall of approximately 1000 mm (Thomas & Shaw 1991; Mendelsohn & El Obeid 2003).

The Okavango is the third largest river in southern Africa and has a headwater basin of 121,700 km^2. Its mean annual discharge is approximately 9.2 billion cubic meters, with a considerable seasonal variation (McCarthy *et al.* 1998a; Mendelsohn & El Obeid 2004). From its source, it flows about 300 km south-southwest before entering the Kalahari Basin. In Angola, the Cubango is gently incised into the Quaternary Kalahari sand cover, with greater incision of up to 30 m occurring sporadically, exposing the sub-Kalahari geology. The river is shared between Angola and Namibia for about 400 km as a boundary between the two states. Its major tributary, the Cuito, joins it at Dirico.

[i] The term "delta", although used frequently in the region, is a misnomer, for the Okavango swamps is indeed an alluvial fan. It is nevertheless used in this study for the sake of regional allegiance, and sometimes interchanged with swamps, fan or fan delta.

Both the Cubango and Cuito show a longitudinal profile steeply dropping in the upper reaches when compared with the rest of their respective course (Figure 3.7). The Cuito contributes about 45% of the total water volume of the Kavango (Mendelsohn & El Obeid 2003).

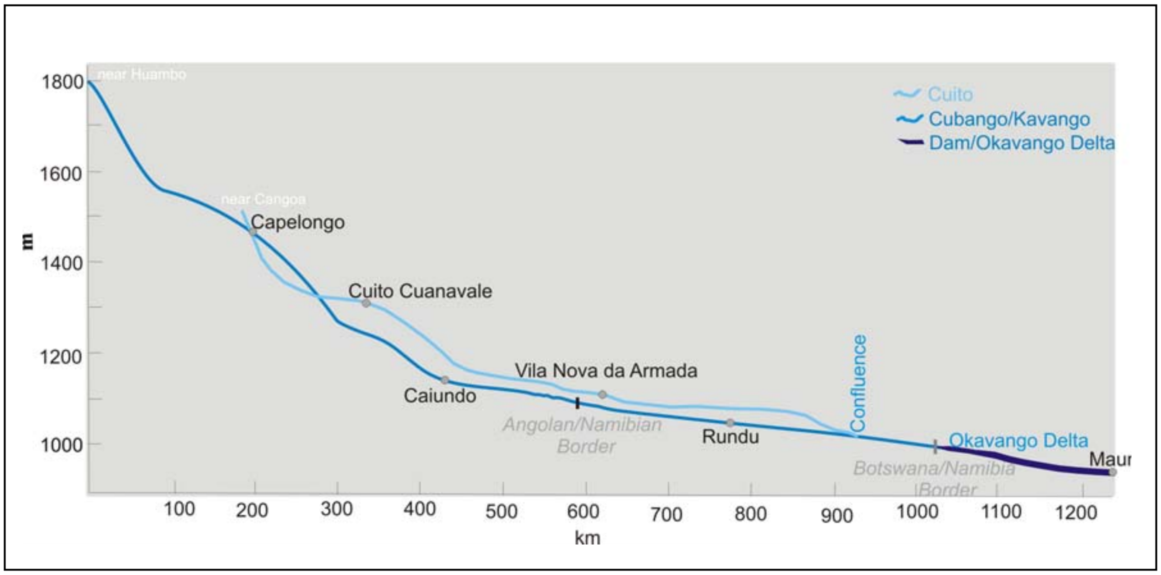

Figure 3.7: Longitudinal profile of the Cubango/Kavango River and its main tributary, the Cuito (Redrawn after Mendelsohn & El Obeid 2003). Vertical exaggeration 1:500.

As the river leisurely flows southeast from Mukwe, it enters a 40 km section of rapids at Popa Falls. These rapids, having a height of 5 m, cross the Damara Sequence consisting of folded quartzite with intercalated limestone (Thomas & Shaw 1991; Siegfried & Schneider 2004). From there, the river continues in a south-easterly direction and enters the (neo)tectonically controlled 'Panhandle' of the Okavango Delta, before it spills its collective load into the wider fan delta. This fan delta is one of the components of a much larger basin, the Lake Palaeo-Makgadikgadi. The following sections will take a closer look at some of these members.

3.3.1.1 Lake Palaeo-Makgadikgadi

Lake Palaeo-Makgadikgadi is a large basin delineated by the present 1000 m contour (Cooke 1979). At its maximum size in excess of 60,000 km^2, it occupied northern Botswana and centred on the modern day Makgadikgadi Pans. Its extent included the Mabane Basin, the Okavango Graben and much of the upper part of the Zambezi trough (Figure 3.8) (e.g. Cooke 1980). Today, these elements are tenuously connected via near-defunct river channels. This palaeo-lake is controlled by post-Gondwana faults, which continue to show seismic activity to the present (Reeves 1972a; Reeves 1972b; McCarthy *et al.* 1993b; Modisi 2000). The largest recorded event with an epicentre located in the centre of the Okavango fan-delta occurred in October 1952 and registered 6.7 on the Richter scale (Reeves 1972b).

It is believed that during the upper Jurassic to Cretaceous, the proto-Upper Zambezi, Okavango, and Cuando rivers found their way to the Indian Ocean via the Limpopo (Wellington 1955; Bond 1963; Moore & Larkin 2001). During the Pliocene, however, there was a flexuring of the Kalahari-Zimbabwe axis and further downwarping of the Kalahari Rift (Cooke 1979; Thomas & Shaw 1988), which interrupted these drainage flows towards the Limpopo. As a result, a huge basin of internal drainage, Lake Palaeo-Makgadikgadi fed by the Zambezi and Okavango rivers, was effectively created (Grey & Cooke 1977; Cooke 1979; Helgren 1984; Shaw & Thomas 1992). Subsequent neo-tectonics and the downwarping of the Gwebe and Chicoa troughs deflected the upper Zambezi into the middle Zambezi (Cooke 1979; Shaw & Thomas 1992), which in the process reduced Lake Palaeo-Makgadikgadi's water budget. On the basis of current geomorphological dynamics and configuration of the area, the modern Okavango River, now draining into the Okavango delta (section 3.3.1.3), is ultimately anticipated to be diverted as well into the Zambezi River (Wellington 1955; Moore & Larkin 2001).

With a down-faulting to the north-west, the north-east trending Kunyere and Thamalakane Faults, located some 110 km upstream of the present day Makgadikgadi Pans, imparted a slow but serious interference to the downstream flow of the Okavango River (Cooke 1979; McCarthy *et al.* 1997). This impinged flow helped create an alluvial fan of very shallow slope that we call the Okavango Delta today (Baillieul 1979; Cooke 1979; Allanson *et al.* 1990).

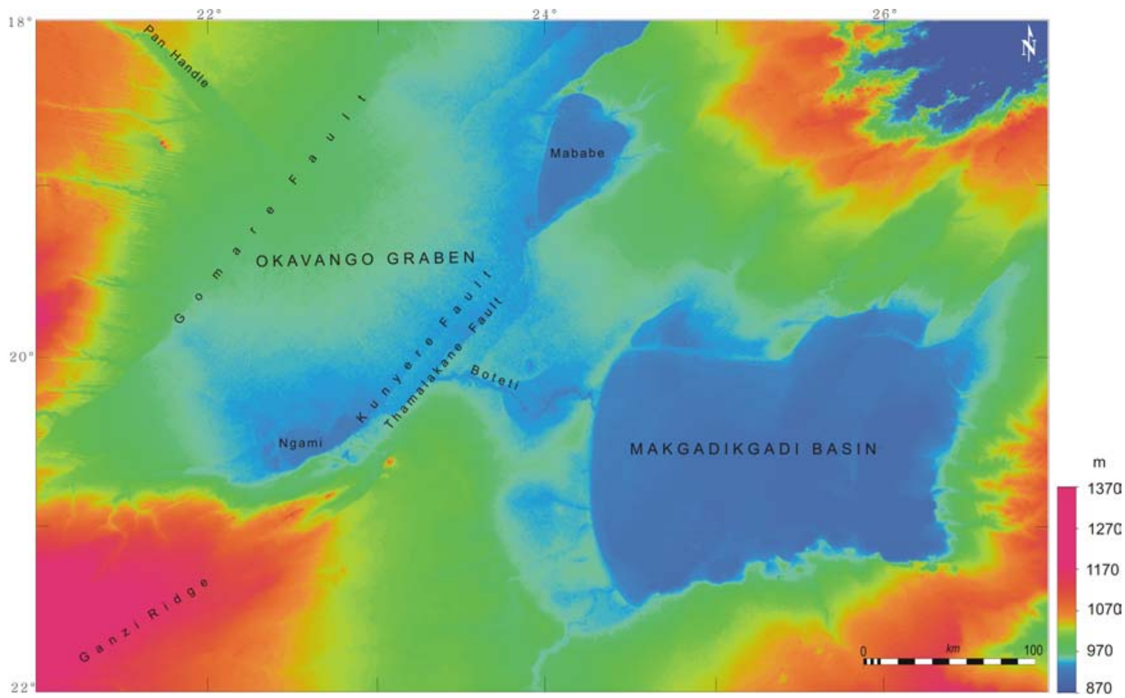

Figure 3.8: Digital Elevation Model (90 m ground resolution) of the Okavango Graben and Makgadikgadi Basin. The Okavango River enters the fan delta via the "pan-handle", top left. The Boteti River, which links the Okavango Delta and the Makgadikgadi Pans, used to be perennial until 1973 when it ceased to flow as a connected stream in successive years (Cooke 1980). It breaks out from the Okavango trough at a height of 931 m and enters the pan at a height of 900 m. (Data source: NASA)

Increased aridity served as a final blow to the desiccation of the Lake Palaeo-Makgadikgadi, leaving behind the Makgadikgadi pans as remnants of this vast endorheic lake (Grey & Cooke 1977; Baillieul 1979; Cooke 1979; Cooke 1980). Since there is no outlet identified for the lake, it is generally assumed that a static state was reached where inflow was balanced by evaporation. Some of these events are still inscribed to the characteristics and sediments associated with the contemporary Makgadikgadi Pans. A brief look at this landform will thus serve to place these events into perspective.

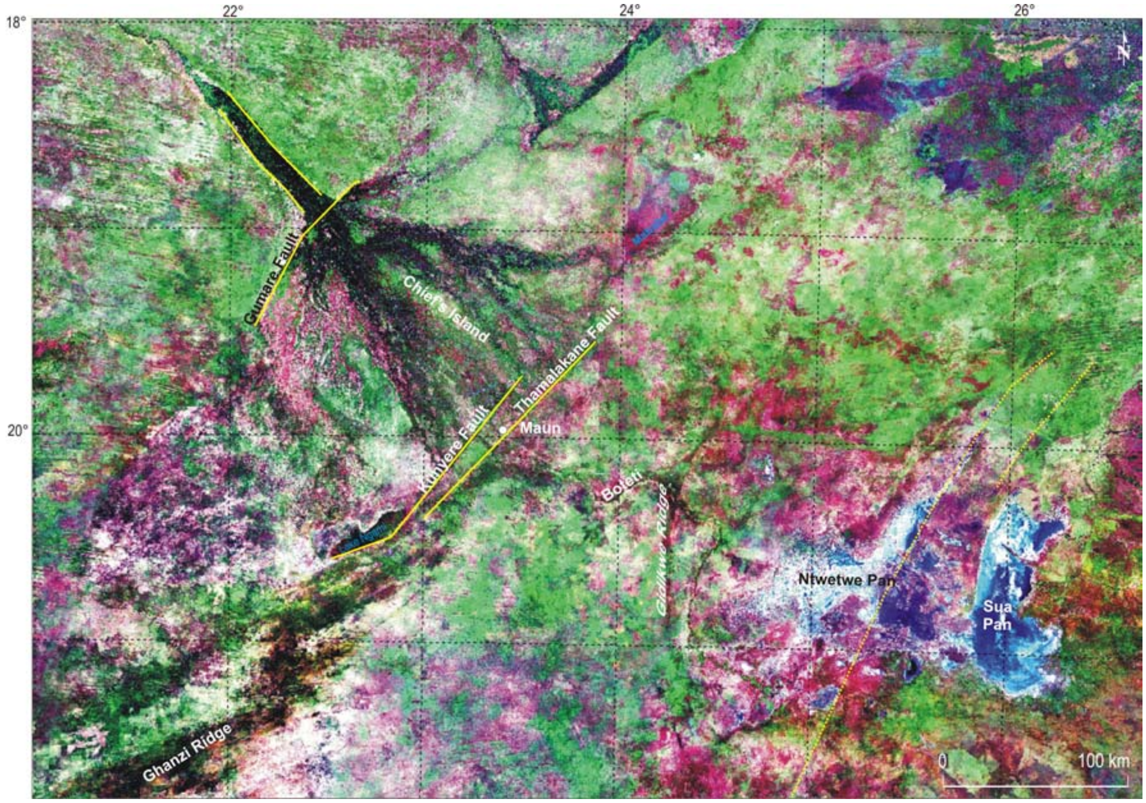

Figure 3.9: Satellite Imagery (Landsat TM) covering the Makgadikgadi Basin and Okavango Graben. The down-throw of the Kunyere and Thamalakane Faults is to the delta side and the faults now form scarps of 2 m and 12 m, respectively. The dark delta-shaped feature is the Okavango swamps.

3.3.1.2 The Makgadikgadi Pans

The present-day Makgadikgadi Pan is practically a complex of shallow basins occupying the lowest point of a drainage system extending from Botswana into Namibia, Angola and Zimbabwe (Figure 3.8) (Grey & Cooke 1977; Baillieul 1979). Of the main 11 pans that make up this complex, the Ntwetwe to the west and Sua Pan to the east are the most salient members (Figure 3.9). The sump level of these pans lies in Sua Pan, with a height of 890 m above sea level (Cooke 1979). Their near-flat surfaces are covered with saline clays and salt efflorescences, overlying up to 100 m of inter-layered clays and sands (Thomas & Shaw 1991), which in turn rest on Karroo bedrock (Cooke 1980).

The Makgadikgadi is regarded as a vestige of the Lake Palaeo Makgadikgadi described in the preceding section. It is fed by the Botletle or Boteti River which has a conveyance capacity much in excess of present needs. This river used to be perennial, but as recently as 1973, discontinuous flow has been recorded for several years in succession. It enters the Makgadikgadi Pans from the northwest, whilst a number of other rivers flow from the west, south and east (Cooke 1980). Draining from the east, i.e. from the higher lands rising towards the uplifted Kalahari – Zimbabwe axis, are the ephemeral Mosupe, Lepashe, Mosetse, Semowa, and Nata rivers (Cooke 1979). From the south, extensive abandoned valleys, such as the Okwa and Mmone drained from the crescent of higher land which runs from eastern Namibia into the pans (Cooke 1979; Cooke & Verstappen 1984; Shaw & Cooke 1986). A series of fossil fan deltas has been deposited on to the pan, particularly by the Boteti River (Thomas & Shaw 1991).

Three parallel, gently curved ridges lie approximately 945 m, 920 m and 912 m above sea level. The 945 m and 920 m ridge levels are the most conspicuous, occurring more distinctly in the western part of the pans and known as Gidikwe and Magwikwe, respectively. The former runs for more than 200 km around the western rim of the pans, while the latter forms a gently curving arc about 70 km long, on the same side of the pans (Grove 1969; Grey & Cooke 1977). The 945 m Gidikwe ridge has a relative relief of about 10 - 15 m above the adjacent plain. The 912 m ridge, known as Dautsa, is usually marked by a low sandy rise, whilst the sand of the Magwikwe is frequently capped with a layer of calcrete. A layer of calcrete caps the Gidikwe Ridge as well, which has also a thick bed of well-rounded or flat smooth gravels and cobbles that occur in places. Ringrose *et al.* (1999) assumed that these layers of calcretes were formed when calcium carbonate precipitated mainly in marshy conditions around plant roots and stems and in association with bacteria in embayments along the lakeshore. Over time and with increasing lake levels, calcrete thickened and became partially indurated. Such a process is comparable to modern-day accumulation of calcium carbonate in the neighbouring Okavango swamps (e.g. McCarthy & Ellery 1998) (section 3.3.1.3). However, Ringrose *et al.* (2005) reviewed that stance, by recognising and separating biogenic-related calcrete from climatically induced precipitation in littoral pore waters during drying phases. Silicification episodes were thus thought to have post-dated early relatively fresh-water palaeolake stages, while subsequent calcretisation in multiple stages occurred after palaeolake events as a result of watertable fluctuations in a semi-arid environment.

Beach deposits, consisting of poorly sorted flat ellipsoidal and discoidal pebbles of polygenetic origin, varying in size from about 1 cm to 15 cm, were found in places, particularly at the 945 m and 920 m levels in the southern side of the basin (Grey & Cooke 1977; Cooke 1979). Based on the relatively gentle slope on which they lie and because high storm beaches are often of a well-sorted nature (King 1972), Grey & Cooke (1977) concluded that the gravel beds represent a low level storm beach. These ridges are thus believed to be barrier beaches formed offshore by wave action as evidenced from sediment characteristics and augmented by the presence of bivalve shells, particularly the Ostracoda species (Cooke 1980). The supply of sand is thought to have originated from longshore or offshore transport or both (Grey & Cooke 1977; Cooke & Verstappen 1984). These ridges variably act as a barrier to channel inflow and facilitate ponding behind them. Their multiple-presence is interpreted to point to the climatic changes that caused the lake to fluctuate in size (Grey & Cooke 1977; Ebert & Hitchcock 1978).

Palaehydrological studies employed calcretes, peat, shell, sand and lithified wood dating to vouchsafe the stages and time-frames of the lake levels (e.g. Heine 1978, 1982; Cooke 1984; Cooke & Verstappen 1984; Shaw 1985, Shaw *et al.* 1997; Ringrose *et al.* 1999; Ringrose *et al.* 2005). From around 2 Ma to 35 ka, lake levels are thought to have stood at the 945 m, some 55 m above the present sump level of Sua Pan. At this level, the Makgadikgadi would have been confluent with a large body of water occupying the present lower Okavango Swamps together with the Mababe and Ngami basins. Low-level lake conditions are deciphered to have returned at around 35 - 26 ka. Between 26 - 10 ka, the Makgadikgadi was occupied by a succession of lakes at the 920 m level, interspersed with reduced levels at around 21 and 19 ka. Fluctuations between desiccation and the 912 m level reigned during the last 10,000 years, with pronounced 912 m levels identified at 3,000 and 1,700 years before present (Shaw & Cooke 1986).

3.3.1.3 The Okavango Swamps

The Okavango swamps occupy a slowly subsiding graben between the northeast-southwest trending Chobe-Gomare faults and the Thamalekane - Kunyere fault zone (Figure 3.8) (e.g. Grey & Cooke 1977). Throws along the Thamalekene - Kunyere faults are reported to be in the order of 300 m, a depth of the same order as the sediment thickness of Recent age (Grey & Cooke 1977; Reeves 1978). This suggests that sedimentation has kept pace with subsidence (Reeves 1972a). Presently, the Thamalekene and Kunyere form a 12 m and 2 m scarp, respectively, located on the upthrown-side (southeast) of the faults (McCarthy *et al.* 1997). The graben is essentially an extension of the East African Rift system (McCarthy *et al.* 1993a) and is mantled with unconsolidated aeolian sand (McCarthy & Ellery 1995) subject to reworking in alluvial fan environments (McCarthy *et al.* 1988).

Two main components, the upper, linear Panhandle section and the lower, delta-shaped alluvial fan, make up the Okavango Delta. The contemporary size of the delta fluctuates with seasons. About 6,000 km^2, including an active peat-forming environment, are permanently inundated, whilst up to 13,000 km^2 of its surface is a seasonal floodplain. The total area of swamp sediment, however, is in excess of 22,000 km^2 (Cooke 1980; Shaw 1988a; Andersson *et al.* 2003). The fan has an annual water budget of about 15.5 x 10^9 m^3 yr^{-1}, of which 10.5 x 10^9 m^3 yr^{-1} is inflow, while the remainder comes from local precipitation (McCarthy *et al.* 1998a; McCarthy & Hancox 2000; Andersson *et al.* 2003).

The climate of the area is hot and dry (Anderson 1976). The average annual rainfall of 513 mm that the area receives is exceeded by the evapotranspiration of 1,400 mm (Cooke 1980; McCarthy *et al.* 1998a). About 96% of its annual budget is thus lost to evapotranspiration, the remainder to outflow, mainly through the Boteti River that leads into the Makgadikgadi Pans (section 3.3.1.2) (McCarthy *et al.* 1998a). Despite the high evapotranspiration loss, however, saline surface water is exceptionally rare in the fan (McCarthy & Ellery 1995). McCarthy & Ellery (1998) suggested that this general absence of saline surface water implies that transpiration far exceeds evaporation.

Rainfall in both the fan and its catchment falls in December to February and inflow into the Delta peaks in March. Because of the gently sloping gradient of the Delta, averaging 1:3,400 (McCarthy *et al.* 1997), each seasonal flood, at a flow-speed of 5 cm per second, takes four to five months to transverse the 250 km across the delta. This is in addition to impinging vegetation, groundwater recharge and the undulating topography, which causes water to fill deeper depressions first before the flood can advance. Collectively, these factors delay the outflow from the fan which peaks in July or August when the maximum inundation of the seasonal swamps takes place (McCarthy *et al.* 1997; McCarthy & Ellery 1998; McCarthy *et al.* 1998a). Water is generally shallow, having a depth of one to two meter (Grey & Cooke 1977; McCarthy *et al.* 1997).

The area around the fan is characterized by relict eolian features, such as linear dunes in the west and south. Transverse to barchan dunes are found in the east. Although the dunes support woodland communities and are heavily degraded, locally their relief may exceed 20 m (Thomas & Shaw 1991). The floodplain itself is devoid of aeolian features, although as mentioned earlier, the soils of the fan are typically very sandy, their size distribution reflecting eolian sands of the region (McCarthy & Ellery 1995) and were subjected to reworking in alluvial fan environments (McCarthy *et al.* 1988).

Likewise, the sediment load entering the delta is dominated by bedload consisting of fine aeolian sand (McCarthy *et al.* 1991a). Due to low local topographic relief and gentle regional slope (i.e. Gumbricht *et al.* 2001), however, its annual 170,000 tonnes of bedload generally accumulates in the channels crossing the permanent swamps, causing vertical aggradation that may exceed 5 cm per year in places (McCarthy *et al.* 1992), with only minor point bar formation (McCarthy *et al.* 1993). As a result, the velocity of channel flow decreases and encourages the invasion of giant sedge, *Cyperrus papyrus*. Over time, this culminates in channel abandonment by avulsion (McCarthy *et al.* 1992). Subsequently, water diversion is introduced and new channels are formed (McCarthy & Ellery 1998c). Areas of aggraded channel beds on the other hand become low, alluvial sandy ridges. This process is believed to be constantly changing, with a channel lifespan of 100 years under present flow conditions. Frequent switching of channel systems evidently results in a very even distribution of sediment across the delta surface (McCarthy *et al.* 1988).

Suspended load is generally low, amounting to some 39,000 tonnes and consisting of organic debris (McCarthy *et al.* 1991; McCarthy & Ellery 1998). In comparison with major rivers of the world (Milliman & Meade 1983) the suspended load of the Kavango River is the lowest recorded thus far (McCarthy *et al.* 1992).

Whereas clastic sedimentation dominates the upper part of the Delta, chemical sedimentation is significantly concentrated in the lower section (McCarthy & Metcalf 1990; McCarthy & Ellery 1995). This is primarily a consequence of the low suspended load of the Okavango River, the very low gradient of the fan and its dense aquatic vegetation (McCarthy *et al.* 1991b). An estimated 380,000 tonnes of solutes are dispersed over the delta each year. Silica is the main constituent (39%), with calcium and magnesium carbonate occurring in lesser amounts. Since most of the water is lost to evapotranspiration and outflow is significantly low, only a small proportion of this load leaves the Delta. About 95% of the solute load accumulates in the Delta (McCarthy & Ellery 1998).

Chemical sedimentation is usually most prevalent on local topographic highs, such as termite mounds and sandy ridges that resulted from channel aggradation and abandonment (McCarthy & Ellery 1994; McCarthy *et al.* 1998b; McCarthy & Hancox 2000). This is mainly influenced by transpiration as the mechanism of water loss, given that these areas contain dense vegetation cover, supporting broadleaved evergreen trees. Plants are known to be very selective in the solutes that they remove from the water. As transpiration occurs from the vegetation, water moves from the surface into the root zone. Ultimately, the selective extraction of resources and the pumping of water into the atmosphere through transpiration lead to the increase in concentration of dissolved solutes in the groundwater. On islands, this process continues all year round, even when the surface water of seasonal swamps has dried up. With time, saturation of calcium and magnesium carbonate occurs beneath island fringes, and salts precipitate out of solution as magnesium calcite (McCarthy & Ellery 1998). The net result of this process is a differing degree of saline brines, which is largely concentrated beneath topographically high features, recharging the reported extremely saline brines that underlie the Okavango region at depth (McCarthy *et al.* 1998a).

In summary, the Okavango swamp is practically a consequence of regional climate, the presence of an internal drainage system and rifting. Within these bounds, the biological processes helped shape the present character of this environment. Despite all this detailed understanding of the fan and the processes involved in its existence, however, the age of the fan is still unknown (McCarthy & Ellery 1998).

As we will see in subsequent chapters, there are a number of similarities between the Etosha Pan and prominent members of the Okavango system, such as the Makgadikgadi Pans and the Okavango Delta. For example, a variety of lacustrine, fluvial, and aeolian landforms in close juxtaposition are a norm in these basins.

3.4 Climate of the study area

The climate of the area immediately around Etosha Pan is largely semi-arid. This climatic characteristic is essentially a function of the global atmospheric circulation patterns which affect southern Africa in association with the elevation and continental location of the area. The development of the atmospheric high pressure, particularly during summer, is being disrupted by the sub-continental landmass, which tapers meridionally southwards. This atmospheric condition is compounded by the continental dimension of the area introduced through the elevation and interior position of the wider Kalahari Basin. It is not surprising therefore that although the Owambo Basin lies above the tropic of Capricorn, its altitude imprints in many respects a more temperate than tropical climate (Tyson 1986; Thomas & Shaw 1991).

Rains, which is highly variable in terms of quantity, timing and coverage, falls in the summer months when temperatures are highest. Like in most of the summer rainfall regions in the subcontinent, rainfall is regarded as of convective in nature, being modulated by the diurnal heating of the near-surface air and related atmospheric and topographic effects. The considerable east-west rainfall gradient of the area is particularly influenced by the interplay of the easterlies forcing and cumulus convection (Tyson 1986). Most storms over the region move in westerly to south-westerly tracks. This translates into the fact that the east-lying areas receive earlier and more rainfall than the west (Mendelsohn *et al.* 2000). The area drained by the Omuramba Owambo, for example, receives an annual average rainfall varying between 550 and 600 mm. The opposite end of the rainfall gradient receives an annual average rainfall of approximately 300 mm. However, due to its climate and sandy nature, much of its falling water budget evaporates rapidly or seeps into the ground (Mendelsohn *et al.* 2000). The mean potential evaporation for the rainy season is estimated at 2600 – 2800 mm, which exceeds rainfall by a factor of 8.4 (Engert 1997).

Although all four seasonal contrasts of the year are perceptible, Berry (1980) distinguished only three climatic periods based on precipitation and temperature stratification. A wet and hot season starts in January and last until April. A dry and cold season follows from May to August, before the dry and hot season commences in September and lasts until December.

Wind data at Okaukuejo between January 1982 and January 1991 were analysed by Kempf (2000). As *Figure (3.9)* depicts, in all seasons dominant wind, amounting to an average of 43%, comes from the north-east. During the dry season, the wind speed of the north-easterlies may exceed 20 m/s at Okaukuejo, up to a frequency of five days a month (Berry 1980; Kempf 2000). Some of these wind storms have been captured through space-borne cameras. Mendelsohn *et al.* (2000), for example, presented a satellite imagery taken in August 1999 showing a dust plume off the Etosha Pan and extending some 300 km towards the coast. The north and east are the next directions from which the prevailing wind blows (Engert 1997; Kempf 2000). In addition, Berry (1980) observed at times a distinct pattern of wind changes from the southwesterly/southeasterly direction during the early hours of the day to northeast around midday, before reverting to the previous vector(s) in the evenings.

June and July are the coldest months at Okaukuejo. During these times minimum temperatures drop to about 7 °C, with temperatures close to zero being also recorded. Day temperatures during the same period rise to about 17 °C. Conversely, December and January are the hottest months, with an average mean temperature of approximately 26 °C. Maximum and minimum average for summer is 35 °C and 18 °C respectively (Van der Merwe 1983; Engert 1997; Mendelsohn *et al.* 2000). These considerable temperature variations between day and night are attributed to the continental climate of the area.

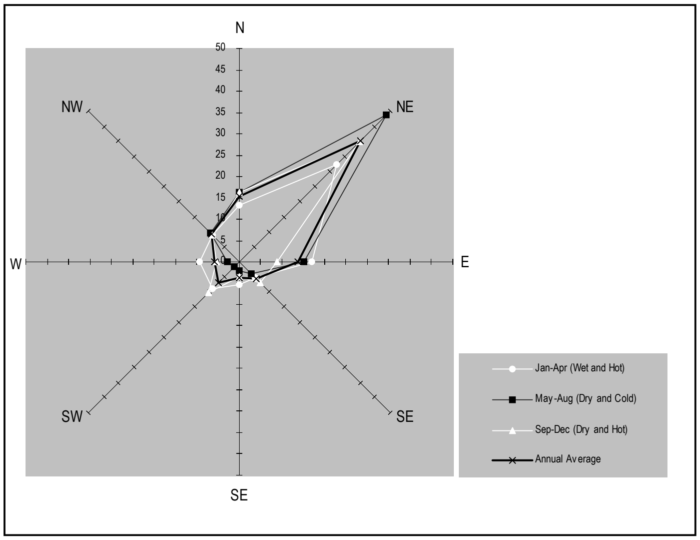

Figure 3.10: Wind rose at Okaukuejo from 1982 to1990 (after Kempf 2000).

The Etosha Pan has been listed as one of the eight large basins of internal drainage in the world noted for contributing to the mineral dust in the atmosphere (Washington *et al.* 2000, 2003; Byrant 2002, 2003). Coupled with tentative evidence of a possible positive feedback mechanism involving rainfall and dust variability (Brooks & Legrand 2000) and the documented higher rainfall anomaly in the leeward of Etosha Pan (Hipondoka *et al.* 2004b), it is plausible that the pan plays a significant role in regulating the micro-climate of the area.

3.5 Soils

Detailed mapping of soils in Etosha National Park was carried out by Beugler-Bell & Buch (1997) based on the locally adapted FAO system. Eight different major soil units were identified. They span the categories of *Arenosols, Regosols, Cambisols, Leptosols, Vertisols, Fluvisols, Solonchaks* and *Solonetzs*. Their distribution and properties show strong influence from the underlying geology and relief. In addition, most soil profiles reflect two main geomorphological processes, namely fluvial and aeolian. These soils are typically constituted of an ochric A-horizon, with a low to moderate content of organic matter. They have weak to moderate profile development, high base saturation and a pH which is slightly acidic to alkaline. Soluble salts and carbonates are known to accumulate in the lower topsoil and subsoil at places.

In detail, soils of the Etosha Pan are mapped as *calci sodic Solonchaks* to *sali calcic Solonetzs* derived from Andoni sandstone or siltstone. *Stagnic Solonetz* with a structural B-horizon occurs at Andonivlakte. Its genesis is uncertain, but it is speculated to have developed either directly over the Andoni Formation or over a thin cover of fluvial sands resting over the Andoni Formation, the parent bedrock of the former pan's surface. Soils of other pans in the area are similar, except that their origin is traced to sandy Etosha limestone (calcrete).

Soils from areas surrounding the Etosha Pan in the north-east, north and north-west are free of carbonate. The soils there, of fine to medium sandy texture, are classified as *cambic* and *ferralic Arenosols* developed in deep sandy substrata, some of which are partly fluvial in origin. Like the islands in the Etosha Pan, the area along the western and north-western edges of Etosha Pan is dominated by (hyper) *calcaric Regosols* to (hyper) *calcaric Arenosols* and *aplic Calcisols* known to have been derived from calcareous sediment above the Etosha limestone and Andoni sand- and siltstone. They are rich in calcium-carbonate, particularly with decreasing distance from the pan.

The area south of Etosha Pan is characterised by the pediment zone of the Otavi dolomite hills. It has extremely shallow to shallow clayey-loamy soils developed from limestone. These soils are mapped as *lithic, eutric* or *rendzic Leptosols* and *eutric Vertisols,* the latter being more common in depressions. Judging from solum depths of 5 to 20 cm in this area, Beugler-Bell & Buch (1997) concluded that the rate of soil formation there is in the order of 2.5 to 10 mm per 1,000 years.

Fluvisols in the park are limited to omiramba and pans. In the environs of Etosha Pan, the most prominent sites are the floodplains of the Ekuma and Oshigambo rivers and an alluvial fan at the south-west edge of the pan which are covered with this soil type. They generally show contents of moderate to high soluble salts (Beugler-Bell & Buch 1997).

In their reconnaissance soil and vegetation survey of the Park, Le Roux *et al.* (1988) summarized the soils around Etosha Pan into two main categories. Soils of the southern and south-western side, being covered with calcrete rubble on the surface, and the extensive aeolian Kalahari sands with calcareous loamy soils found in the west of the pan and thought to be related to the shrinking of the Etosha Pan. North of the Park, the carbonate-free Kalahari sands that cover this section were deemed as recent overburden.

3.6 Vegetation

The distribution of the vegetation in Etosha National Park shows a positive spatial correlation with those of soils (Le Roux *et al.* 1988). Additionally, the rainfall gradient of the area has a significant influence on the vegetation structure. For example, Le Roux (1980) observed that typically trees of over 6 m in height occur in the east where rainfall is high, whereas in the west where rainfall is lower, trees on similar soil barely reach 2 m in height. Plant species adaptation to rainfall is also recorded in the area, where some plants only occur in the wetter part of the park.

On the surface of the Etosha Pan, the higher content of sodium militates against the growth of vegetation (Figure 3.11). However, in above-average rainfall years, *Sporobolus salsus* and *Sueda species* may invade the margin of the pan (Le Roux 1980). The Gaseb alluvial fan, situated in the south-western corner of the pan, in particular, supports a healthy growth of the two mentioned species along with salt-tolerant grasses. Patches of grasses were also found locally in the middle of the pan.

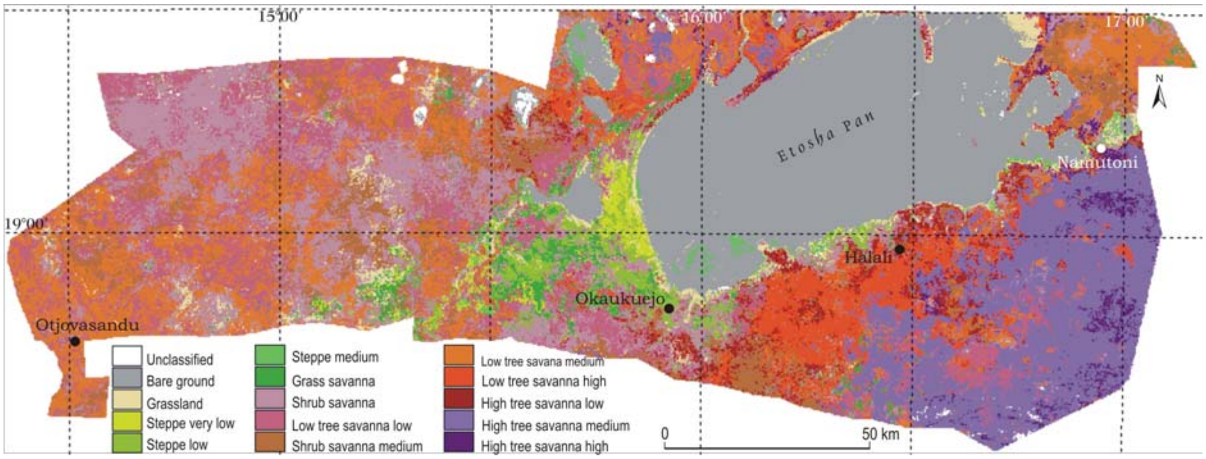

Figure 3.11: The vegetation structure of the Etosha National Park as derived from the classification of a Landsat TM image. Note the presence of vegetation on the surface of the pan, particularly on its south-western corner, being part of the Gaseb alluvial fan. (Source: Sannier et al.*submitted).*

Immediately around the Etosha Pan, Le Roux (1980) and Le Roux *et al.* (1988) mapped communities of tall grasslands along the western margin, the Ekuma River and Poacher's peninsula. The Andoni grass-plains were excluded from the above communities on account of their shallow soils, which are also saline and more or less alkaline. These properties affected the type of grass that grows there. 'Sweet grassveld on lime' is the mapping unit assigned to the extensive grassland bordering the southern edge of the pan. This unit is the most utilized by large numbers of herbivores. Locally this unit is punctuated with stands of woody plant communities, such as *Acacia newbrownii,* which occur in extensive thickets.

The deep Kalahari sands of the north-east of the Etosha Pan support a number of woody plants in which *Terminalia prunioides, Terminalia sericea, Croton menyhartii* and *Lonchocarpus nelsii* are some of the dominant species. Similarly loamy soils in the north-central and north-western of the pan support stands of *Colophospernum mopane*, interspersed with a variety of other plant species.

3.7 Land Use

The area zoned as Etosha National Park today has undergone through two major land use changes in the recent past. The period of common property was superseded by the game reserve era, with a protracted transition phase separating them. Before 1907 this region was utilized mainly by the Hei//om-, Oshiwambo- and Otjiherero-speaking communities as a hunting ground and seasonal (dry season) grazing land. Towards the end of the nineteenth century they were joined in hunting by the European traders and militias (Eirola 1992).

As a hunter-gatherer the Hei//om community, which occupied the core of the park before it was proclaimed, subsisted on the wealth of wildlife that the area once boasted. Their movements in the district were guided by the availability of water. They therefore utilized the area in response to the prevailing climatic conditions and food availability. The Oshiwambo-speaking communities on the other hand were mixed farmers, mainly tending crops and rearing livestock. A good deal of rainfall variability in the region imposed an adaptation to drought, which translated into the establishment of cattle-posts in the periphery when necessary. Etosha region was such an area and was therefore utilized for grazing in times of need. Although common access to land was a norm, the utilization of resources such as game was controlled. Oral accounts acknowledge the existence of a hunting season (generally during winter) which was centrally declared by the king of each ethnic group, in the case of the Oshiwambo-speaking community. The Otjiherero speakers are pastoralists. Like Aawambo, they thus exploited the area as prevailing climatic conditions commanded. Coupled with a relatively small population in the area at that time, the existing cultural practices with respect to the utilization of game helped maintain a thriving wildlife population.

The emergence of transboundary traders and colonial forces ushered in a new dynamic in the region. In particular, trading and the introduction of guns exerted immense pressure on the number, diversity and distributions of game. Due to concerns over the fast dwindling numbers of wildlife in the area, von Lindequest, the then German governor of the territory, declared Etosha as a *Wildschutzgebiet Nr. 2* in 1907 (Cloudsley-Thomson 1990). In so doing, Etosha became the third oldest game reserve in Africa, after Kruger National Park (1892) and Amboseli National Park (1899) (Adams 1996).

While in theory the land use of Etosha changed to the protection of indigenous plants and animals, trading and farming continued in the reserve until the mid-1950's (Cloudsley-Thomson 1990). An exception was the early 1960's when farmers were allowed to get their animals graze within the boundaries of Etosha, as a result of a severe drought that hit most of the country (Berry 1997). Towards the end of 1950's the desire to keep Etosha as a fully-fledged wildlife conservation area, in conjunction with a fledgling tourism industry, gained the necessary momentum. What followed was the drilling of numerous boreholes throughout the park, burrowing of pits for road constructions, and the erection of fences along the park boundaries (de la Bat 1982; Berry 1997). That conservation and tourism efforts took an earnest step forward during that time is perhaps evidenced from the construction of a tourist camp in 1967, named Halali. Borrowed from the German tradition, the word '*Halali'* is used as a bugle call for announcing the end of a hunt. Effectively, hunters turned poachers, a reputation that is often accompanied by a heavy fine, imprisonment or both.

The Etosha Pan itself is implicitly zoned as a wilderness, the highest category for the conservation of the natural environment. As such, human activities in the pan are strictly prohibited. An exception is extended to scientific research, which only after a stringent review may be allowed. It was under that umbrella that this study sought and obtained approval for accessing the pan interior.

Figure 3.12: 'Land use' of Etosha Pan at different environmental conditions. A) As the pan dries-up at the end of the rainy season, flamingos that previously flocked to the pan for breeding depart in phases. During the transition period the still remaining waterbirds make use of pools of water in the vicinity, such as the Ekuma River shown above. B) In contrast, when the pan dries-up, ostriches take refuge to the pan's interior, away from scavenging jackals, for nesting. (Photos taken in August 2004).

Most recently, the Etosha Pan was declared as a Ramsar site (Kolberg *et al.* 1997). This declaration was partly made due to the recognition of Etosha Pan as one of the only two (the other is the Makgadikgadi Pans) inland wetlands in southern Africa where lesser (*Phoenicopterus minor)* and greater (*Phoenicopterus rubber)* flamingos flock to breed (Sauer & Rocher 1966; Berry *et al.* 1973; Archibald & Nott 1987; Berry 2000). Interestingly, when the pan dries-up after the rainy season, ostriches move into the pan interior for nesting (Figure 3.12b).

CHAPTER 4

The Kunene River and the Catchment of Etosha Pan

Introduction

As mentioned earlier, the alleged geomorphic connection between the Kunene River and the Etosha Pan is prominently factored into the controversy surrounding the origin of the Etosha Pan. Hence, the subject of this chapter is the Kunene River, along with the Cuvelai system and the rest of prominent elements of the Etosha Basin catchment.

4.1 The Kunene River

Along the South Atlantic, between Benguela and Cape Town, there are only two solitary and exoreic, perennial rivers, the Orange and Kunene that drain into the ocean from the sub-continent. These rivers, some 1,400 km apart, outline respectively the southern and north-western boundaries of Namibia. The drainage area of the Kunene lies side by side with that of the Okavango on the crest of the Angolan highlands, and the watershed between them has poorly pronounced topographical features (Kanthack 1921). This watershed scenario repeats itself between the Kunene and the river systems draining north-west to the Atlantic. In contrast, the northern water divide follows the Lunda swell, which has an east-west orientation. This uplifted belt, reaching the Atlantic coast to the south of Lobito bay in Angola, is an extension of the mineral-rich Katanga belt of southern Zaire. North of this swell the drainage is directed towards the northern part of the Kalahari Basin into the Congo sub-basin (Kanthack 1921; Beetz 1933).

Rising near Huambo at an altitude of 1,840 m (Feio 1970), the Kunene has a catchment area of 106,500 km^2, and only 13% of that area lies within Namibia, the rest in neighboring Angola (van Zyl 1990; Burmeister & Partners 1993). Annual rainfall over the upper catchment averages 1300 mm. With a mean annual runoff of 5.2×10^9 m^3 (Chapman & Shannon 1985), 40% of this runoff is generated upstream of Jamba-la-Mina, an area equal to 13% of the total catchment (Burmeister & Partners 1996). The river has a total length of 1050 km (Simmons *et al.* 1993; Burmeister & Partners 1993; Burmeister & Partners 1996). Its fluvial environment, however, has undergone modifications since the 1950's due to a number of man-made structures directed on regulating its flow. Today, the river hosts the Gove Dam, constructed in 1973, Matala weir, in existence since 1954, Calueque weir, completed in 1977 and the Ruacana hydropower station, which dates from 1979 (Figure 4.1).

From its source, the Kunene falls fairly rapidly (Figure 4.1) for the first 300 km and flows generally due south. It then enters the Kalahari Basin at around latitude 15°45’ S (Wellington 1938). For the next 260 km it forms a flood plain, with its east side bank having a continuous scarp of soft sandstone averaging 21 m above the bed of the river. This scarp is broken only in three places in the vicinity of Calueque by iishana (Figure 4.2), thought by Wellington (1938) as palaeo-spillways, but now leading into the Kunene. They converge to form Oshana Olushandja / Etaka, which leads southwards towards the Etosha Pan, via Lake Oponono (discussed in section 4.2) (Kanthack 1921; Wellington 1938). In contrast, the western bank of the Kunene floodplain slopes very gradually.

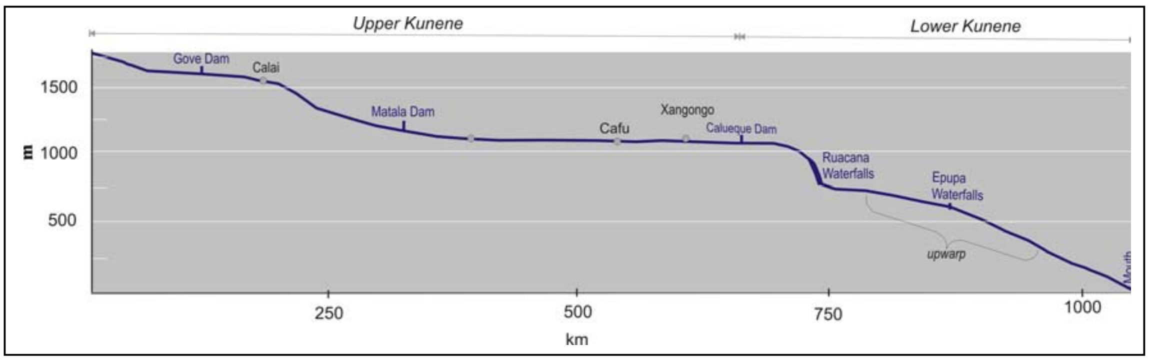

Figure 4.1: Longitudinal profile of the Kunene River. (Note: The Upper Kunene belongs to an ancient drainage system, whilst the Lower Cunene is geomorphologically younger and exposes a steeper gradient. Vertical exaggeration 1:160 (redrawn with modifications from van Zyl 1991; Burmeister & Partners 1996))

As it snakes through the Kalahari Basin of south-western Angola, the river pursues a sluggish course. It then takes a south-westerly direction for about 160 km, while the sluggish course is still in place and skirts the Kalahari Basin before reaching the western margin of the great Plateau at Erickson’s Drift (Kanthack 1921), in the vicinity of modern-day Calueque. Here, the floodplain track ends (Figure 4.2) and abrupt changes in the valley character begin (Wellington 1938). After about 10 km from this drift, the river makes a sharp bend westward, a direction it generally maintains for 340 km in its descent down through a 1000 m-escarpment to its mouth. The sharp bend of the river is made around a small cluster of kopjes, known as Okarundu, composed of red sandstones and grits. Kanthack (1921) regards the crest that lies here as the Kunene's natural critical point that regulated its bed-levels for many kilometers upstream, before the introduction of the man-made structures mentioned above.

From Calueque to Ruacana Falls, the river is characterized by no less than five rapids. Just a short distance downstream from the Calueque Dam is the crest of the Kavale / Kazambue Rapids. Here the Kunene, having just left the floodplain, narrows considerably while attaining a small gradient. The Kavale Rapids consist of red grits and conglomerates, along with greenish biotite schists at lower levels. These rapids are generally known as Small Cataract, and attain a total drop of around 15 meters. Below Kavale, the river cuts steadily deeper into the plateau, before it descends through a 21 m rapids, known as Chimbombe. Granite and a variety of igneous rocks make up the river bed at these rapids. A westerly course persists below these rapids and the river flows through yet another succession of pools and rapids for about 20 km. Suddenly, a transient southward course is pursued until the river plunges over the Ruacana Falls. For the 37 km distance of the river from the Kavale Rapids to Ruacana Falls, the river has a total drop of 55 m (Kanthack 1921; Burmeister & Partners 1996).

Ruacana Falls, 700 m wide, has a total drop of 124 m (Kanthack 1921; Petzel & Schneider 2004). It is located at the contact of gneisses of the Epupa Complex, one of the oldest rocks in the country, with the sediments of the Nossib Group. Due to its higher susceptibility to erosion, sediments of the Nossib Group are cut northward by the river, forming a meandering gorge of near-vertical walls. Thus, the formation of the falls is largely due to local influence and has little relation with what has been described as the "capture" of the Upper Cunene. The formed gorge, 1.6 km long and 120 m deep, follows the contact of the gneiss with the Nossib Group sediments (Wellington 1938; Petzel & Schneider 2004).

Immediately downstream of the falls, the Kunene exposes fluvio-glacial sediments of the Dwyka Formation of the Karoo Sequence. Further downstream from the Ruacana Falls, much of the river landscape is characterized by the bedrock geology of the Otavi System (Petzel & Schneider 2004). Some 120 km downstream from Ruacana, the river reaches an intrusive gabbro-anorthosite complex in which the Epupa Falls are hosted (Burmeister & Partners 1993). The catchment area between the Ruacana and Epupa Falls shows a lower gradient of the river and its large tributary valleys with wide-open pediment slopes. Downstream from the Epupa Falls, the river morphology changes its character and a deep gorge with several cataracts is cut. From there on, short seasonal streams, including the Marienfluss, occasionally enter the river on either side of the bank. The Marienfluss, in particular, is a dry sandy riverbed located on the Namibian side of the Kunene draining a Tertiary basin dominated by terrestrial sandstone (Burmeister & Partners 1993).

As the Kunene River mouth approaches its mouth, a 70 km wide coastal desert, the Namib, reaches the left bank of the river. In this area, sand dunes are up to 100 m high. A plethora of aeolian, fluvial, deltaic, and marine processes define much of the lowermost section of the Kunene. Part of this section receives wind-blown sediments, which the river transports to the sea. In turn, upon deposition the northward-bound littoral drift transports these materials in that direction, ultimately leading to coastal dunes formation to the north of the river mouth. Time-series studies of the river mouth revealed its continuous northward migration due to littoral drift (Feio 1970; Burmeister & Partners 1993; Simmons *et al.* 1993). There are also signs of a diminishing mouth due to changing flow patterns (Simmons *et al.* 1993).

For a river of its size, the Kunene river mouth is relatively small. It varies between 30 m to approximately 80 m across during low and high flow, respectively (Simmons *et al.* 1993). Similarly, the marine shelf at the mouth is reported to be narrow (45 km), with a shelf break of approximately 200 m depth (Shannon 1985). More than 75% of the river mouth sediments are terrigenous, mainly consisting of sand particles. Silt is deposited to the north of the Kunene, resulting from a redistribution of the river sediments by the northward littoral drift. The low amount of organic matter (up to 0.9%) contained in these sediments reveals the past and present (high) water quality of the river (Burmeister & Partners 1993).

The nature of sediment load of the river can be differentiated based on its major sections. Suspended fine-grained sediments and dissolved load prevail in the upper Kunene, whereas the lower Kunene is dominated by bedload. Much of the sand that constitutes the bedload is presumed to have been injected into the river system from dune fields of the Namib Desert through which the river passes before reaching the coast (Burmeister & Partners 1993).

At Ruacana the annual suspended sediment and dissolved load is estimated to be 430,000 and 200,000 tonnes, respectively (Burmeister & Partners 1993). Senin (1969), however, had earlier estimated the annual sediment delivery to the sea to be 5.8×10^6 m^3 (386 000 tonnes). Clay of the suspended load mainly consists of smectite (59%), kaolinite (32%) and some illite (9%). Numerous amphibolite bodies, the Kunene Igneous Complex and the soils of the drainage area are believed to contribute to the high smectite content of the riverine clays (Bremner & Willis 1993). Gingele (1996) suspected that kaolinite may have formed from bedrock of weakly acidic terrains. The illite assemblages are similar to those of aeolian dust raised from the Namibian interior and the rest of the Kalahari (e.g. Bremner & Willis 1993). There is also a high concentration (max 92%) of glauconitic facies which occur off the Kunene River (Bremner & Willis 1993).

Unlike the mature profile of the Kunene River upstream of the Calueque dam, the section downstream from that point has a juvenile character. Among others, the upper Kunene has mature, low gradients (averaging 1:1,100), flood plains inundated on an annual basis, and natural deposition environments. In contrast, the lower Kunene has steep gradients (averaging 1:400), rapids and a channel controlled by the bed-rock structure.

These distinct geomorphological expressions are believed to have been rooted in the regional geological development of the subcontinent. The upstream section is traced to the proto-Upper Cunene River, believed to have been an internal drainage into the Owambo Basin (Wagner 1916; Hedberg 1979; Miller 1997). Part of the proto-Lower Cunene on the other hand, was initially thought to follow an exhumed glacial valley formed during the Late Carboniferous. The old valley is taken to have responded sensitively to deep erosion associated with the break-up of Gondwana and was subsequently re-exploited as a short, exorheic river (e.g. Thomas & Shaw 1988; Allanson *et al.* 1990; Mendelsohn *et al.* 2000; Roesener & Schneider 2004). Thus, the river morphology of the lower Kunene is lumped into the same category as the deeply cut gorges of ephemeral rivers, such as the Hoanib and Huab, located south of the Kunene River (Maack 1969; Martin 1981). However, recent studies disputed that assumption of a Dwyka valley. In particular, Sander (2004) concurred with Feio (1970) and concluded that the river valley was influenced by a graben which prevented Dwyka sediments from being eroded.

Nevertheless, it is widely assumed that the proto-Upper Cunene and the proto-Lower Cunene fused in the Pliocene / Early Pleistocene under the influence of neo-tectonics, when the former was deflected into the latter and unified as the contemporary Kunene River (e.g. Wellington 1938; Stuart-Williams 1992a). In comparison with the Kavango, the Kunene is a much steeper and less sluggish river (Kanthack 1921). In addition, the Kunene has a very small discharge in the winter and spring months and a very heavy flood discharge during the rainy season.

4.2 The Kunene River and Etosha Pan

In light of the lingering view that draws a geomorphic connection between the Upper Kunene River and Etosha Pan, via Oshana Etaka/Olushandja, it is deemed as appropriate that this topic be dealt with here. Technically, the two names, Olushandja and Etaka, refer to one and the same oshana, flowing in opposite directions. Oshana Olushandja, flowing northwards, joins the Kunene some 3 km upstream of the Kavale / Kazambue Rapids, whereas Etaka slopes southwards towards the Omadhiya Lakes (Figure 4.2). The average gradient for the oshana Olushandja and Etaka is 1:2,700 and 1:7,700, respectively, which points to the youthful character of the former. A number of investigators (e.g. Schwarz 1920; Wellington 1938) regarded this channel as a former "spillway" from the Kunene to the Etosha Pan.

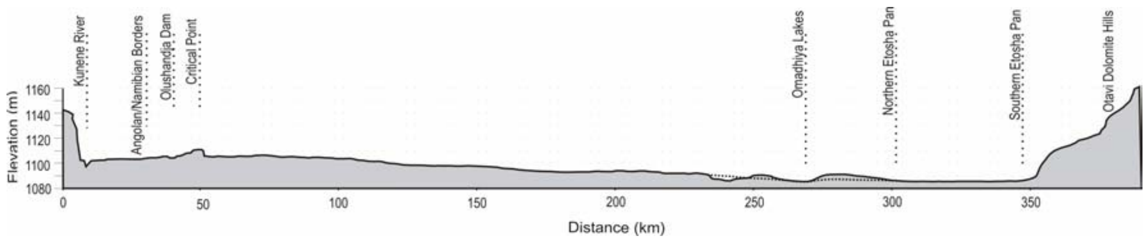

Figure 4.2: Longitudinal profile of Oshana Etaka/Olushandja from the Kunene River to Etosha Pan. (Note: Double ridge around Omadhiya Lakes resulted from striping effect error in the DEM data, discussed in section 3.1; the overlapping dotted line estimates the ideal profile at that location). Vertical exaggeration is 1:930.

The spatial setting of the oshana Olushandja in relation to the Kunene River is presented in Figure 4.3, while the generalized geomorphological map of the area is portrayed in Figure 4.4. Both figures illustrate that oshana Olushandja enters the Kunene River valley at almost a 90° angle. As observed by Kanthack (1921), this entry occurred at the most westward bend of the Kunene River, where its broad floodplain terminates, and its stream enters a narrow, bed-rock controlled valley. The SRTM data show that the oshana Olushandja joins the Kunene valley at approximately 1102 m a.s.l., being roughly the same slope gradient the river maintains for a distance in excess of 10 km upstream from the confluence (Figure 4.5). Around that point, the immediate upstream valley narrows from a width of approximately 6 km to 100 m over a distance of about 15 km. At the confluence of the oshana Olushandja and the Kunene, the valley has a width just over 2 km in extent.

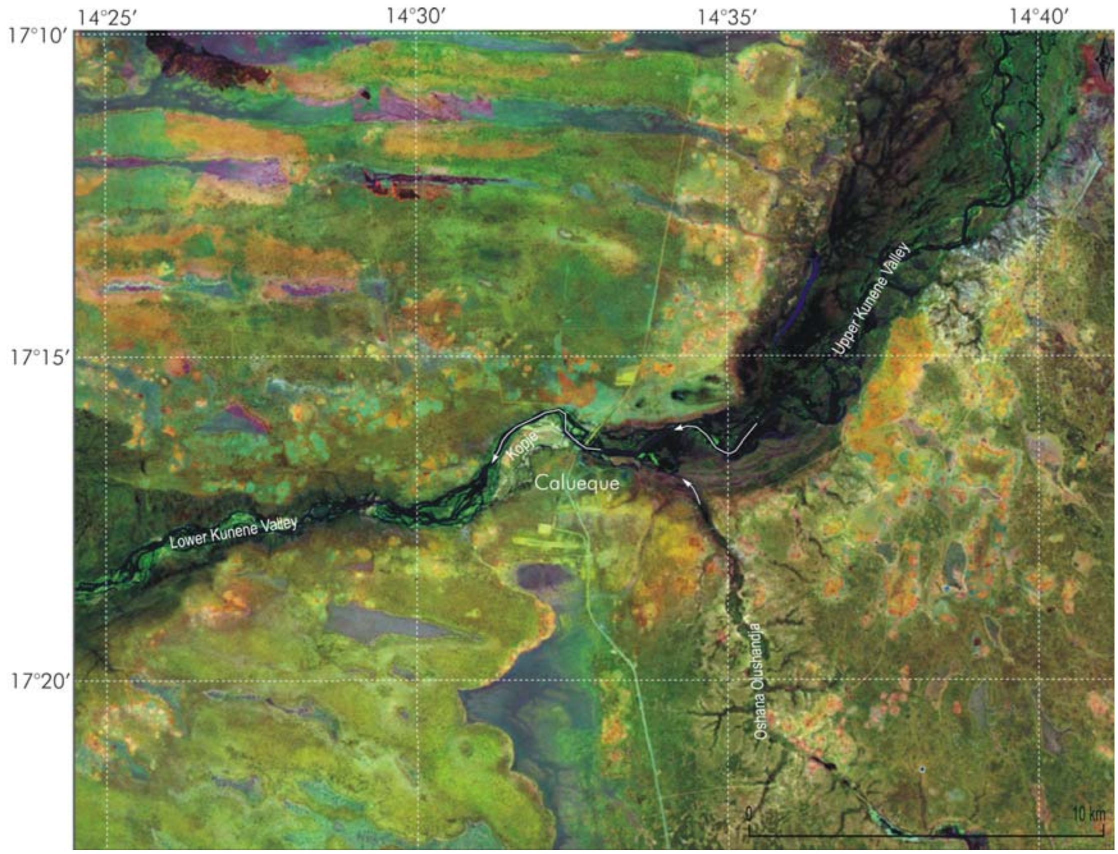

Figure 4.3: Spatial setting of oshana Olushandja and Kunene River from a Landsat TM perspective, betraying through the land cover that the Kunene once may have continued flowing southward at almost the same width as the floodplain upstream of the narrow valley. Note: Almost all the tributary channels to the Oshana Olushandja join it at a right angle, and the Olushandja itself enters the Kunene valley at almost the same angle. The lithology along the oshana Olushandja is Kalahari sands. The Oshana Olushandja joins the Kunene valley, just before the river changes from a floodplain to a narrow valley, characterized by numerous rapids (Bands 7-4-2 in RGB; image captured in November 1986).

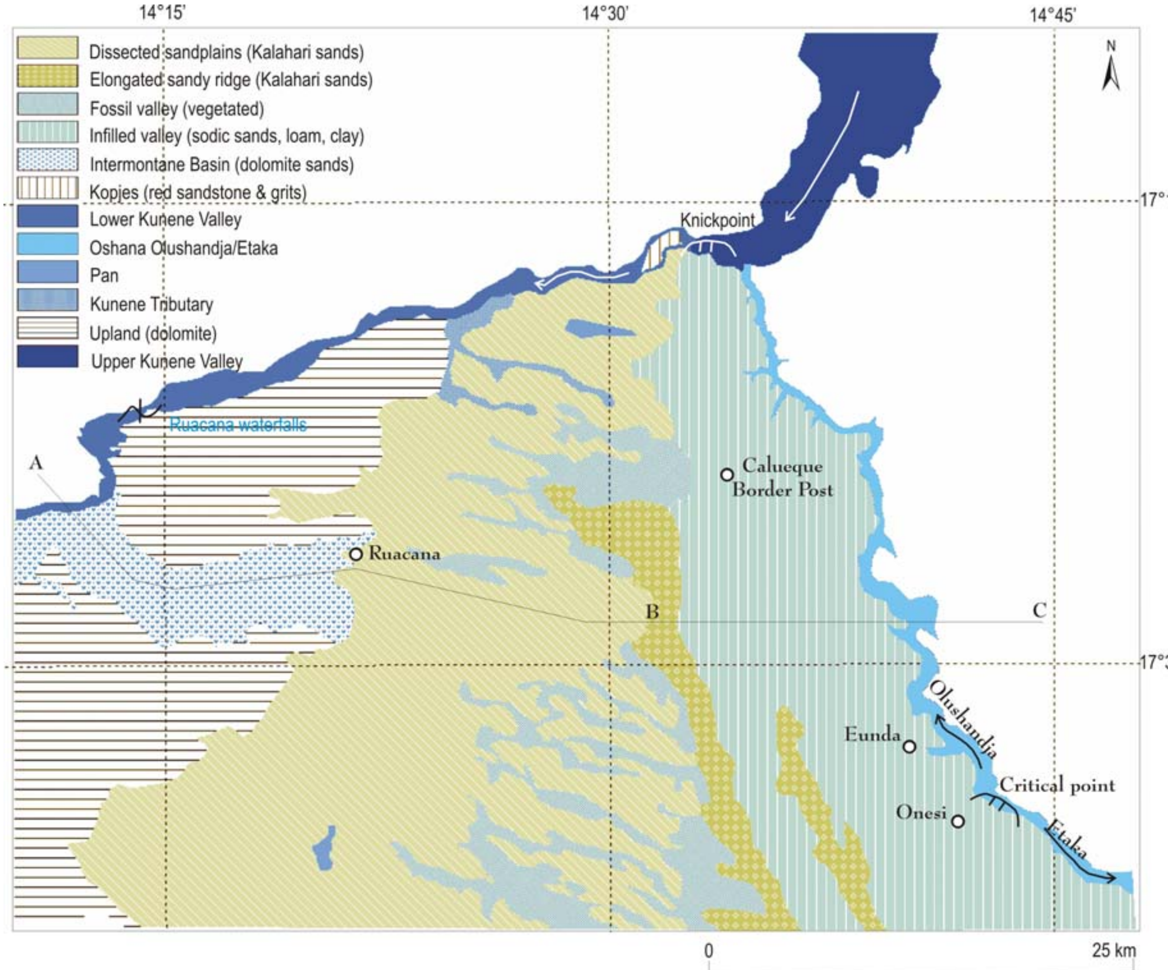

Figure 4.4: Generalized geomorphological map of Ruacana area.

Approximately 3 km downstream from the elbow, the Kunene turns northwest to pass around the Okarundu kopje (Figure 4.3, 4.4). Although the main tributary passes this kopje to the north, there are also traces of spillway channels along its southern edge. This elongated kopje measures some 500 m across with a length of about 2.5 km, and Kanthack (1921) regarded it as the river's natural critical point regulating its bed levels upstream; that was before the man-made structures mentioned above were constructed. Downstream from the Okarundu kopje, the valley slope drops to less than 1090 m a.s.l. as it enters a deeply cut zigzag gorge, with almost vertical walls (e.g. Roesener & Schneider 2004). Given the river morphology and geomorphological setting of this confluence, it is fair to suspect here that this kopje is a remnant of the "wall" that was breached by the upper Cunene River to drain coastwards.

In supporting the above pre-emptive suspicion of the diversion of the upper Kunene River, further evidence was found along the Oshana Olushandja / Etaka. Wellington (1938) had already concluded that the oshana Olushandja has an up-slope gradient from the Kunene southwards to Onesi, before a slope reversal of the oshana Etaka defines the rest of the channel as it drains southwards (Figure 4.2). He also derived that the slope of oshana Olushandja has a rise of 13.5 m over a ground distance of 45 km along the channel. Presently, the last 19 km of that distance is occupied by the Olushandja reservoir, having a maximum water depth of 4 m and a surface area of 26.6 km² (Department of Water Affairs 1975).

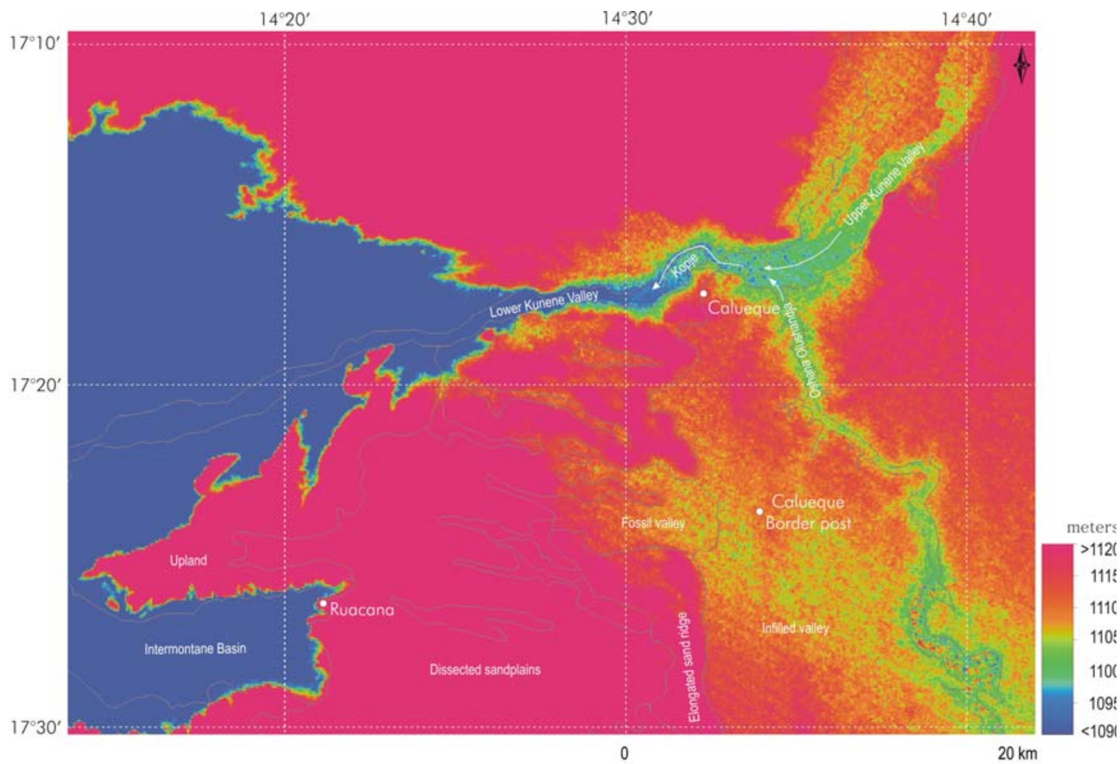

Figure 4.5: Selectively stretched (1090 -1120 m) DEM of Caluegue area where oshana Olushandja joins the Kunene River with polygon overlay of the generalized geomorphological map. Much of the area below 1095 masl, is part of the Kunene River gorges and the intermontane basin.(Data Source: NASA)

Prior to the construction of the said reservoir, precise levelling of the area resulted in topographic data of one-meter resolution. That data set was used here to construct a longitudinal profile (Figure 4.7) along the bed of these iishana. Together with satellite imagery, analyses of the topographic data revealed that the water-divide between these two iishana is located on an uplifted block, interpreted here as a lowland horst. On the Olushandja side, the lowest part immediately around the horst is some 3.6 m below the knick point. Likewise, the lowest part of oshana Etaka, immediately adjacent to the horst, is 2 m below the critical point. Part of the horst, including the critical point, hosts the reservoir, which, quite understandably, has embankments on both flow directions of the channel. The horst measures 21 km along the bed profile. From remote sensing data analysis, this horst was found to be oriented in a northwest – southeast direction, lying diagonally to the channel.

In Wellington's (1938) view, the gradient reversal of Oshana Olushandja substantiated the hypothesis put forward by Schwarz (1920) that the Kunene at one time flowed towards Etosha Pan along the Oshana Etaka. Both Schwarz (1920) and Wellington (1938) erroneously assumed that the oshana Olushandja cut its bed southwards and in the process "moved" the critical point towards the Etosha Pan. In endorsing that assumption, Burmeister & Partners (1993) predicted that in the absence of human interference in the area, such as the regulating of the river, the natural processes would have led to the shifting of the critical point to the Etosha Pan, and divert the Etosha drainage into the Kunene River. The data presented here, however, attribute this up-slope gradient of the Oshana Olushandja from the Kunene River to the role of neo-tectonics, with its structural surface expression still detectable. In that light, the critical point has remained seated over the horst, and has not shifted for tens of kilometers as purported earlier.

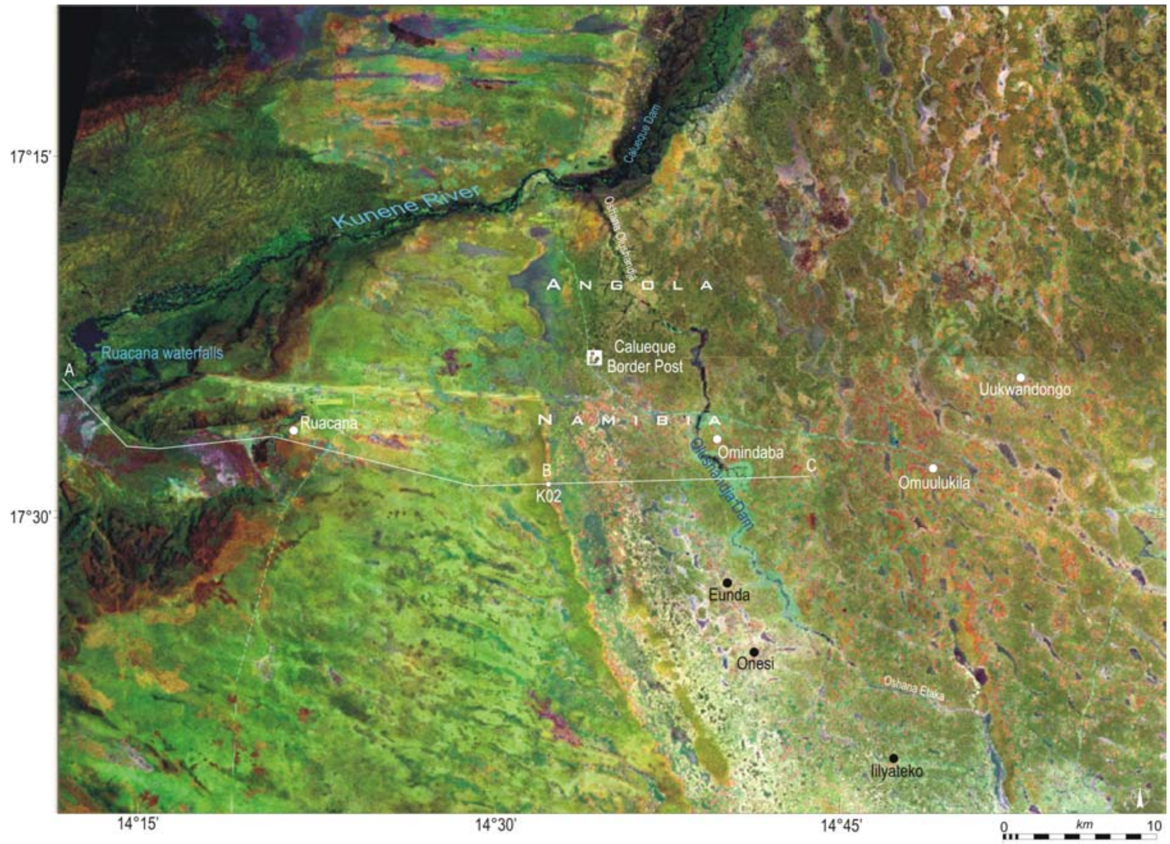

Figure 4.6: Landsat Imagery (November 9, 1986; 742:RGB) covering Olushandja Reservoir and Kunene River environs.

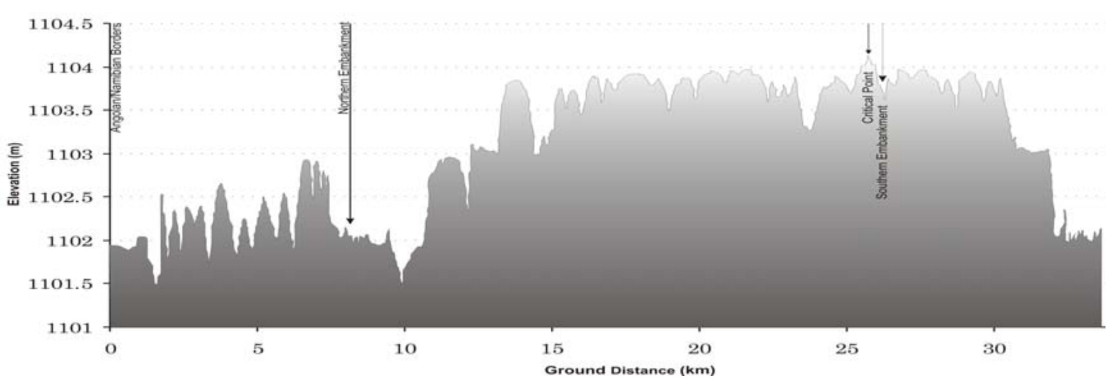

Figure 4.7: Longitudinal profile of Oshana Olushandja/Etaka showing the critical point. (Countours digitised from Hydro-consults, 1970[i], at a scale of 1:25,000 and contour interval of 1 m; vertical exaggeration 1:1,665).

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[i] *Values of the elevation data presented here differ from those generated from SRTM; despite this discrepancy, the general outlines of both profiles where they overlap are comparable. Since the Hydro-consults data were specifically generated for engineering purposes in relation to the construction of the Olushandja Dam, they are deemed to have a higher degree of precision. Its accuracy with respect to sea level is also in a general agreement with the topographic maps of the area at a scale of 1:50 000, both being referenced to the Bessel 1841 (Namibia) ellipsoid, and not to the WGS84 system like the SRTM data.*

Another thread of evidence for supporting the drainage of the upper Kunene River flowing southwards is gleaned from the remote sensing data which was supplemented with geomorphological surveys. From these analyses, it emerged that the Oshana Olushandja / Etaka is located on the eastern edge of a wide infilled valley (Figure 4.4, 4.8). This valley narrows down towards the entry point of the oshana Olushandja into the Kunene River and broadens as it fades away in the opposite direction. As it can be observed in Figure 4.6 the northern base of this valley is projected into and fits the size of the current valley of the upper Kunene. Characteristically, the broadening of the infilled valley signals a change from relatively confined beds.

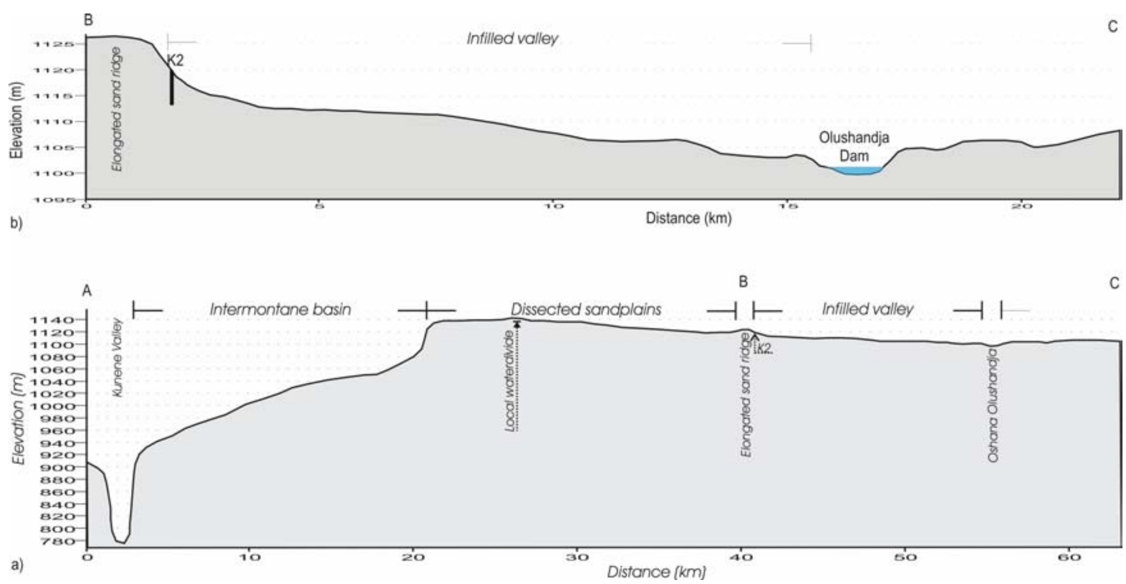


Figure 4.8: a) Cross-section through major geomorphological units around Ruacana area; b) enlarged cross-section covering the infilled valley and Olushandja Dam (see Figure 4.6 for the location of the profile; vertical exaggeration (a) – 1:42 and (b) – 1:180).

Some 15 km southwards from the borders of the infilled and Kunene valleys, there is an elongated sand ridge, which curves along the infilled valley. In addition to its spatial setting, its orientation suggests that it is a former natural levee of the infilled valley. The Etunda Irrigation Scheme, where over 300 hectares of various crops are grown, is located on this sand ridge.

A glimpse of sub-surface sediments of this area was revealed in a gravel pit (K2) located at the line of contact between the elongated sand ridge and the infilled valley. Figure 4.9 shows a section of the southern flank of this gravel pit, which has a depth of about 6 m. Four distinct layers were recognized. The upper layer is composed of colluvial material, containing dark sodic soils. They have been mapped by Mendelsohn *et al.* (2002) as *eutric Cambisols*. With a total thickness of about 1.5 m, this layer unconformably rests on a 3 m stratum of calcrete, with nodules in places.

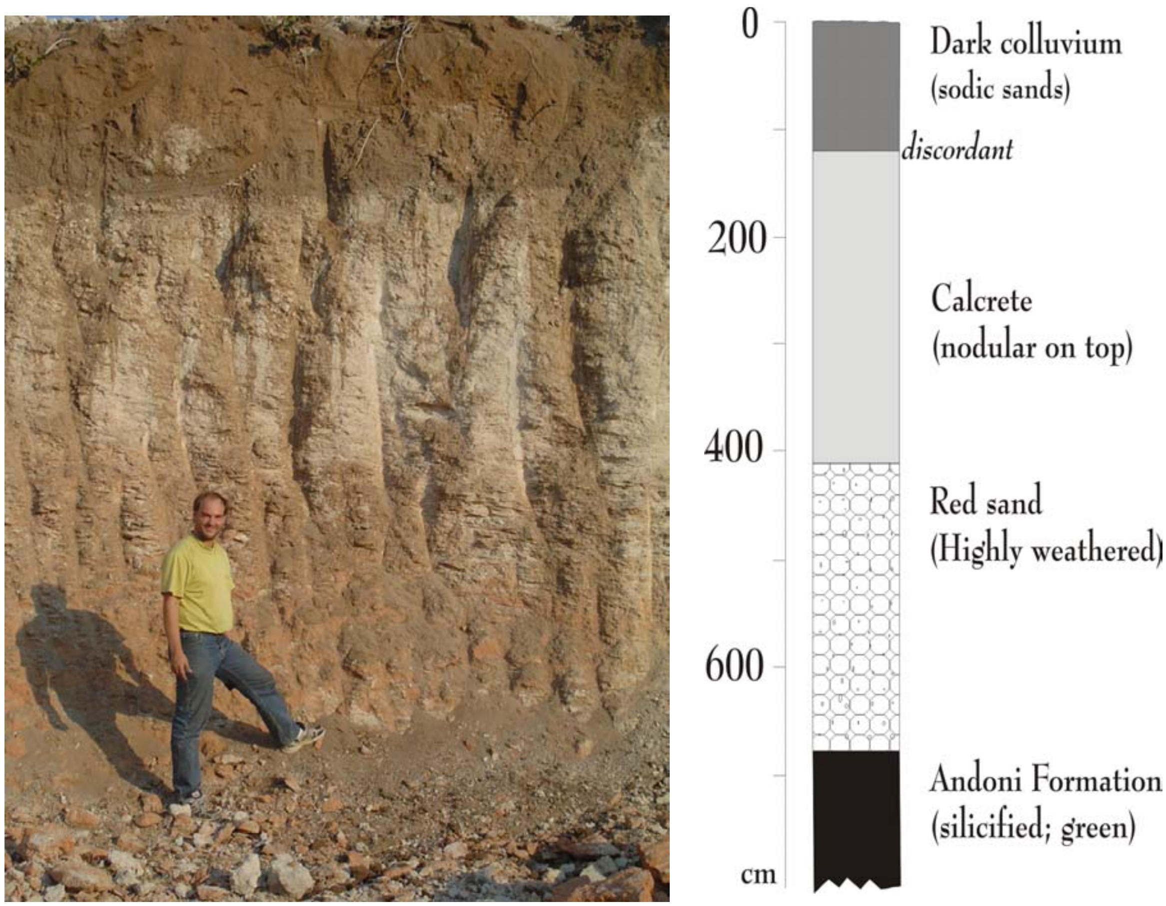


Figure 4.9: Photo showing the southern flank of the gravel pit, and corresponding profile at K2.

On the northern flank of this pit, a section of the calcrete layer displays angular to semi-angular pebbles, laid sub-horizontally and dipping to the east. The calcrete layer grades gradually into a highly weathered red sand deposit with a thickness of up to 3.5 m. The bottom of the pit is covered with olive-green sediments of the Andoni Formation. The uppermost section of these sediments is silicified, whilst the rest of the exposed layer is finer and more weathered. As mentioned in section 3.2.3, the Andoni Formation is regarded as the uppermost layer of the endorheic Etosha basin.

The existence of an infilled valley was also confirmed on the basis of groundwater distributions. Boreholes data from the area revealed that unlike the nearby dissected sandplain, the infilled valley is characterized by a shallow aquifer. A number of wells tapping this near-surface groundwater have depths in the range of 10 m. In contrast, boreholes located on the sandplain have depths in excess of 100 m (Christelis & Struckmeier 2001). Like the abandoned valleys (Figure 4.4) the flow of the groundwater in the sandplain is directed eastwards. Although the presence of more than one different aquifers at different depths is known to exist in the Cuvelai (e.g. Mendelsohn *et al.* 2000; Niemann 2000), the absence of the shallow aquifer on adjacent and nearby terrain units implies the existence of an infilled valley.

Due to its size, location, prominence in the remote sensing data, and the additional attributes detailed above, this infilled valley is therefore taken as the former valley of the Kunene River, with the oshana Olushandja / Etaka now occupying its eastern-most section. Sediments are thought to have brought into the valley from the elevated part, notably the western section. However, the most vexing question that, unfortunately, remained unanswered is related to the time when the Kunene River abandoned its previous south-oriented valley, before sedimentation took place.

4.3 The Cuvelai System

The core and major surface of the Owambo Basin is drained by the Cuvelai System. This fluvial system lies wedged between the catchments of the Okavango in the east and Kunene rivers in the west (Figure 4.10). In Angola, the region drained by the Cuvelai system has a surface area of approximately 50,000 km^2 (Mendelsohn *et al.* 2000). This system is virtually a maze of shallow water courses, called iishana (*sing.* oshana) funneling towards the Etosha Pan. Perennial tributaries only occur in Angola (Bittner & Plöthner 2001), where some of their headwaters originate in the region of Serra Encoco, a southerly spur of the central Angolan highlands located east of Cassinga (Stengel 1963). This area receives an annual average rainfall in excess of 800 mm (Mendelsohn *et al.* 2000).

The Cuvelai is the principal river of this system. It has two major tributaries, the Mui-Mui in the west and the Caundo in the east. The Mui-Mui enters the Cuvelai below Evale, after running parallel to the Cuvelai for a considerable distance. The Caundo empties into the Cuvelai north of Namakunde (Stengel 1963). This tributary has three major streams, of which Oshigambo is one.

Iishana are typically of very low gradient (averaging 1:5,300; Figure 4.11), covered with bush and grass, through which the water flows. Often, these flows have not developed a well-defined flood channel. Their banks are usually lined with trees and bushes. As water progresses further south, the topography becomes gentle. Along with the basin effect, this topographic change causes iishana to meander along with frequent silting and re-routing being noted (Groundwater Consulting Services 1990; Mendelsohn *et al.* 2000). Higher grounds, locally known as *omifitu*, 1.5 m to 6 m in height (Wellington 1938), are located between the iishana (Figure 4.12). Omifitu are mainly mantled with Kalahari sands and support a variety of trees and shrubs (e.g. Marsh & Seely 1992). They are often considered as levees. Collectively, the Cuvelai landscape is an inland delta.

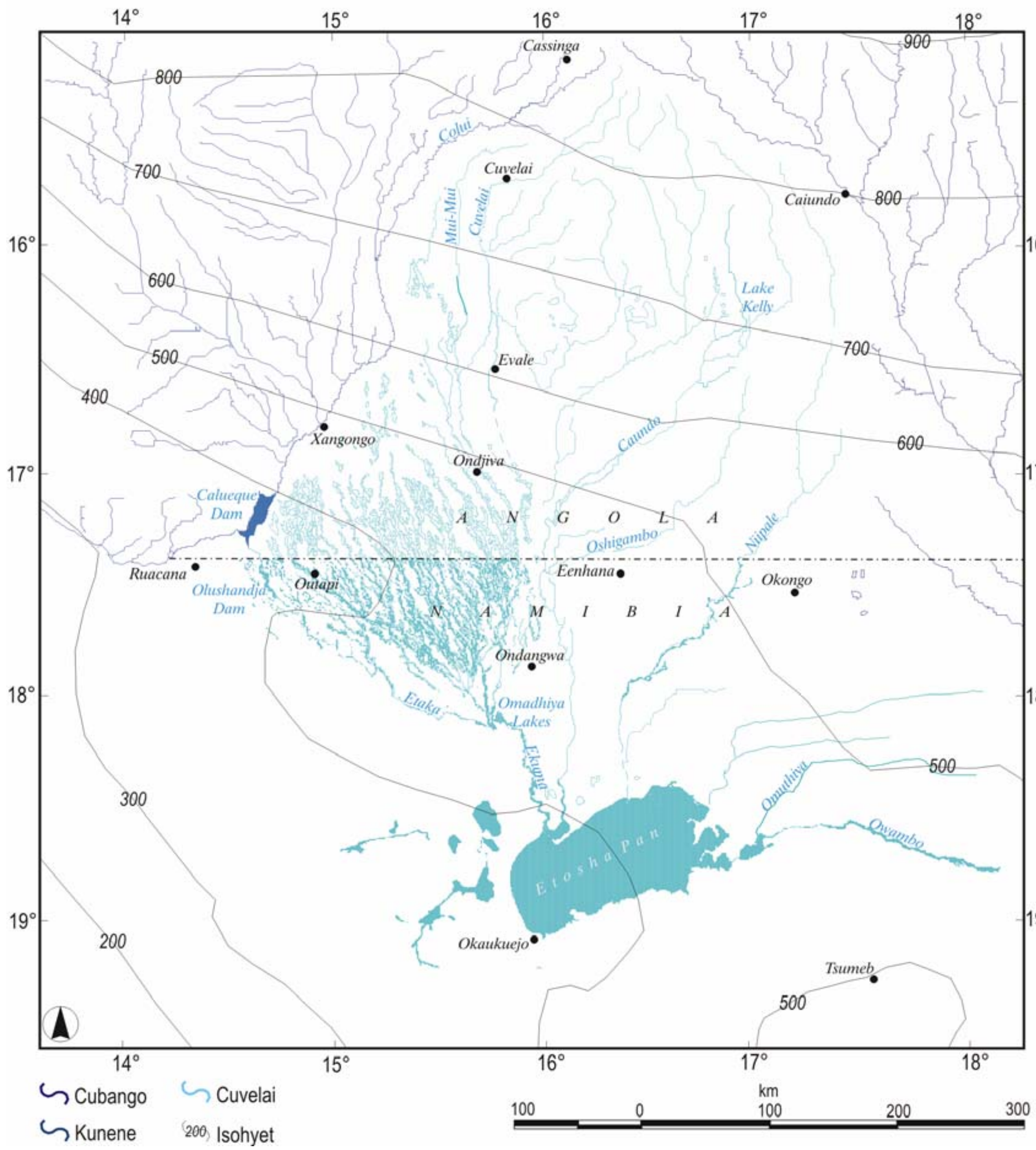


Figure 4.10: The Cuvelai System (Redrawn after: Stengel 1963; van der Waal, 1991; Mendelsohn et al. *2000)*

In the Namibian part of the Cuvelai, iishana flow only during the rainy season. This is particularly so when rains fall over the entire catchment, which may cause a large flood locally known as efundja. Usually, iishana start filling up with rainwater early in the season. By February, floodwater from Angola reaches northern Namibia. Flows last only for a few months, ending around May (van der Waal 1991). From a series of records compiled over a period of 58 years, Mendelsohn *et al.* (2000) concluded that on average, a major efundja would be expected in three out of 20 years, while no flow will be experienced in seven years of that cycle. The other half of the 20 year period would witness either a moderate or minor flow, in equal proportion.

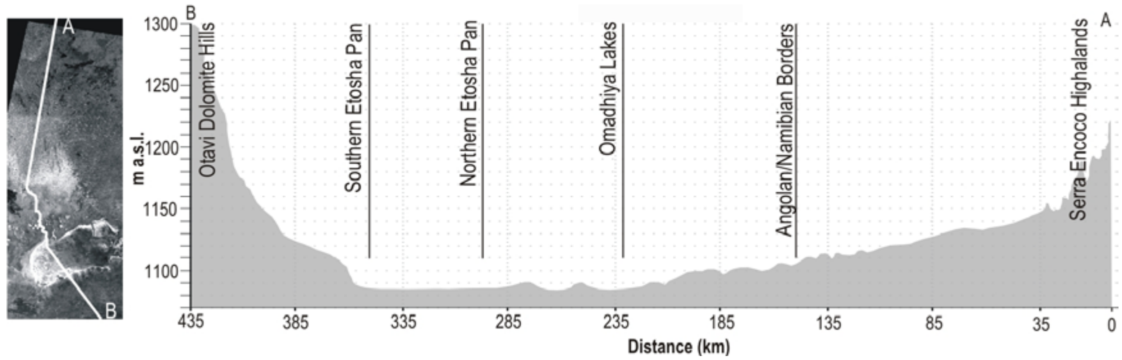


Figure 4.11: Longitudinal profile across the Cuvelai System from the southern section of the Encoco Highlands to the northern part of the Otavi Dolomite Hills Note: Ridges between Omadhiya Lakes and the Etosha Pan are artifacts propagated from the original data (see section 2.1). Vertical exaggeration is 1:580 (Data source: NASA).

In surveying fishes of the Cuvelai in April 1976, van der Waal (1991) reported a flow velocity of 0.25 m/s from a well vegetated oshana. Based on the category made by van der Waal (1991) and Mendelsohn *et al.* (2000), that year fell in the group of a major efundja, with Oshakati measuring an annual rainfall of 716 mm. Stengel (1963) reported that its velocity in these channels rarely exceed 0.5 m/s, a condition endorsed by Marsh & Seely (1992) who attributed the sluggish velocity to the roads built across iishana which partially altered the flow of water.

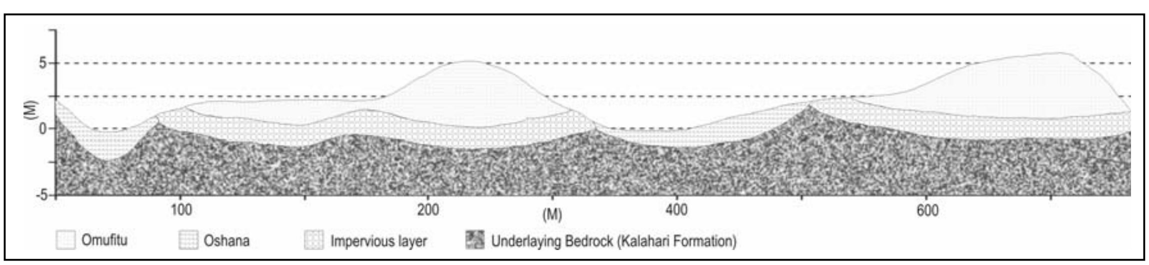


Figure 4.12: Typical cross-section of Oshana and Omufitu in Angola. (Redrawn after Stengel, 1963, which in turn was drawn based on the Portuguese publication "Abastecimento de aqua ao baixo Cunene relatório dos trabalhos efectuados em 1954".)

Typically, iishana vary between one and seven meters in depth and 100 to 500 m in width (Groundwater Consulting Services 1990, 1991; van der Waal 1991). However, narrower and densely spaced iishana are common in the east, whilst the western area has much broader and flatter channels spaced farther apart. Additionally, the western iishana begin their courses near the sandstone scarp of the eastern bank of the Kunene River (section 4.1), except for the three iishana near Calueque, of which Oshana Etaka/Olushandja is one. These three exceptional iishana are connected to the Kunene River, to where the connecting channels drain (i.e. section 3.2.1). Etaka oshana in particular, is also unique in that it is deeper than the rest of the western iishana, but still flatter than those in the east. Wellington (1938) suggested that either iishana were spillways during an earlier stage in the evolution of the main stream, or they are simply rain-water channels that gradually cut back towards the sandstone ridge.

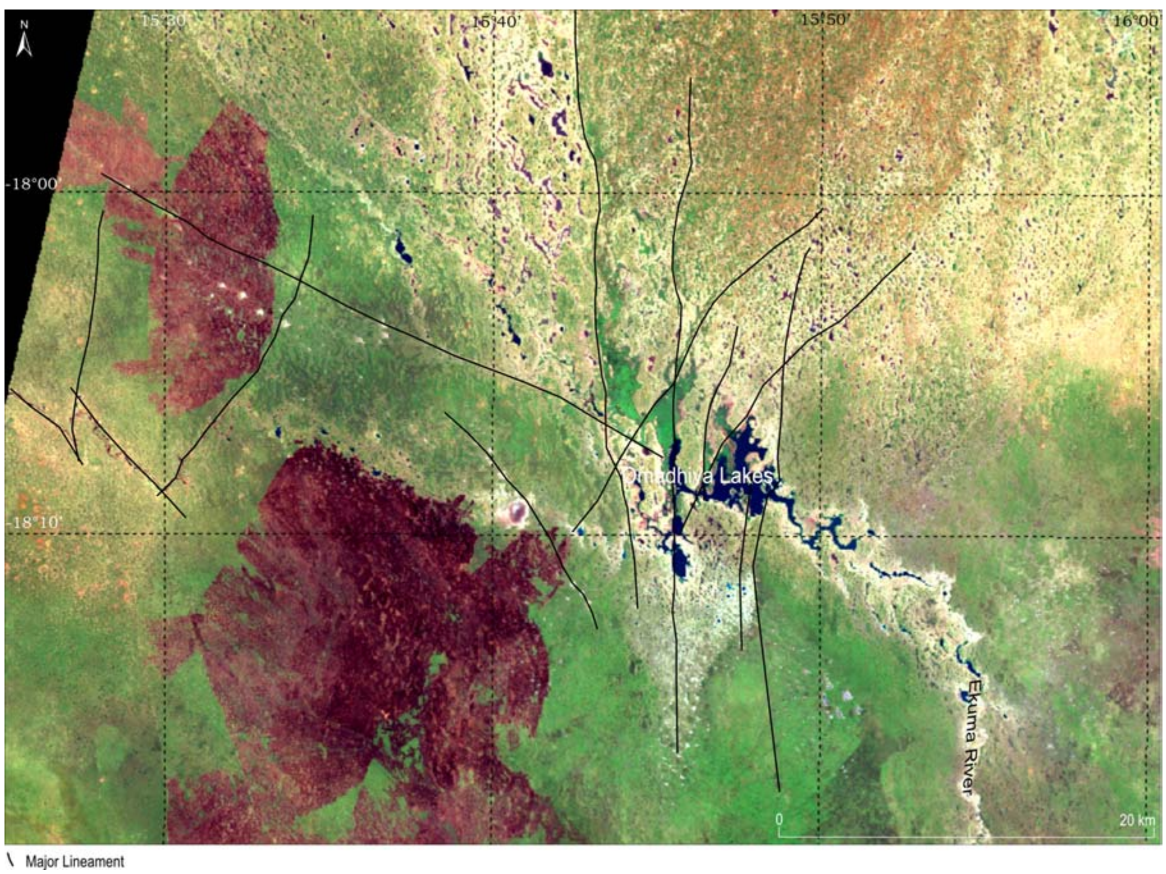


Figure 4.13: Spatial setting of the Omadhiya Lakes and major lineaments in the area. The lakes are relatively at low levels, the blue shades designate inundated area (Landsat image Bands 742 in RGB, acquired 08/08/1984).

Most of the iishana, coming in from the western to northeastern sections of the Cuvelai, converge at the Omadhiya Lakes (Figure 4.13), located some 60 km northwest of the Etosha Pan. Lake Oponono is one of the best-known members of these interconnected, semi-permanent water bodies. It is from these clusters of lakes and pans that water issues in a single channel, the Ekuma River, before it debouches into the Etosha Pan. Lindeque & Archibald (1991) reported, however, that although water may flood into the Ekuma River, only in exceptional flood years it will reach the Etosha Pan. This condition is conceivable, as insight from topographical data revealed that the Omadhiya Lakes are some 2 m below the beds of the northwestern section of the Etosha Pan, at the mouth of the Ekuma River. Moreover, satellite image interpretations show that the Omadhiya Lakes are aligned along major lineaments and the surrounding area is also criss-crossed by similar structural features (Figure 4.13). Hence, the topographic difference between the Omadhiya Lakes and the Etosha Pan, which in turn is attributed to neo-tectonics, is the likely factor that affects this supply of water into the Ekuma River but coming short in reaching the Etosha Pan. The role of neo-tectonics could also be the significant player for the strikingly funneling of (semi) parallel iishana, that were up to 140 km apart, to converge at the Omadhiya Lakes (Figure 4.10).

As the Ekuma River enters the Etosha Pan, it forms a bird-foot delta. In Wright's (1985) classification of these landforms, the Ekuma Delta falls in the category of a fluvial or river dominated delta. It measures approximately 11 km in length and is 4.5 km wide at its broadest base. Its natural levees are tens to hundreds of centimeters above the channels. Channels are bare, whilst natural levees are covered with grass. Two major distributary mouth bars caused the channel to bifurcate first at 3 km and then at 500 m from the delta apex.

Unlike the Ekuma River that has its root in the Omadhiya Lakes, the channel of the nearby Oshigambo River has no connection to these lakes where most of the iishana converge. Its channel generally runs east of the area where the density of iishana is highest. Some 40 km from its entry point into the Etosha Pan, however, the main channel of the Oshigambo River becomes indistinct and fades away, for a distance of 13 km towards the pan (Figure 4.14). Thereafter, four incipient streams have joined to form the downstream channel of the Oshigambo River that reaches the Etosha Pan. It is likely that either the indistinct area of the Oshigambo riverbed is sediment-choked or it is affected by neo-tectonics or both.

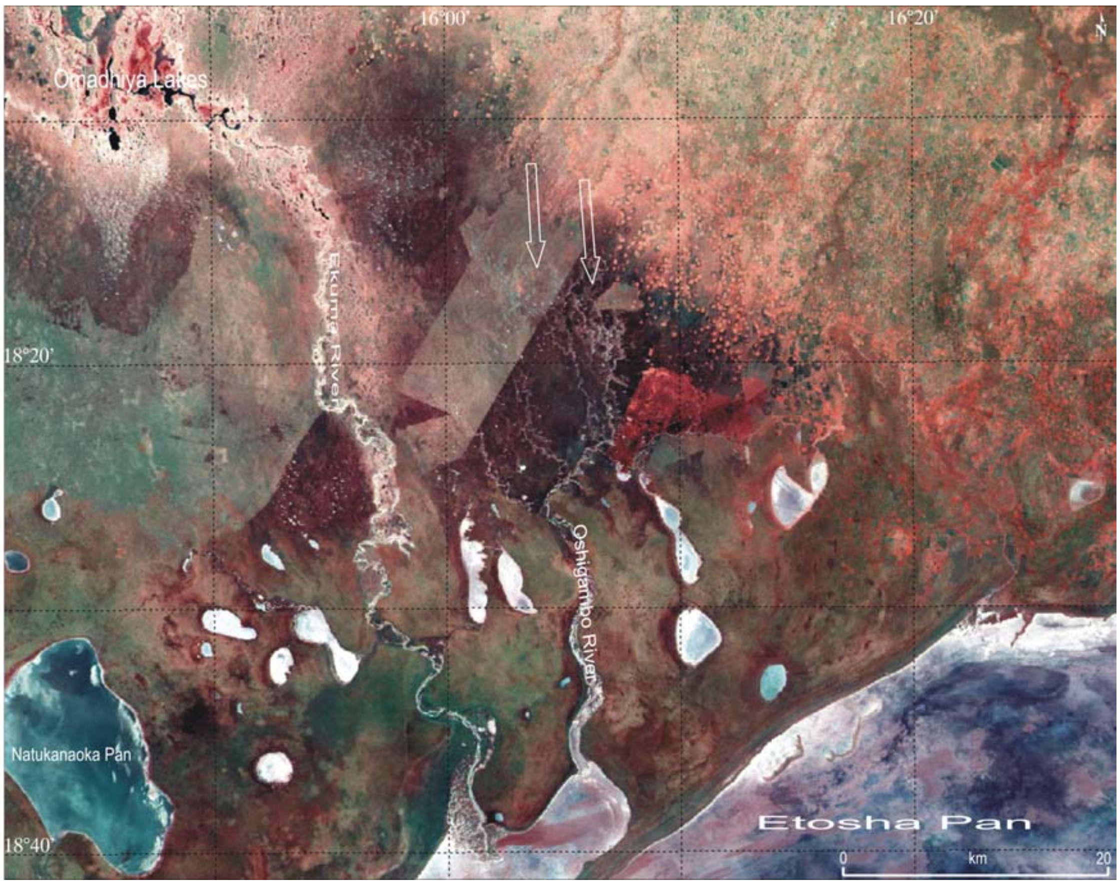


Figure 4.14: Landsat TM+ image (August 2002) of the area between the Omadhiya Lakes and the northern section of the Etosha Pan. Two down-pointing arrows show the area where the upstream channel of the Oshigambo River is indistinct.

Despite its size and prominence in the area, the Oshigambo River has not formed a delta like the neighboring Ekuma. Instead, its adjacent segment to the Etosha Pan has a bar-braided channel with significant in-channel sediment storage. These bars terminate some 4 km before the river discharge into Oshigambo Bay. It is also assumed that the discontinuity of the Oshigambo River mentioned above, contributed to the lack of sediments for deposition at the river mouth. Collectively, approximately 75% of the water budget that Etosha Pan receives is thought to come through the Cuvelai System (Groundwater Consulting Services 1990).

4.4 The rest of the Etosha Drainage system

As an endorheic basin, channels feeding directly into or towards the Etosha Pan flow virtually from all directions. The most prominent and largest is undoubtedly the Cuvelai system described above, which drains mainly from the north. The next significant inflow into the pan comes from the east and northeast, primarily via the Omiramba (*sing*. omuramba) Omuthiya and Owambo. From the west, although there is a pronounced drainage system, it does not have a direct link to the pan at present. The southern margin on the other hand has a poorly defined surface drainage, generally due to the lithology of the area. This subsection will look into relevant characteristics of these drainages and place them into perspective.

The Omuramba Omuthiya drains a system of parallel, vegetated dunes situated at the northeast of Fisher's Pan, the eastern-most tongue entering the pan (Lindeque & Archibald 1991). Much more significant in terms of volume, however, is the Omuramba Owambo, which drains the Tsintsabis area and the distal northern portion of the Otavi Highlands (e.g. Jaeger 1927; Gevers 1930; Lindeque & Archibald 1991). The upstream section of this catchment receives an annual average rainfall of 500 mm (Mendelsohn *et al.* 2002). Schwarz (1920) assumed that in the past the Omuramba Owambo had a direct link to the Omuramba Omatako, hence the Okavango River. He pointed out that the longest tributaries of the omuramba Omatako on its western flank, are projected towards concatenated vleys, which are aligned with the omuramba Owambo. These alignments were then taken to suggest that the vleys are the remnants of a stream that once connected omuramba Owambo. Older topographic maps, such as the Kriegskarte von Deutsch-Südwestafrika (Reiner. 1904), also show a tenuous link between the two omiramba. In Schwarz's (1920) reasoning, this would have therefore meant that the Kunene and Okavango rivers communicated via the Etosha Pan, with oshana Etaka and omuramba Owambo as the connecting network. In other words, the Kunene was essentially a tributary of the Okavango River, and the Etosha Pan was seemingly a 'riverine' lake. Ironically, he also cited oral accounts of water (with fish) spilling over from the Okavango River and reaching the Etosha Pan via those channels in historical times. It is interesting to note that in the region and inland of the Great Escarpment, the Makgadikgadi has the lowest elevation (893 m a.s.l., Ollier & Marker 1985). Insight from the available data is not sufficient enough, however, to substantiate or refute the hypothesis of the Kunene River as a tributary of the Okavango.

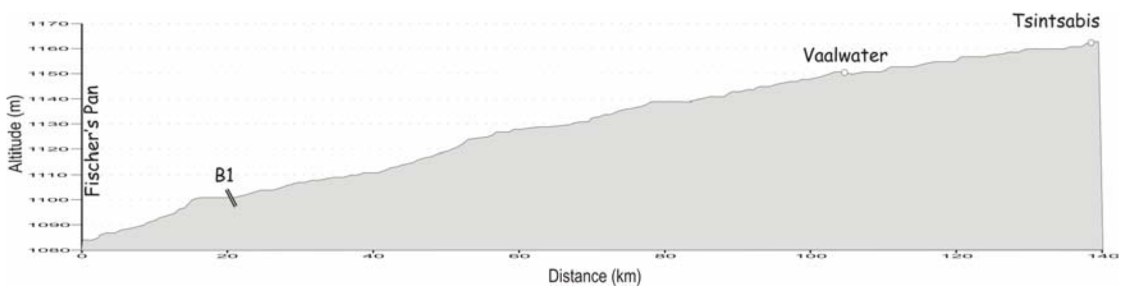


Figure 4.15: Longitudinal profile of the Omuramba Owambo derived from SRTM data (NASA). Amongst the drainages of the Etosha Basin, the Omuramba Owambo has the second steepest gradient, after the Adamax Pan channel. (Vertical exaggeration 1:355).

Among the channels that drain directly into or towards the Etosha Pan, the Omuramba Owambo has the second steepest slope gradient. The longitudinal profile of the Omuramba Owambo shows that in its 140 km distance from Fisher's Pan to its sources near Tsintsabis, the channel has a total drop of 78 m (Figure 4.15). This gives a gradient of 1:1,795, which is four times steeper than that of the Oshana Etaka.

The only drainage channel flowing from the west of the Etosha Pan has the steepest gradient (1:1400) in the area. This shallow channel traces its source from the southern section of the Otavi Dolomite Hills and terminates in the Adamax Pan. It has a length of approximately 72 km. The high gradient steepness that characterizes its profile is attributed to the fact that it drains a gradually rising area, while its headwater region receives an average rainfall of less than 300 mm annually. The Adamax Pan, in which this channel terminates, does not have a direct surface link to the Etosha Pan today. The area between the Etosha and Adamax pans is presently covered with dune ridges, which are considered to have blocked off the inflow channel. Thus, the Adamax Pan is essentially a lagoon that was formed as a result of water that was backed up against the dune ridges. Interestingly, Adamax and Natukanaoka pans are tenuously linked by a very shallow channel, and water between the two pans may flow to either direction. There is, however, an interference of this occasional flow due to a damming of the channel approximately 7 km from the Adamax Pan.

The lack of drainage lines west of the Etosha Pan is compensated for by the presence of numerous pans, most of which being less than 30 m across (Figure 4.16). The most conspicuous area dominated by this landform feature is locally known as "pannetjiesveld", which has a pan density of up to 30 pans per 10 km^2. This area is virtually flat, covered with either low tree savanna or shrub savanna (Sannier *et al.* submitted). The area is also characterized by moderate to very high water-holding capacity, either in the top- or sub-surface layers (Beugler-Bell & Buch 1997). Annual average rainfall in this area is below 300 mm. Most of these pans are covered with dark clays.

Figure 4.16: Example of innumerable (up to 30 pans per 10 km²) minor pans of western Etosha National Park. The mud hole in the centre of the pan was dug by elephants. For well over a century, animals have been dubiously cited as one of the prominent geomorphological agents in the origin of pans (e.g. Alison 1899; Passarge 1904). Their role in pan development, however, can only be extended to opportunistic pan enlargement or deepening (secondary modifying processes), as recorded here.

As the only means in which this area drains itself, the development of these pans is ascribed mainly to a delicate interaction between topography, climate, soil and vegetation. This inference was fostered owing to the fact that, in general, pans in the region occur in areas of low topographic relief, that they are associated with semi-arid rather than with arid or hyper-arid climates, and that they arise in poorly drained areas (e.g. Graf 1988; Verhagen 1990; Goudie 1991; Goudie & Wells 1995). Other key elements for their development are the absence of fully integrated fluvial processes and a lack of aeolian accumulation which obliterate the infilling of any irregularities in the land surface (Goudie 1991; Goudie & Wells 1995). Thus the modest moisture that the area receives, often in the form of localized thunderstorms, is not sufficient enough to initiate fully-fledged surface channels, while at the same time, given the low infiltration rate, the received amount is sufficient enough to withstand evaporation for a short period. Coupled with a near-level topography, a predisposing condition for promoting standing water is therefore set. In the wake of controversies surrounding the development and modification of pans, it is essential to differentiate these two processes and that the above understanding is limited to the prerequisite setting and the incipient formative processes.

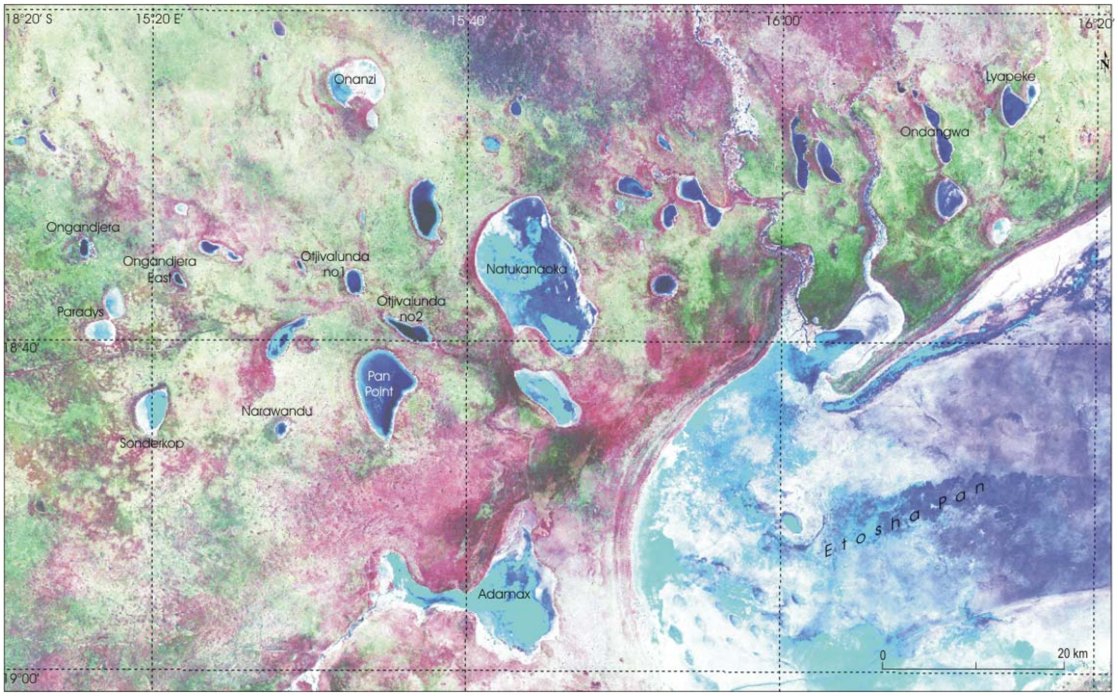


Figure 4.17: Landsat TM mosaic showing the locations and distribution of medium-size pans around the Etosha Pan. In principle, the blue shades signify higher moisture content, the reddish represent low vegetation cover, whilst the green one reflects green vegetation.

The minor pans discussed above are morphologically different from the medium-size depressions, such as salt pans, which occur predominantly in the northwest of the Etosha Pan. In addition to the sedimentological differences, the two types of pans also differ in the sense that the medium-size pans are deeper on their eastern sides, which are usually rimmed with calcrete, and that their lee sides have smooth surfaces, mantled with the so-called lunette dunes. In that respect, they have affinities with the Etosha Pan, as we will find out later in the following chapter. Moreover, all of these pans are connected by the 1,100 m contour, a level used by Buch (1997) to trace the incipient, erosional form of Etosha Pan at the Pliocene / Early Pleistocene.

Prominent among other pans of this character are Adamax, Narawandu, Natukanaoka, Onaiso, Ondangwa, Ongandjera, Otjivalunda, Pan Point, Paradys and Sonderkop (Figure 4.17). These pans can be broadly categorized into salt and non-salt pans. Sediments of non-salt pans, such as Sonderkop, Paradys, Natukanaoka and Adamax are very shallow. Buch & Rose (1996) documented that a weathered sedimentary rock is present within a depth of 40 to 80 cm, whilst the parent bedrock, the Andoni Formation, is recorded around 100 cm. The surface cover is interpreted as parautochthonous, with textural, mineralogical and chemical properties very similar to those of the underlying layers.

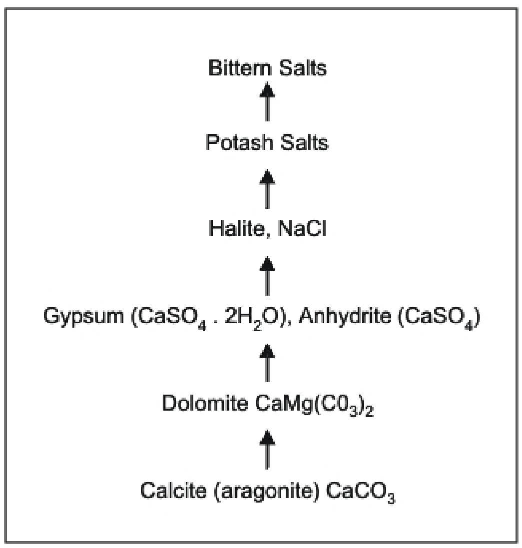


Figure 4.18: Succession of evaporate minerals after Usiglio's (1849, cited in Davis 1983) experiments. The first mineral to precipitate is on the bottom (redrawn from Davis 1983).

*Figure 4.19: Otjivalunda salt pan. The thickness of the surface salt varies from place to place and from year to year due to solution by rain water and redeposition (Schneider & Genis 1992a); this salt layer rests on black mud (inset) about 30 cm in thickness. The mud is underlain by a layer of thenardite and trona reported to have a thickness of more than 1 m (*Schneider & Genis 1992a*).*

Seven of the medium-size pans, starting with Otjivalunda, are known for sizable deposits of halite (NaCl), thernadite (Na_2SO_4; sodium sulphate), and/or soda ash (Na_2CO_3) with lesser amounts of other complex sodium salts (Figure 4.19) (Foshag 1933; Schneider & Genis 1992a, 1992b; Miller 1997). The two Otjivalunda pans totaling a combined surface area of 12 km^2, for example, have a deposit of 190 000 tons of halite, and 6.5 million tons of thenardite and trona ($NaHCO_3$, sodium carbonate) (Schneider & Genis 1992b; Miller 1997).

The presence of halite and other salts in some of these pans can be related to the evaporate sequence first recognized by an Italian chemist Usiglio (1849, cited in Davis 1983) (Figure 4.18). In his genetic approach to saline lakes, for example, Davis (1983) applied this sequence to the precipitation of minerals sequence. This sequence demonstrated that as the brine becomes more and more concentrated, the first minerals to become supersaturated are carbonates. Continued concentration causes co-precipitation of carbonates and gypsum. After much more concentration of the brines, halite is precipitated. Eventually, the lake either dries up or become ephemeral, but only after (considerable) accumulation of evaporates have precipitated.

Given the maturity stage of the mineral precipitation of some of these pans, combined with the altitudinal location, their relative density, size and proximity to the Etosha Pan, it is probable that the development of these pans is related to that of the Etosha Pan. They might be, for example, vestiges of a much larger lake, left behind as isolated sub-basins after the Kunene River cut ties with the Owambo Basin. As a result, the pans evolved independently to their present character as individual entities. Like the Etosha Pan, however, Buch & Rose (1996) believed that these pans are incised into a sandy variation layer of calcrete, which overlies the Andoni Formation found at about 100 cm below the surface of the non-salt pans.

In contrast to all other sides around the Etosha Pan, the southern section is devoid of surface drainage in the upper catchment, whilst the lower part has very short, isolated and poorly defined drainage lines. This unique characteristic is due to its lithology. The Etosha catchment in this area has its source in the Otavi Mountains, formed of carbonate rocks. A pediment zone consisting of calcrete mediates between the Otavi Dolomite Mountains and the Etosha plains (Buch 1997; Miller 2001). Although Miller (2001) acknowledged the occurrence of pedogenic calcrete in the area, he regarded much of this calcrete as non-pedogenic or "groundwater" in origin. His model associated this calcrete cover with springs along the contact zone between the Otavi Group dolomites and limestones and the largely impervious Mulden Group arkoses and phyllites (Figure 4.20). Having the spring water subjected to ensuing evaporation, the precipitated calcrete (*not tufa*) was then deposited at and downhill of the springs. With time, the Mulden phyllites were covered with this calcrete layer, as the spring water seeped through the calcrete and emerged in places, particularly at the furthest edges of the calcrete.

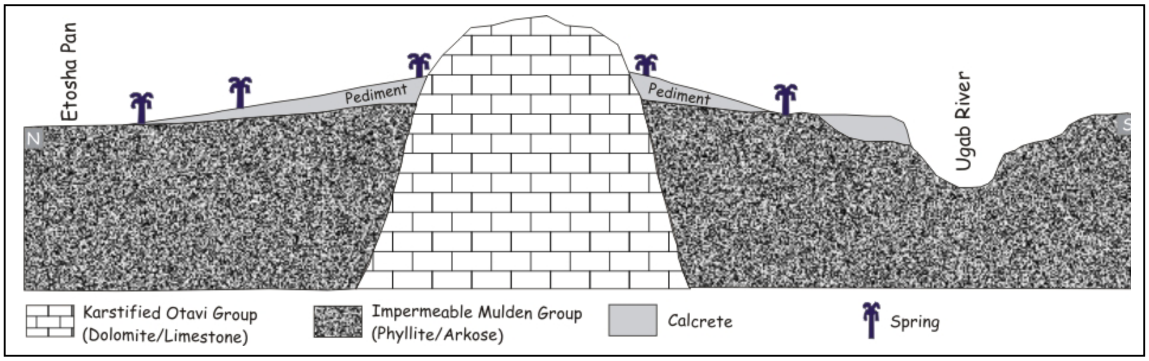


Figure 4.20: Idealized cross-section through the karstified Otavi Group carbonate rocks and impervious Mulden Group phyllites and arkoses illustrating the development of "groundwater" calcrete in the southern section of the Etosha Pan's catchment. Primary springs at the contact of the carbonates and phyllites are buried beneath their own calcrete and secondary springs emerge down slope where the thickness of the calcrete decreases (redrawn from Miller 2001).

Downhill-spreading of calcrete was also accompanied with its thickening, which ultimately buried incipient springs. Buch (1993) reported, for example, a thickness of this calcrete to be in the range of up to 50 m along the southern rim of the basin. As the distance between the Otavi carbonate rocks and the newly emerging springs increases, the calcrete interfingers with clastic Kalahari sediments. Through substantial karstification of the calcrete, its porosity and permeability is enhanced. The manifestation of this process today, albeit subdued, is believed to account for the presence of springs in the area, such as Gobaub and those at the southern edge of the Etosha Pan, which are regarded to be fed by this sub-surface contact.

Field evidence in support of the above hypothesis lies in the zonal composition of the calcrete with increasing distance from the source. Away from the Etosha Pan, near the zone of contact springs, calcrete is reported to be composed of pure calcium carbonate. Closer to the Etosha Pan, however, the calcrete becomes more dolomitic, with dolocrete boulders found in and around the Etosha Pan itself. This change in calcrete composition is taken to have resulted from the greater solubility of dolomite, where typically calcite gets deposited closer to the source, while dolomite further away (Miller 2001).

However, Rust (1984, 1985) and Talma & Rust (1997) argued earlier that much of the calcrete surface that characterize most of the Etosha region, including the southern section, is pedogenic. Their conclusion was based on the absence of the A-horizon and a widespread combination of a weathered, Bv horizon and a calcareous, Ca horizon.

Due to the presence of this calcrete pediment, surface channels from the southern section of the Etosha Pan developed almost exclusively, further in the plains. These channels are usually of the order of 10 km, with few exceptions of one or two channels that are up to 30 km in length. The most notable channels are the Gaseb in the southwest and the Springbokfontein in the southeast. Both channels are broad and shallow, forming noticeable fan delta on the floor of the Etosha Pan (Figure 5.5).

In summary, the primary source for surface inflow into the Etosha Pan is from the north and east. Groundwater flow, on the other hand, tends to be more significant from the southern section (Rahm & Buch 1997), as evidenced from the presence of numerous springs and seepages. The west, with its appreciable internal drainage and being drier than the rest of the catchment, contributes significantly less to the water budget of the Etosha Pan at present.

CHAPTER 5

The Etosha Pan

Introduction

The build-up of previous discussions was directed towards the presentation of the Etosha Pan in its regional setting. Although the coverage was largely selective, it highlighted amongst others that the assertion of the geomorphic relation between the Etosha Pan and the Kunene River is credible. What came short was the fact that the timing of when that connection ended was not firmly secured. Nevertheless, the following chapter, devoted to the Etosha Pan, will take place with full cognitive of that background and the constraint that was presented. In so doing it will also avail an objective evaluation of the existing theories regarding the development and evolution of the Etosha Pan.

5.1 Etosha Pan

The Etosha Pan is defined by Shaw (1988b) as a large, fault-defined endorheic basin. It is situated in the southern margin of the larger Owambo Basin, discussed in section 3.2.3. Today, the pan is a regional sump, with much of both surface- and ground-water flowing towards or directly into it.

The surface area of the Etosha Pan measures approximately 4,760 km^2 (Lindeque & Archibald 1991). Its major axis has a maximum extent of 120 km, whilst its width is up to 55 km across. As already determined by Jaeger (1926), the major axis of the pan has an ENE – WSW orientation (Figure 5.1). Although its surface is markedly flat, the pan is slightly tilted to the east. The eastern section is about 5 m below that of the west. This sets the pan's gradient to an average of 1:22,400. Occasionally, differential surfaces a few decimeters high and few meters across are encountered on the surface of the pan. The marginally higher areas are typically coated with saline flakes on the surface.

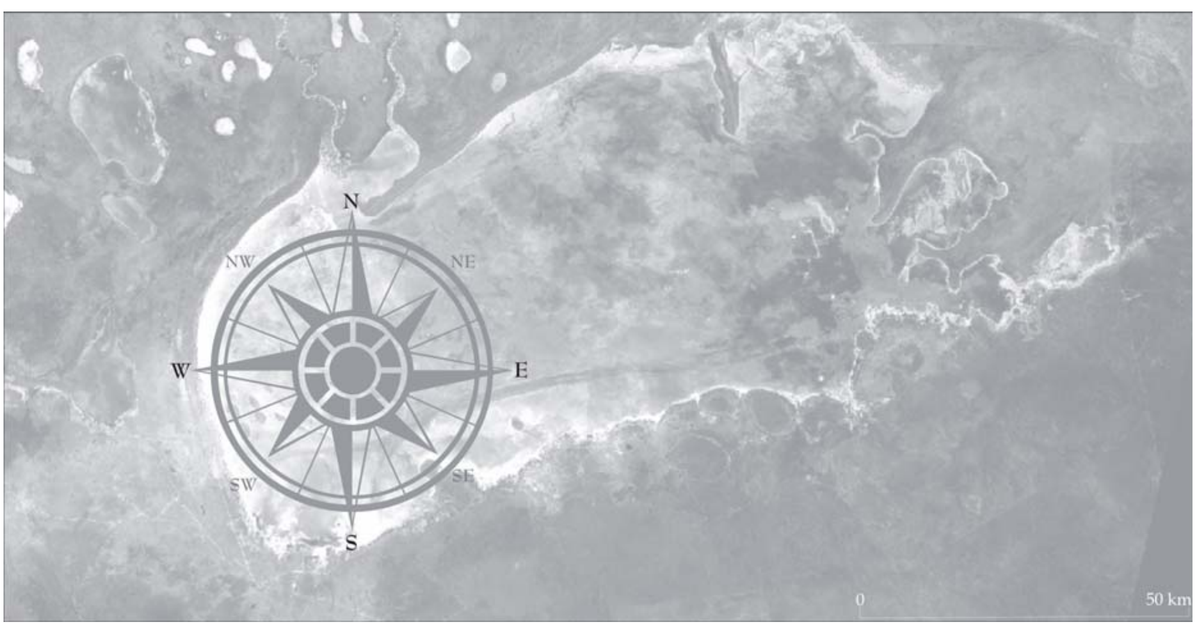


Figure 5.1: The orientation of Etosha Pan

Apart from the Digital Elevation Model (DEM) of the pan floor, both its flatness and gradient are best illustrated through the aid of remote sensing data captured around the rainy season. Figure 5.3a captures the effect of widespread local rainfall in the area, which resulted in a pan-wide coverage by a thin layer of water, elucidating the horizontal surface of the pan. With a significant inflow, however, the pan's gradient tends to be exploited as depicted in Figure 5.3b.

On the floor of the pan itself, there is a straight and very shallow channel that originates from the southwestern part of the pan and runs eastward (Figure 5.3b; 5.5). Started off in the vicinity of Wolfness, now a dry spring, this channel has a couple of short, first order channels. At its upstream, the most prominent stream from the southern side connects with the Gaseb channel, whilst the major one from the northern direction links with minor pans located near Okondeka waterhole. Water that enters the pan from both the Ekuma and Oshigambo rivers also find its way to the east, via a broad, ill-defined waterway situated on the floor of the pan. The depth of both channels is only a few centimeters and they are not easily detectable in the field.

There are also a limited number of places on the pan's surface where water collects locally in small depressions, hardly perceptible on the ground, but salient in remote sensing data. Additionally, numerous springs and seepages occur at the southern pan margin (mainly around Halali), the western shoreline and in the middle of the pan. For those located in the middle of the pan, one such spring is located some 3.5 km from Logan Island and at present forms a 2 m high mound (Figure 5.4b). Another spring is located five kilometers southeast of the same island. Other springs are located near the Springbokfontein alluvial fan, in the very centre of the pan, and at the Oshigambo Bay. The majority of these springs have water flowing radially from the eye, which in turn gives a circular shape to the site where a spring is located.

Apart from the more familiar Halali seepages, most of the little known ones are located in the vicinity of Pelican Island. These seepages take a linear shape, oriented mainly in the north-south direction and parallel to the island (Figure 5.4a). It is likely that they follow zones of weakness below the surface of the pan. There is also a weak seepage located 10 km north of Okondeka waterhole. Kempf & Hipondoka (2003) attributed its existence, along with Okondeka and Wolfness, to the percolating groundwater streams from Adamax Pan. A common feature between these springs and seepages is that they tend to have a temperamental flow intensity. Like springs, their surroundings are also covered with salt flakes.

Sediments of the pan, particularly in the subsurface, are chiefly composed of olive-green clay. Two representative assays obtained from Eto 10 and Eto 216, taken respectively at a depth of 0-20 cm and 110 cm, reveal that more than 95% of the pan sediments is mud, of which clay constitutes more than 80% (Figure 5.2).

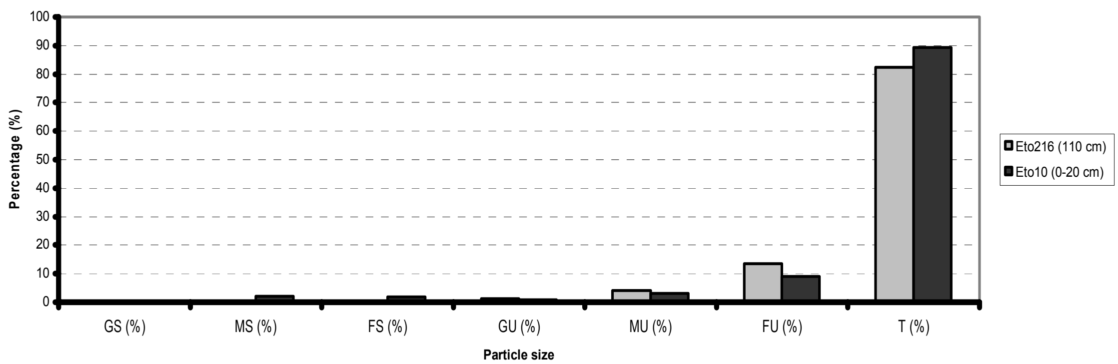


Figure 5.2: Grain size distribution of pan sediments from two typical profiles. Key: GS – coarse sand, MS – medium sand, FS – fine sand, GU – coarse silt, MU – medium silt, FU– fine silt, T – clay.

An assortment of geomorphological units in the area is predominately associated with lacustrine, fluvial and/or aeolian processes. These units are presented in Figure 5.5. Prominent among these geomorphological units are sandy beach ridges, lagoons, delta, fan delta, islands, flood plains, bays, and peninsulas. The only delta in the area is formed by the meandering Ekuma River (section 4.3).

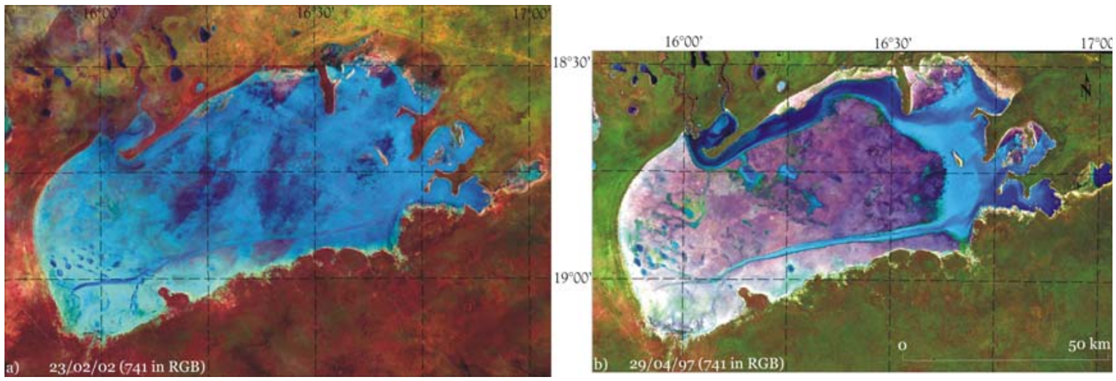


Figure 5.3: Landsat images betraying the topography and gradient of the Etosha Pan; a) pan-wide coverage of a thin layer of water resulting from local rainfall reflects and illustrates the pan's horizontal topography b) water inflow from the catchment area accentuating the subtle eastward slope of the pan.

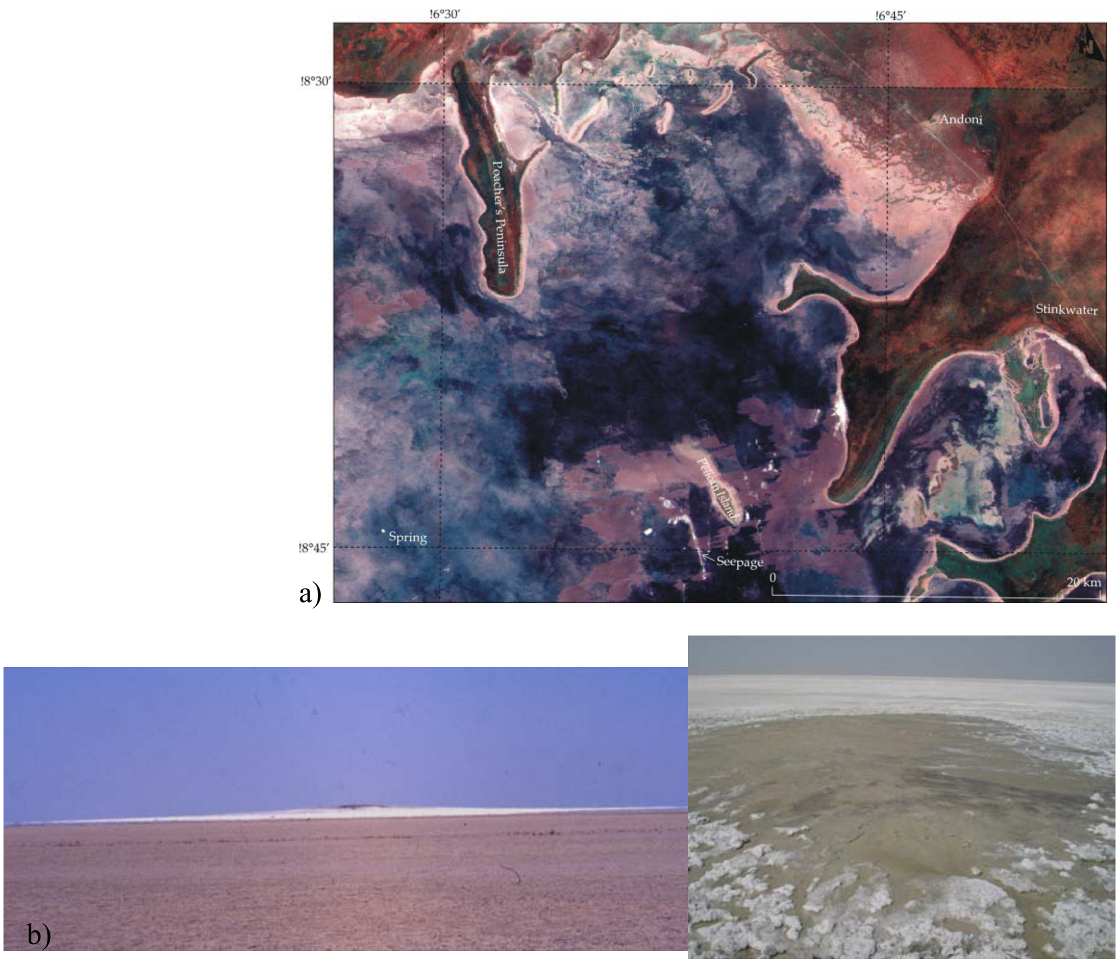


Figure 5.4: Sample of circular springs in the middle of the pan and seepages; a) location b) horizontal perspective of the 2 meter mound-spring from a distance, and an inset of an eye.

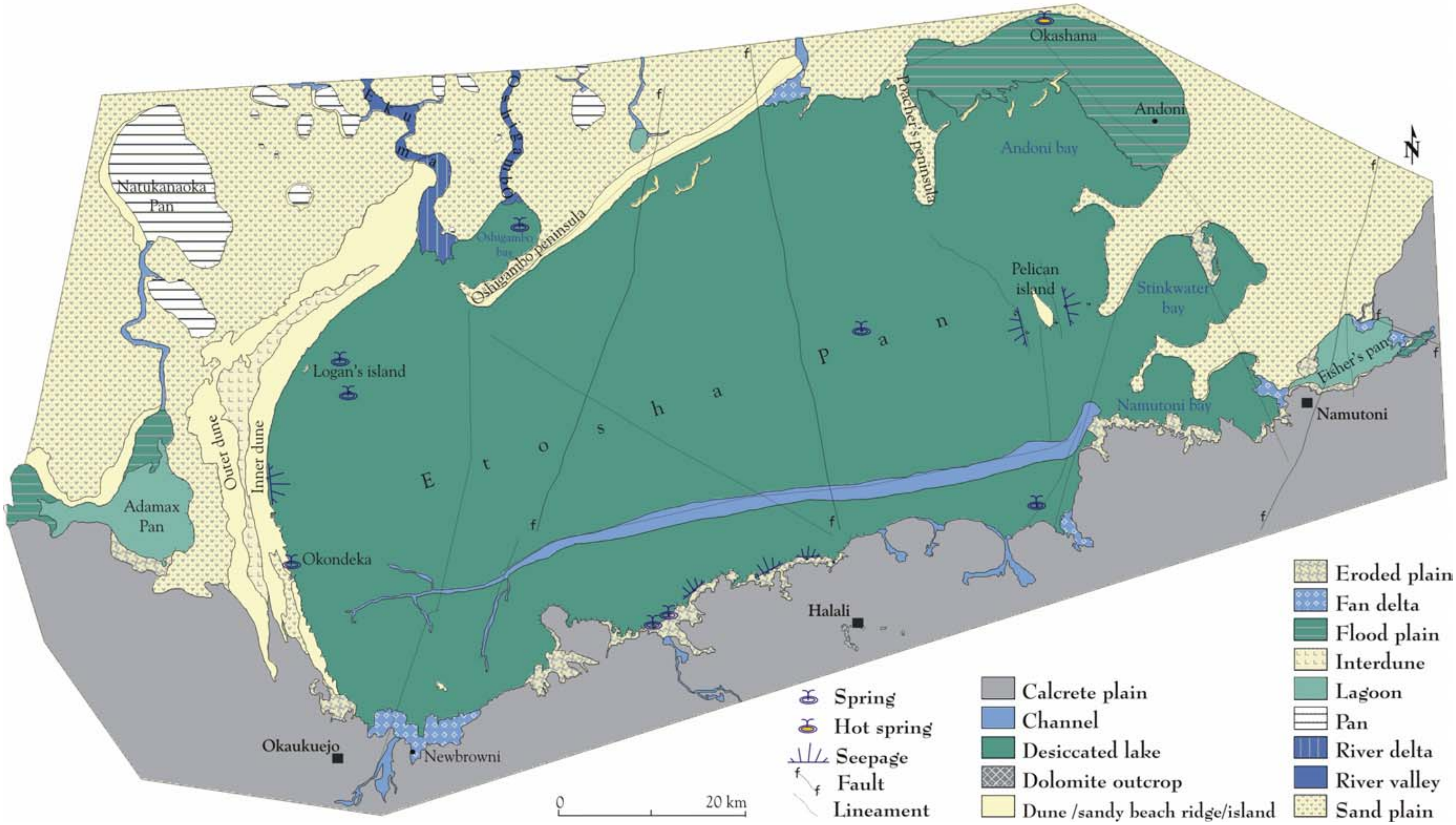


Figure 5.5: Generalized geomorphological map of Etosha Pan and its environs.

Five main alluvial fans occur on the surface of the pan. These alluvial fans are formed at the entry points of selected feeder channels. Niipele, Omuthiya, Fisher's Pan channel, Springbokfontein and Gaseb channel all have produced noticeable alluvial fans. They average 9 km^2 in size.

Five bays, Oshigambo, Andoni, Stinkwater, Namutoni and Springbokfontein were delineated. The Andoni, located in the northeastern part of the pan, is by far the largest. It fades into a floodplain, covered with grass (section 3.6).

Terraces of various heights and material compositions border with the pan margin. The northwestern and western section has multiple ridges, believed to be beach dune ridges, located at varying distance from the pan margin. They are often grouped into inner and outer "dunes". Likewise, the northern part, west of Niipele alluvial fan possesses similar beach dune ridges or beach bars. Whereas a smooth, regular shoreline with sandy beaches features in the western margin of the pan, the plan outline in the eastern side is rather irregular and is intervened by peninsulas with pronounced scarps in places. With the notable exception of the Mushara sand ridge, the eastern flanks are often lined with calcrete walls or capping.

Adamax and Fisher's pan are the most prominent lagoons in the area. Under current geomorphologic configurations, Adamax Pan, along with the only channel that drains from the western sector of Etosha Pan, has no direct surface connection to the pan (section 4.4). The 8.5 km stretch between Adamax and the western section of Etosha Pan is covered with beach dune ridges. The Fisher's Pan on the opposite end has a direct connection to the Etosha Pan. However, water may flow to either direction via a narrow channel that links them.

The Pan is transected by two major faults, with a third skirting it at the eastern side. These faults are aligned broadly in a north-south direction. They reflect little sensitivity to local topography. Numerous lineaments are also delineated in the area (Figure 5.5)(Kempf & Hipondoka 2003).

Recent studies model Etosha Pan to have formed in the late Quaternary by virtue of pluvial erosion of intervening scarps (1 to 20 m high). In tandem with aeolian deflation during the dry season, the pluvial erosion had subsequently led to the coalescing of several smaller pans, which ultimately resulted in a super-pan that we call Etosha today (Rust 1984, 1985; Buch & Zöller 1992; Buch *et al.* 1992; Buch 1993, 1997; Buch & Rose 1996). Within the latitude of this postulation, the Kunene River is specifically and emphatically ruled out as a player in the evolution of Etosha Pan.

The model introduced above unseated a previous assumption that evolved from the 1920s (e.g. Schwarz 1920; Jaeger 1926) and matured with Wellington (1938). Building on results from previous studies, the latter author conceived Etosha Pan as a desiccated palaeo-lake that resulted from the 'capturing' of the proto-Upper Cunene by a fast flowing, coastal stream, the proto-Lower Cunene (section 4.2). Eventually, this cut-off lake was believed to have degenerated into its present dried-out condition.

Despite the controversy surrounding the evolution of the modern day Etosha Pan, a shallow, low energy and saline-alkaline lake is generally recognized to have existed as a terminal evolution of the Owambo Basin. That terminal stage was estimated by Buch & Rose (1996) and Buch (1997) to have taken place during the Oligocene and Miocene. From the Etosha Pan itself, evidence for such a lake was deduced from the presence of stromatolites at Poacher's Point and Pelican Island (Martin & Wilczewski 1972; Smith 1980). However, these fossils were variously dated to a Pliocene age (Martin & Wilczewski 1972), Early to Middle Pleistocene (Smith & Mason 1991) or are assigned a minimum age of Upper Pleistocene (Talma & Rust 1997). Assuming these stromatolites were eroded from the Kalahari sediments, the latter authors employed stable isotope (^{13}C and ^{18}O) analysis in deriving the age estimate. Such an assumption of assigning a wide range of age to the Kalahari sediments illustrates the illusive nature of this environment and why the age of the Kalahari Group sediments is still unknown.

Smith & Mason (1991) identified the stromatolite-bearing sediments of Poacher's Point to comprise the lower green clay, oosparite, oomicrite and an upper layer of calcrete. Apart from stromatolites, these sediments are also rich in ooids and silica nodules. The ooids are reported to have formed around micritic peloids, ostracod valves, detrital grains of goethite, and detrital grains of quartz. Radial-fibrous calcite laminae are the main composition of the ooid cortices. The presence of these laminae is taken to be indicative of relatively calm water. Their various shapes, ranging from spheroidal to ellipsoidal and asymmetrical tend to point to the environment in which they were formed. The former two types are thought to have developed in calm, sheltered conditions, whilst the latter is attributed to an intermittently agitated and calm setting (Smith & Mason 1991).

The presence of stromatolites does not necessarily provide unequivocal evidence for saline conditions (Gray 1988). They are known, for example, to exist in Tertiary lakes of the northern half of France (Freyte *et al.* 2001) and along the middle Yangtze River in China (Geyer, G., personal communication, 2005). The only variety that is exclusively of saline environment has a columnar structure visible in thin section. Those that grow in a marine environment attract sulphate to their vacuoles which then turn into gypsum crystals.

The lack of authoritative age of the Kalahari Group sediments, particularly the upper Andoni Formation, in which these stromatolites are hosted, has a detrimental effect to the reconstructions of the evolution of the Etosha Pan. Within that limitation, however, there exist proxy data that once evaluated systematically, might lead to a plausible model of the evolution of the Etosha Pan. These data accumulated mainly from studies that examined the area, despite the fact that their applications led to diverging conclusions with respect to how the modern Etosha Pan came into being. The following sections will therefore look into details of what led to the formulated theories regarding the Etosha Pan. In tandem with findings of this study, substance of each argument will also be assessed and evaluated as they are introduced and discussed.

5.2 Etosha Pan as an erosional landform

The theory of Etosha Pan as an erosional landform was developed by Rust (1984, 1985). His model commenced with the recognition of a continental unconformity layer covering the entire surface situated inland of the Great Escarpment. This continental planation surface is of unknown age and is represented by a fully developed soil (MBv-ca-C) which specifically covered the Otavi, Nosib and Kalahari Group sediments. Stratigraphically, the calcrete (pedogenic) horizon is overlain by iron-coated sands. It was on this old planation surface that complex landforms developed opportunistically and on the basis of pedogenetically inherited lithologic varieties found above, at and below the MBv-ca-interface. Thus the highly developed landforms were predominantly pluvial (endorheic relief) in origin, where differential and intermittent denudation of the old surface layer began.

According to Rust (1984, 1985), the above-mentioned process occurred specifically on the then surface of the Etosha Pan. The accident of Etosha Pan's surface was determined by the superimposition of the epeirogenetic (Etosha sub-basin) and stratigraphic (Kalahari sediments) conditions. Denudation that ensued followed the morphology of cuesta landforms, in a form of scarp retreat process. This process brought about the coalescing of smaller pans by means of broad valleys and hose-shaped (*sic*) features. The intervening old surfaces on their part, dissolved into peninsulas and islands that feature in the area (Figure 5.5). The resulting endorheic relief has a maximum range of 20 m. Over time, the net result of the above processes was the development of a super pan that we call Etosha Pan today, which simply formed by the fusion of scarp retreating of several pans. Due to the fact that the pan developed in stages, some of the dated materials gave dates which exceeded the carbon-14 method that was employed. Thus the date for initial formation of the pan could not be firmly constrained by Rust (1984, 1985).

In its current form, the above model has an obvious weakness, as it offered no explanation to the origin of the base level, given the fact that the pan itself is already the largest and deepest part of the fluvial system, most notably the Cuvelai. Moreover, the export of material from the erosional base level was not duly treated. It is nevertheless recognized that the process of endorheic pluvial erosion was augmented by both the aeolian and fluvial processes as we are about to witness it below.

In advancing the hypothesis discussed above Buch *et al.* (1992) and Buch (1993), for example, re-affirmed the formation of a thin sediment cover during the Quaternary. This cover was also situated on what was later to become the surface of Etosha Pan. These authors assumed that since at least the Late Pliocene/Early Quaternary, Etosha Pan was subjected to continuous denudation processes operating in seasonal rhythms of water inundation during the rainy seasons and extensive deflation processes during the dry seasons (Figure 5.6).

These denudation processes left behind two conspicuous levels at 1,100 m a.s.l. and 1090 m a.s.l., which were incised into the calcrete horizon, termed by Buch (1993) as 'Etosha Limestone'. Buch (1993, 1997) regarded this limestone as the terminal facies of the fluvial and fluvio-limnic depositional history of the Owambo Basin, which took place during the Miocene/Pliocene. In that judgment, part of the limestone developed simultaneously with the Andoni Formation. This is part of the same calcrete layer considered by Rust (1984, 1985) as pedogenic and marking a continental unconformity within the Kalahari succession; it was also discussed in section 4.4 how Miller (2002) attributed the southern section of the same calcrete layer to "groundwater" formation. Today, the thickness of this limestone or calcrete in the western side of the Etosha Pan is alleged to be not more than 15 m and 5 m below the 1,100 m and 1090 m levels, respectively (e.g. Buch *et al.* 1992). Levels below the 1085 m a.s.l., including the current pan margin, have been further incised into the limestone layer and reached the green-colored claystone, siltstone and sandstone facies of the Andoni Formation. Localized exposures of the Andoni Formation in the area are revealed only on the western flank of Poacher's Point Peninsula (Smith & Mason 1991), at the southern end of Pelican Island (Martin & Wilczewski 1972) and at a section of the eastern side of the Oshigambo River terrace (this study). The so-called Etosha Limestone (Buch 1993) has a much more significant but laterally discontinuous exposure along the margins of the Etosha Pan and river terraces in the area.

According to Buch *et al.* (1992) and Buch (1993), these underlying disintegrated rocks of the Kalahari Group sediments and the continental planation surface that covered the surface of Etosha Pan were blown out by dominant northeasterly winds. Two distinct sand ridges located in the northwestern and western margin of the pan and overtopping the 1,100 m and 1090 m levels were exploited as primary evidence for aeolian deposition of those deflated sediments. The sand ridges were interpreted as lunette dunes.

Buch & Zöller (1992) and Buch *et al.* (1992) applied pedostratigraphy and the thermoluminescence-chronological method to determine the timing of that 'lunette dune' deposition. Results from 8 TL-dates show that the longest record of aeolian sedimentation in that area dates to the last 140 ka. This date was taken as pointing to the formation of the 1,100 m level. Since these sediments rest on the Etosha Limestone that Buch (1993) assigned a Miocene to Pliocene age, Kempf (2000) articulated how baffling it is that such an assumption entails that the intervening 5 million years were morphologically static. For that reason the age of the so-called Etosha Limestone and the younger part of the Andoni Formation cannot be that old as Buch (1993) envisaged.

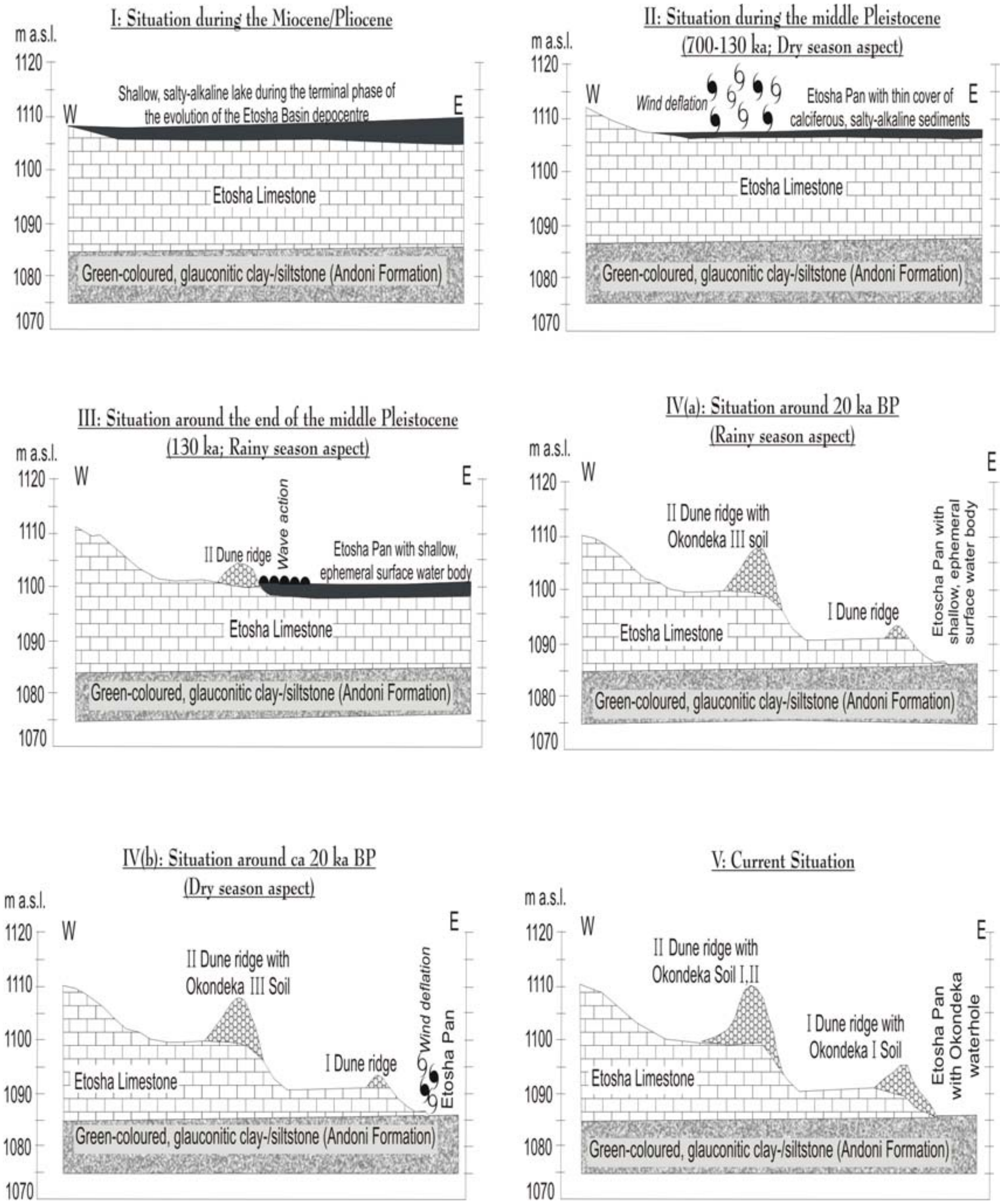


Figure 5.6: Model for the development of the Etosha Pan according to Buch (1993). The area covered for illustrating the model is western Etosha Pan, where the so-called lunette dunes occur. (Redrawn from Buch 1996, 1997).

The 1090 m level on the other hand was constrained to the minimum age of 19.7 ±4.1 ka, the age of sediments taken from the base of the inner dune. This was also approximately the time when erosion in some parts of the pan is assumed to have reached the boundary surface between the Etosha Limestone and the Andoni Formation. Deposition of sediments near the surface took place in the Middle to Late Holocene. For the inner dune that time was estimated at 5.6 ± 2.2 ka, while for the outer dune the age of 2.4 ± 0.5 ka was derived. The overall pedostratigraphy record provided no indication of substantial interruption of sediment removal from the pan floor. For that reason reported pluvial phases and the presence of a perennial lake at Etosha Pan particularly at around 12 ka (Heine 1979) was repudiated by among others, Buch *et al.* (1992) and Buch (1993), as they were unable to substantiate it with their respective findings.

To validate the origin of sediments that made up the lunette dunes, provenance was factored into the investigation. Buch *et al.* (1992) quantitatively determined heavy minerals of a sample taken at the southern part of Logan's Island (Figure 5.12), and qualitatively at two locations located directly across from that island, at an interdune depression and the outer dune (1,100 m level). The profiles yielded similar mineralogical results, with major constituents of pale-pink garnet, dark-green amphiboles, and dark-bluish-green to olive green tourmaline as a secondary component. Staurolite, epidote/zoisite, kyanite, zircon, xenotime and rutile were also encountered, along with traces of andalusite, corundum, monazite, titanite, and opaque minerals. Conclusively, the heavy mineral associations between sediments taken at Logan's Island and those sampled from the interdune and outer dune were regarded as indicators for a shared genetic relationship between these units. In this setting, Logan's Island is interpreted as a remnant of a thin layer that once covered the rest of the modern surface of Etosha Pan, whilst the dune and its associates are made up of those materials that disintegrated and were then blown out to make way for the pan (e.g. Buch 1993).

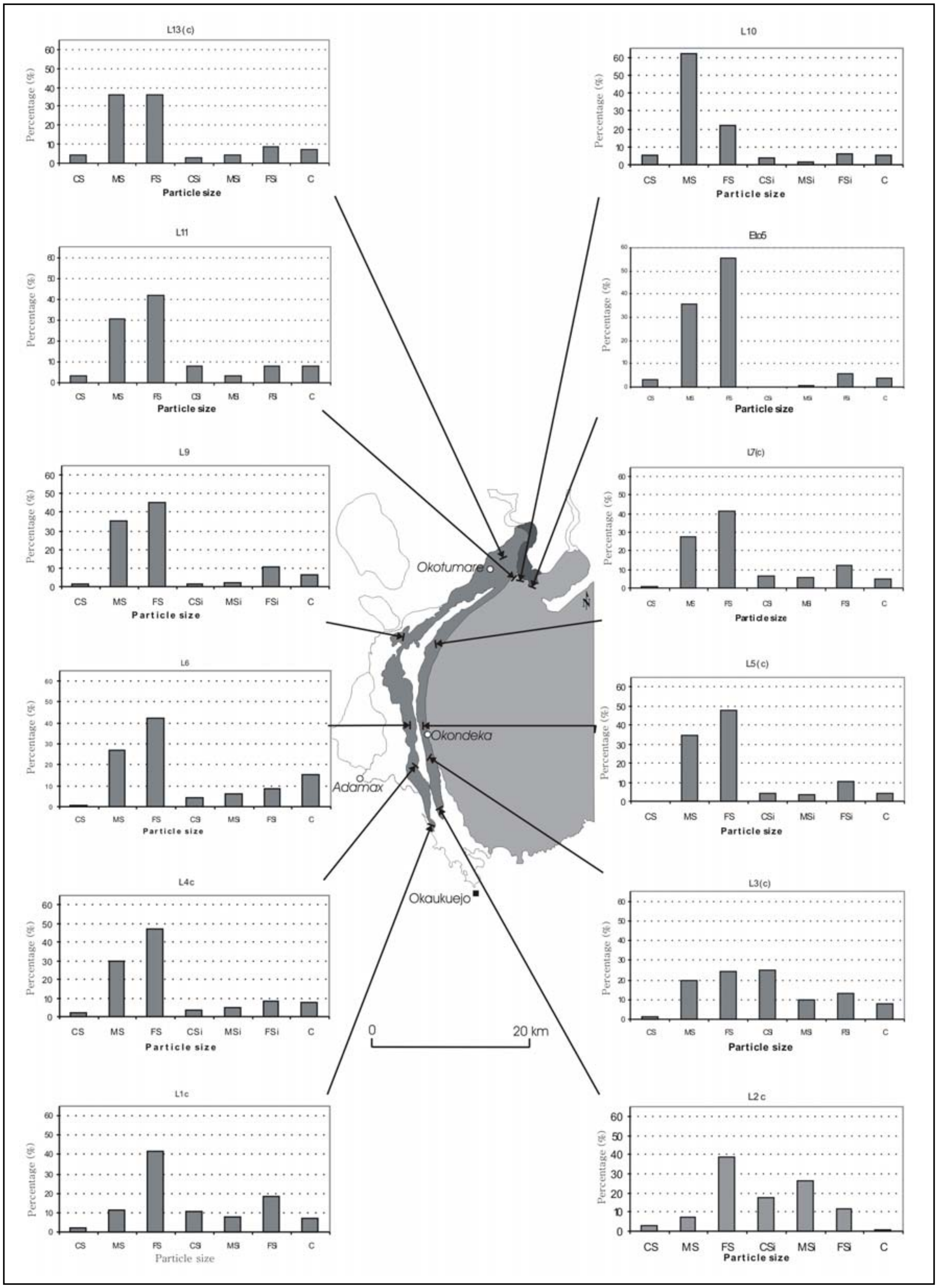
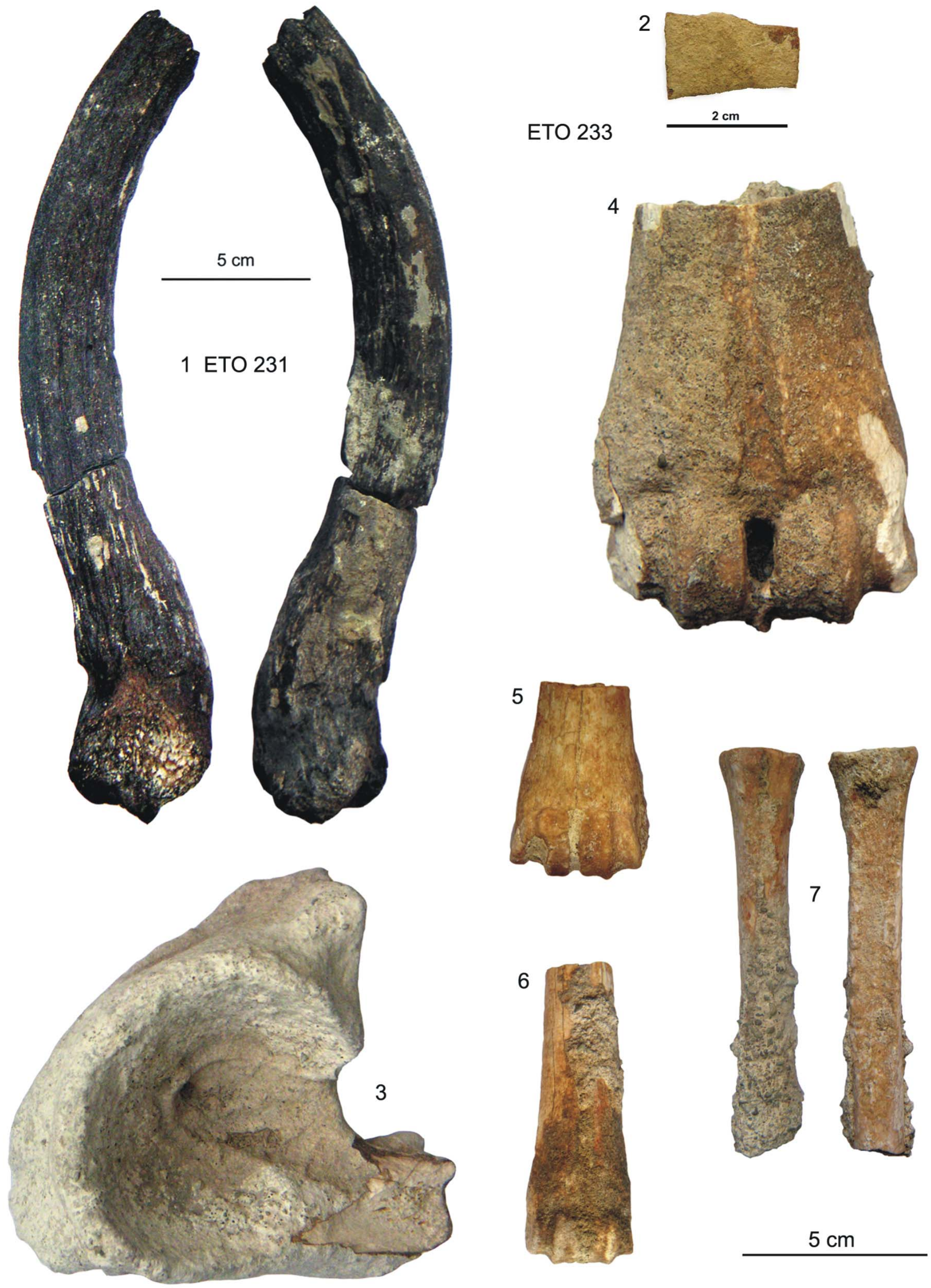


Figure 5.7: Granulometry results of selected soil samples along the dunes and at Ekuma delta. Key: CS – coarse sand, MS – medium sand, FS – fine sand, CSi – coarse silt, MSi – medium silt, FSi – fine silt, C – clay. (After Hipondoka et al. *2004a)*

However, Hipondoka *et al.* (2004a) revealed that sediments of these lunette dunes are distributed almost in an orderly fashion, with coarser particles close to the Ekuma Delta and finer ones deposited farther from the delta (Figure 5.7). Additionally, the lunette dunes are found on the northwestern and western side of the Etosha Pan, whilst the declared wind vector that is believed to have played a formative role in the dune's development, blows from the northeast. Thus there exist a discrepancy between the spatial setting and orientation of the dunes and the identified dominant wind vector. In reconciling this inconsistency and by recognizing the textural gradient of the dune sediments that coarsen with decreasing distance from the delta, Hipondoka *et al.* (2004a) concluded that these sediments have not necessarily originated directly from the basin floor of the pan. They resulted instead from sediments transported to the downwind margin of the water body by wave action during the pan’s full stages, with the inlet supplying the bulk of the materials. That process is analogous to coastal dune development (e.g. Campbell 1968; Thomas *et al.* 1993).

As Figure 5.5 shows, there are also similar sand ridges along the southern section of the Oshigambo Peninsula. These ridges were found in this study to bear a host of fossils belonging to a variety of mammals. Presented in Figure 5.8, and determined by Jousse (University of München, 2005), these fossils include a pelvis of a *Perissodactyle* (equid or rhino), *metatarsal* of a *Tragelaphus spekii* (sitatunga) and *metracarpal* of an *Aepyceros melampus* (impala) and of an unidentified small antelope. These fossils were located in an eroded dune beach ridge, at Eto 233 (Figure 5.9, 5.15). The eroded area is littered with fragments of fossils, while others are buried in these loose sediments. The frontal section of the dune ridge has been affected by illuviation for a distance in excess of 150 m, running parallel to the pan margin.



ETO 231
1. Indeterminate antelope horn

ETO 233
2. Ostrich egg fragment
3. *Equus quagga* (quagga) pelvis

4. *Tragelaphini oryx* (eland) metatarsal
5. Indeterminate small antelope metracarpal
6. *Tragelaphus spekii* (sitatunga)
7. *Aepyceros melampus* (impala) metacarpal

Figure 5.8: Major fossils found at Eto 231 (Ekuma Delta) and Eto 233 (eroded beach dune ridge). Fossils positively identified by Jousse, H., University of München, 2005.

Figure 5.9: Photos showing a section of the Oshigambo shorelines at Eto 133. a) A litter of fossil fragments eroded from the dune ridge. b) Rib fossil of an unidentified mammal, protruding from eroded dune sediments.

Figure 5.10: Approximate present-day geographical distribution of Tragelaphus spekei *(sitatunga) on the continent (Source: Skinner & Smithers 1990).*

Among the mammals listed above, sitatungas do not occur anywhere near the Etosha Pan today due to their ecological requirements. Amongst other needs, it thrives in environment with permanent water and it is considered as the only truly amphibious antelope on the planet. As Figure 5.10 depicts, the nearest place from Etosha where sitatunga occur is now the Caprivi and the Okavango swamps, located some 600 km away (Skinner & Smithers 1990; Estes 1991).

Figure 5.11: Common gastropods in the environs of the Etosha Pan; a) gastropods scattered in loose sediments at Logan Island (Eto 234); b) gastropods coated with calcrete contained in soil sample taken at 90 cm below surface at Lun 9/Eto 212; c) Xerocerastus burchelli. *(Fossils positively identified by Sirgel, F., University of Stellenbosch, 2005).*

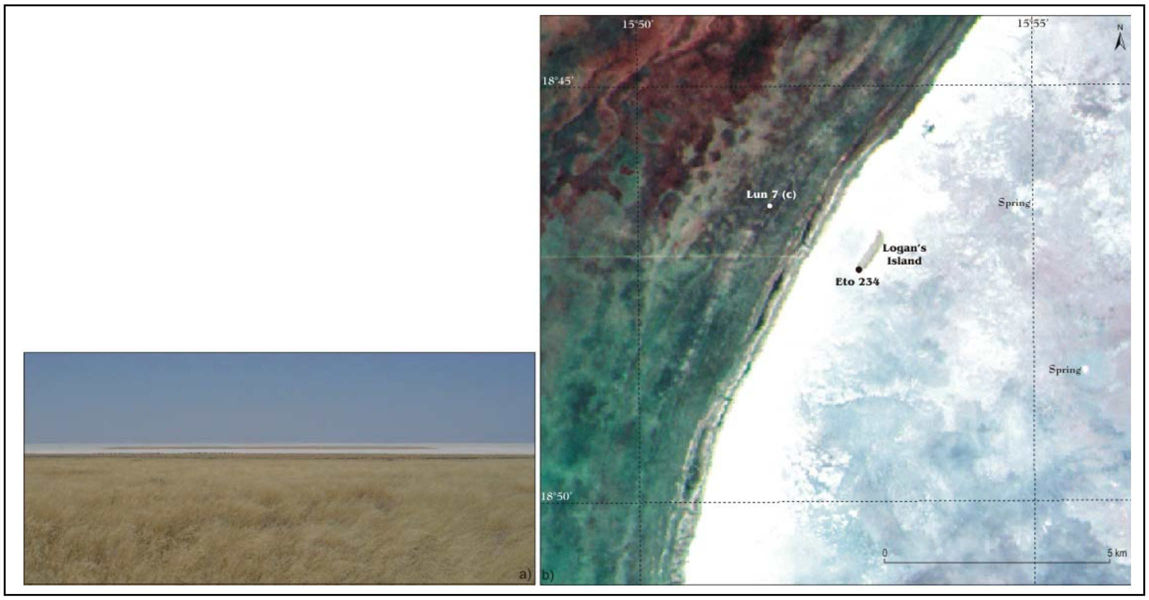


Figure 5.12: Spatial setting of Logan's Island. a) Ground perspective from the west (lunette dunes) and b) view from space (Landsat TM+, bands 748 in RGB, Aug 2002).

Figure 5.13: Geomorphological dynamics of the current shoreline at western Etosha Pan. a) Erosion of the current shoreline; b) Oblique aerial photograph of western Etosha Pan showing the damming of water behind shoreline and the prolific of beach cusps. The erosion of beach cusps, supplemented by ponding of water on the opposite side, contribute to the denudation of the barrier and hence the fluctuation of shoreline in a long-term. (Note: In 1994 when photo (b) was taken, Okaukuejo recorded a rainfall amount of 305 mm (long term average is 320 mm)).

The carbon-14 method was used to date a piece selected from these fossils. The result showed an age of 4275 ±80 BP. Using the new IntCa104 calibration curve (G. Skog, personal communication, 2005) for deriving the calendar years, the age was adjusted to a period between 4500 and 5000 years BP. This age range coincides with the sediment accumulation of the upper section of the inner dune which was revealed above from Buch & Zöller (1992) and Buch *et al.* (1992). As it is against the grain for wind excavation to take place on a perennial swampy surface, the dichotomy, in space and time, and the head-on collision in the interpretation of the data is ominously self-evident. Inevitably therefore the conclusion reached by Hipondoka *et al.* (2004a) with respect to the formation of the so-called lunette dunes is plausible. This would specifically imply that soils of the Okondeka dune complex are in situ Arenosols of former beach ridges.

Although the application of provenance was well intended in tracing the origin of the lunette dune sediments to the pan floor, it was virtually misplaced from a terrain analysis perspective. Logan Island is oriented parallel to the current pan margin, and the two units are approximately 1 km apart. Besides the fact that its sediments, consisting of 68% sand-size particles (Appendix A), are closely comparable to samples (i.e. L7c, Figure 5.7) taken directly across from the lunette dunes, sediments of the island also bear gastropods found in the so-called lunette dunes (Figure 5.11).

Specifically, these gastropods were found at Lun 9/Eto 212 located in the Okondeka outer dune (from 90 cm below the surface) and Eto 236 situated at Poacher's Point peninsula. At Lun 9/Eto 212 and Eto 236 these shells were either coated with calcrete to form calcrete-nodule-like assemblage or were interbedded in the calcrete layer. In contrast, these fossils were buried in loose sediments at Logan's island, where its flank is eroded by water. Similar gastropods were also encountered in the pan's environs by Heine (1979) and Buch (1993). The most abundant species was positively identified for this study by W.F. Sirgel (2005, University of Stellenbosch, South Africa) as *Xerocerastus burchelli*, which thrive in a wide range of ecological tolerance, including seasonally wet terrestrial environment (van Bruggen 1964). Heine (1979) assigned them to a freshwater environment, whereas Buch (1993, quoting Rust, U., personal communication) believed they were marine. They are catalogued at the National Museum of Namibia with the accession number *SMN 77105*.

The distribution of the gastropods species in sediments of different age is taken as an indicator for Pleistocene and Holocene environmental fluctuations. Although some individuals of *Xerocerastus burchelli* are well preserved in the sediments, no living individual was reported so far from Namibia. It is thus likely that their optimal living conditions in the area were in an environment with a more pronounced wet season and perennial lake conditions.

Needless to say, the geomorphological configuration of Logan Island, the sediment composition and their inclusion of these fossils suggest that the island is a remnant of a former pan shoreline and not a residue of an alleged thin layer that once covered the rest of the modern surface of Etosha Pan. As Eugster & Kelts (1983) and Boggs (2001) noted, it is rather a rule than an exception that unlike open lakes, instability of shorelines in closed lakes is a common feature. Such a fluctuation is mainly due to the fact that the water level is maintained by the balance between local precipitation and inflow on one hand and evapotranspiration and infiltration on the other. Even though the climatic regime is a cardinal factor in the fluctuation of shorelines (e.g. Hammer 1986), its occurrence does not necessarily call for a radical shift in the overall environmental conditions. Part of the current shoreline, for example, is experiencing denudation today due to isolated, local events. A sparking case in point is the erosion of the beach cusps, along with the damming of water behind the ridge (Figure 5.13a), which exerts stress on the wall and enhancing the weakening of materials through seepage or throughflow. These processes, particularly the erosion of the beach cusps, lead to the preferential breaching of the shoreline, with erosion spreading laterally along the barrier. Given enough time and continued erosion, traces of the original shoreline may thus remain in places. It is not surprising therefore that the heavy minerals of Logan Island and the lunette dunes are similar; in fact, it would have been baffling if they were not comparable.

Additional support for interpreting the Etosha Pan as a deflational landform was based on the supposedly insignificant thin layer of the pan sediments. Although Buch *et al.* (1992) conceded that the surface morphology of Etosha Pan has the hallmarks of an aggradational feature such as a flat and horizontal surface (Shaw & Thomas 1989), they also maintained that typical pans defined as erosional landforms are characterized by a thin or missing sediment cover.

Thus, based on the findings of Buch (1993) which revealed that the thickness of the pan's sediments does not exceed 150 cm, Buch *et al.* (1992) concluded that such a sediment record does not insinuate a history of a desiccated palaeolake as it was put forward earlier (e.g. Jaeger 1926). The sediment thickness was determined from 12 profiles augered with a hand-operated drill (Buch 1993).

In details, Buch & Rose (1996) classified the pan's sediments into four main zones (Figure 5.14). This classification was based on sedimentological, mineralogical, and geo-chemical properties. The only noteworthy sediment input from the surrounding areas into the pan was found in the southeast (zone A) and the southwest of the Etosha Pan (zone B) and then the Fischer's Pan (zone D). These allochthonous sediments showed vertical differences in terms of calcite and dolomite content. The latter was more apparent at depth, whereas calcite was found to predominate near the surface. The rest of the Etosha Pan, excluding its centre, is characterized by parautochthonous sediments, which made up the third zone (zone C). These materials are thought to have resulted from weathering of parent sedimentary rock of the Andoni Formation. The weathered products were then re-deposited locally through the fluvio-lacustrine processes. This denudation process is believed to have taken place under saline environment, comparable to current climatic conditions (Buch & Rose 1996). Sediments of the centre of the pan were not zoned, due to a lack of data.

Allochthonous sediments differ from those of the underlying parent sedimentary rock in terms of colour, calcium carbonate content and texture. Additionally, considerable mineralogical and chemical differences exist between the imported sediments and the Andoni Formation. The silt and claystone facies of the Andoni Formation are free of calcium carbonate. Since the parautochthonous sediments were locally re-deposited, their properties proved to have been influenced heavily by the underlying parent sedimentary rock. Therefore, their color and calcium carbonate content resemble very closely those of the underlying parent sedimentary rock (Buch & Zöller 1992).

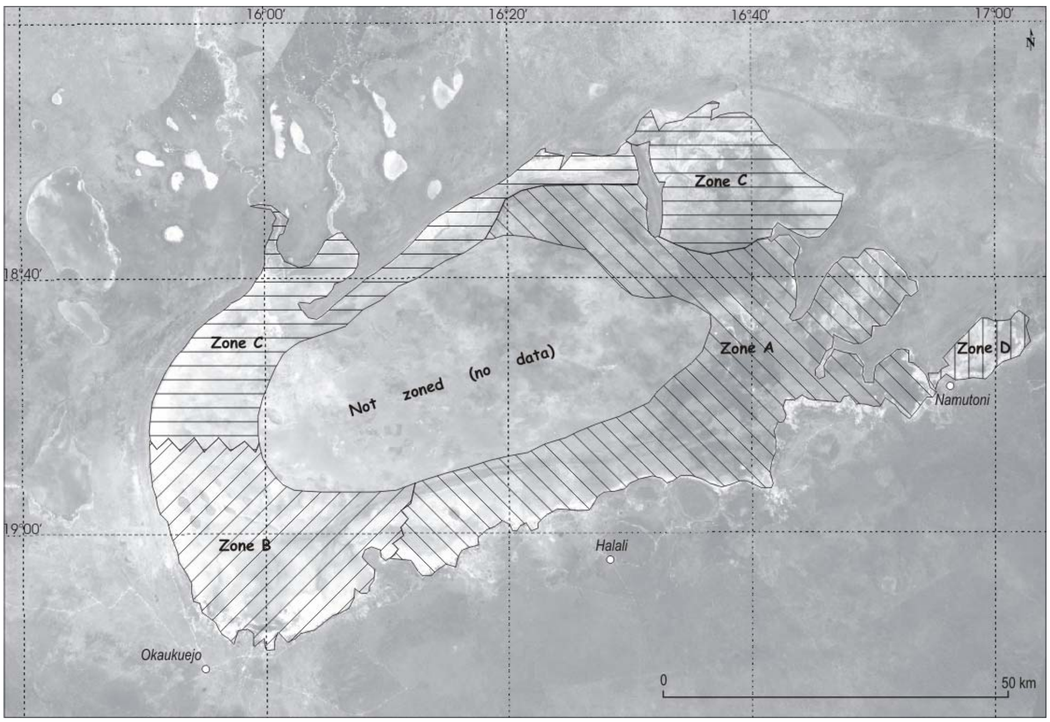


Figure 5.14: Four zones of the loose sediments of the Etosha Pan, according to Buch (1993) and Buch & Rose (1996). Zoning was based on sedimentological, mineralogical and chemical characteristics or properties. Zone C was regarded to consist of par-autochthonous sediments, while the rest of the zones are allochthonous in origin (Redrawn after Buch 1993).

The weathered sediments of the Andoni Formation on the pan floor are associated with the analcime or K-feldspar and mica. This mineral association usually extends to the parautochthonous sediments of the pan. In their absence, sepiolite (loughlinite) is the characteristic mineral. In consulting the literature, Buch & Zöller (1996) endorsed results from a number of studies that associated sepiolite to saline-alkaline conditions in other environments.

The above in-depth demonstration of the absence of appreciable allochthonous lake sediments spoke against the assumption of the Etosha Pan as a palaeolake that was hypothesized to have existed over a long period of younger geological times (e.g. Stuart-Williams 1992). In accordance with minerals of the widespread parautochthonous sediments in the pan and the presence of the underlying Andoni Formation, advocates of Etosha Pan as a super-pan could only recognize a large, shallow, permanent and saline-alkaline lake that they considered to have existed during the Oligocene and Miocene (Buch *et al.* 1992; Buch 1993; Buch & Zöller 1996; Buch 1997).

The alleged lack of lake sediments on the pan floor was pursued in this study. Unfortunately, on the basis of 25 profiles (Appendix B), measurement of the thickness of these sediments was rendered inconclusive due in part to the compaction and heavy nature of the clay. This prevalence of sediment compaction points to the overburden pressure that variously affects the volume of sediments. Friedman *et al.* (1992) observed that the greatest reduction in volume takes place in fine-textured sediments such as silts and clay. As the volume decreases, so does the porosity; simultaneously, the density increases.

Despite the limiting factors mentioned above, eleven of these profiles measured a depth of equal to or in excess of 150 cm. The deepest depth attained without reaching the bedrock was 3.9 m at Eto 217 (Figure 2.5). Here, after digging a soil pit to a depth of 1.1 m, soil augering was further done at its bottom and was abandoned at 2.8 m depth (from the bottom of the pit) without reaching the bedrock. Examples of other exercises with equally inconclusive results were made at Eto 31 (3.7 m), Eto 25 (2.3 m) and Eto 22 (1.95 m). Conversely, a cemented layer was encountered at eight of the pan profiles, at a depth ranging between 70 cm (Eto 27) and 2.4 m (Eto 12).

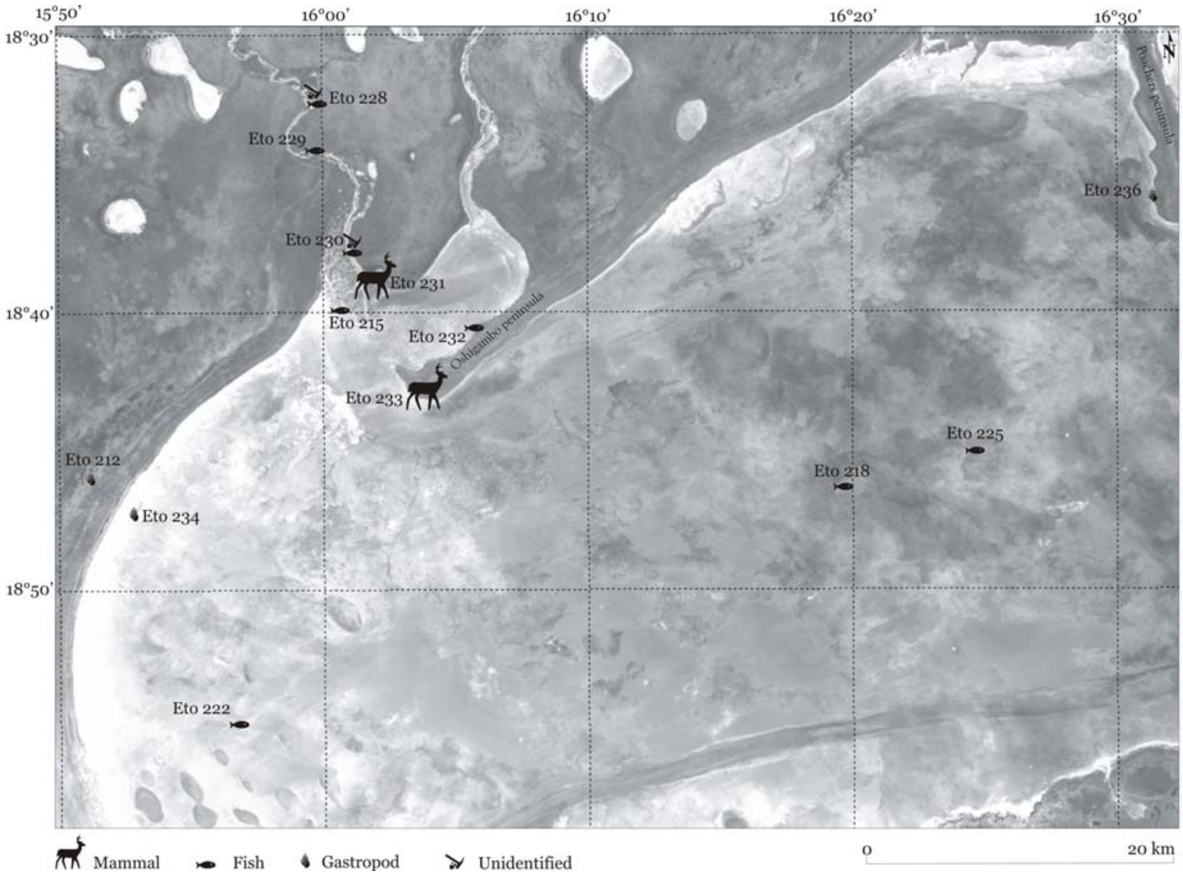


Figure 5.15: Locations of major fossils discovered in and around the Etosha Pan.

Three of the profiles (Eto 218, Eto 222, and Eto 225) from the pan yielded fossils of a freshwater fish (Figure 5.15). These fossil types were also found in the terraces of the Oshigambo Peninsula (Eto 232), terraces of Ekuma River (Eto 228, Eto 229 and Eto 230) and the delta (Eto 215). Fossils found in the pan sediments were buried at a depth of 70 cm at Eto 218, 170-190 cm at Eto 222, and 80 cm at Eto 225.

In spite of their abundance, however, fossil fragments suitable for positive identification to a higher level were rather limited. The best available fragments were a *pectoral spine, dentale* and a *supraoccipital*, all belonging to a *Clariidae species* (Figure 5.17; fragments positively identified by Jousse, H, University of München, 2005). In comparison with modern skulls, these pieces belong to an individual measuring approximately 90 cm. This specific suite of fossils was found at the lower delta of the Ekuma River, at a depth ranging between 70 cm and 90 cm. Figure 5.16 shows the cross section of the soil pit in which these finds were unearthed.

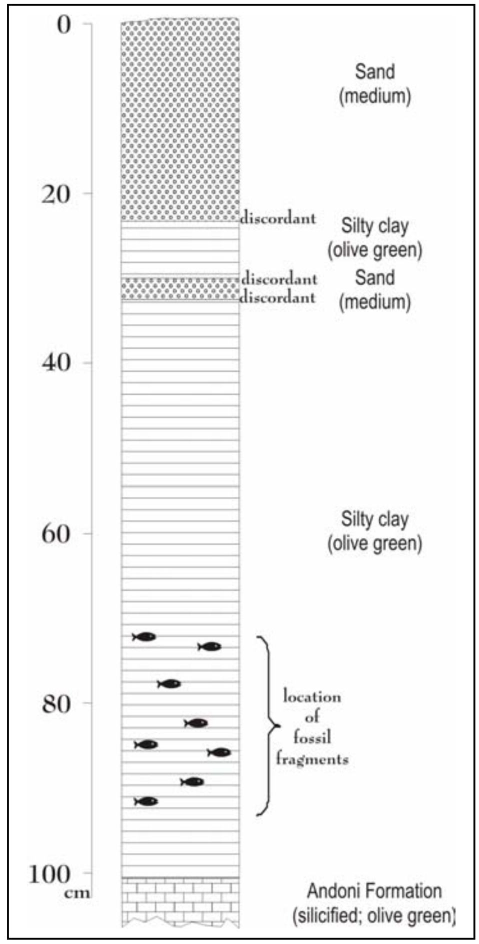


Figure 5.16: Profile of soil pit at Eto 215 and vertical location of fossil fragments belonging to Clariidae *species in the pit.*

The first 20 cm of that pit are characterised by sandy soil, very similar in all respect to that of the nearby L10 (Figure 5.7), located at the delta apex. Below the uppermost layer lies a silty clay film, olive green in colour and only 10 cm thick. This olive green section rests on a very thin sandy layer, measuring 6 cm in thickness, before an olive green silty clay strata grades into the cemented layer of the same color at the bottom of the pit (100 cm). It was in that silty clay layer that the positively identified fish fossils were buried. The fossil fragments were generally scattered in the mud, spanning some 20 cm in thickness and covering most parts of the pit base (80 x 130 cm). There was an unconformity between each of the first three successive layers.

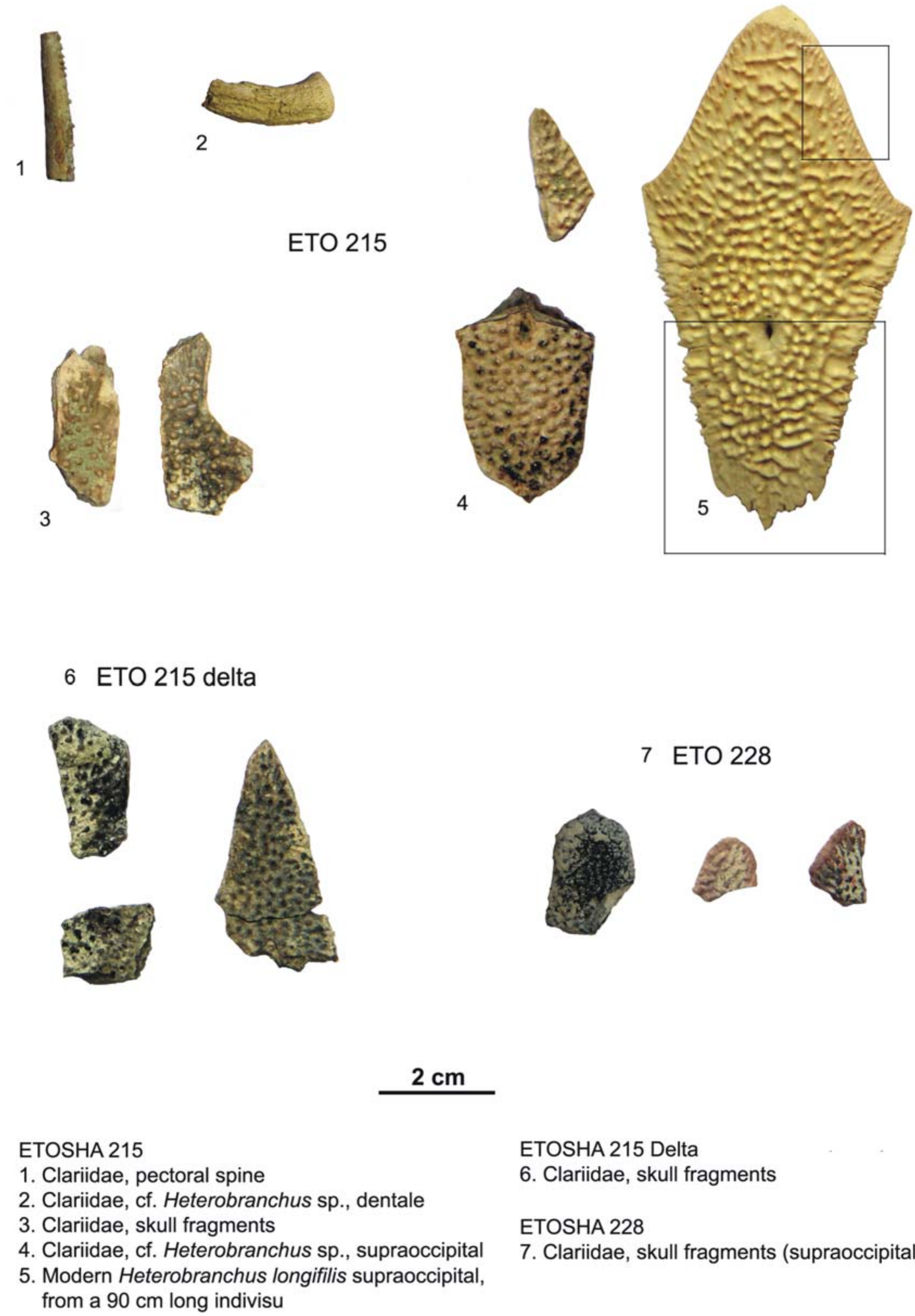


Figure 5.17: Fossil fragments belonging to a Clariidae *species. (Fossils fragments positively identified by Jousse, H., University of München, 2005).*

Figure 5.18: Sample of fossils embedded in sediments of Ekuma River terraces; a) and d) - unidentified; b) and c) - fragments of Clariidae species.

Other fish fragments, found along the river terraces, were embedded in calcrete that makes up the lower terraces. Characteristically, these fragments are sub-angular, measuring up to 4 cm across. Some of the calcrete blocks in the area are composed of fluvial sands, while others consist of finer grains with olive and light green mottles. The bearing of these mottles in a fine muddy matrix in some of these sediments might be a result of remobilization of fluvio-limnic deposits, which is known to be a frequent occurrence in close-basin lakes (e.g. Eugster & Kelts 1983). Terraces that were found to be most prolific with these fossils are at Eto 230, situated 50 to 100 cm above the active river channel. Figure 5.18 presents some of the images of these fossils in their natural environment.

The exact age of these fossils is unknown. Despite their abundance and the relatively large size of the fragments, they proved to contain little or no collagen. This status compromised their quality for dating, hence evading the Carbon-14 method that was to be employed. Reasons for a lack of collagen are many, but the likely contenders in this case are old age and/or leaching by water. However, judging from their close resemblance to modern species, these fossils are likely to be of Pleistocene age.

Essentially, the sheer presence and abundance of these fossils testify to the existence of freshwater life in and around the Etosha Pan. Such large fish species are by far the largest individuals known from mainland Namibia outside the perennial border rivers. Thus this evidence alone points to perennial lake conditions and tributary inflows, which in turn unambiguously dispels arguments for Etosha Pan as a deflational landform as formulated above, as well as the assumption that the previously existing water body was a saline-alkaline lake. Traces of saline-alkaline conditions can only be associated with dried-up phases of the lake.

It is also vital to point out that the assumption of allochthonous sediments limited only to the entire southern section and a substantial portion of the eastern and northern part of the Etosha Pan reflects inconsistency in the interpretation used for zoning the pan sediments. What specifically needs to be raised here is when and why sedimentation exceeded erosion in those parts where allochthonous sediments are zoned, particularly the entire southern half of the pan, while at the same time the rest of the pan continued to experience unrelenting denudational processes. It is also crucial to reflect on the fact that the surface of the pan is flat and according to Buch (1993) the bedrock was universally encountered at a depth of not more than 150 cm across the pan, which ought to illustrate the horizontal layering of the underlying bedrock. The only exception to that depth was the Fisher's Pan, where the depth obtained by the same author was in excess of 2.6 m. Given such a stratigraphical setting, it is therefore unrealistic to have the southern section serving as loci for sediment deposition, whilst the rest of the pan was exclusively and simultaneously undergoing localized denudation and reworking of sediments. Thus loose sediments on the surface of the pan are either allochthonous or parautochthonous; hitherto collective evidence unanimously leans towards the former.

Although the sediment thickness of the Etosha Pan was not conclusively determined, Friedman *et al.* (1992) offered that clues for recognizing lake deposits include a central-lake suite typically composed of fine materials and a marginal suite leaning towards coarse deposits that resulted from deltas, fans, fluvial plains and beaches. Thickness of lake sediments largely depends, among others, on the rate of basin subsidence, which is partly due to sediment loading and the rate of sediment supply. Shallow lacustrine conditions may persist where subsidence is low and the rate of sediment supply is small. In fact, a study of the lacustrine oncoids at the Etosha Pan by Smith & Mason (1991) revealed that the limestone at Poacher's Point which bear the fossils lack detrital nuclei or clastic materials. That led these authors to conclude that the lake during that time (Late Pleistocene, in their estimate) was either not fed by streams or that the channels were of very low gradient and consequently carried very little sediment. Therefore, even if the sediment thickness of the Etosha Pan did not exceed 150 cm as Buch (1993) contended, it does not automatically entitle and relegate the pan to an erosional landform. The declaration of that status to the Etosha Pan was thus premature and was done without taking all other relevant factors into consideration.

To summarize, the hypothesis of Etosha Pan as an erosional landform was built on three main pillars, namely the presence of islands, peninsulas and lunette dunes, provenance and a lack of thick lake sediments on the pan floor. The foundation of this hypothesis was the presence of a thin sediment cover that once blanketed the pan's surface. We have evaluated the merit of most of these assumptions and their premises, and subsequently passed judgment accordingly. Attention will now be turned to the content and arguments for Etosha Pan as a dried-up lake, which is the subject of the following section.

5.3 Etosha Pan as a desiccated palaeolake

The theory of Etosha Pan as a dried-up lake was first documented by Schwarz (1920). His work was precipitated by a grand plan he conceived for redeeming the Kalahari thirstland. In this scheme, he proposed the construction of two weirs, one in the Kunene River, near Calueque and the other in the Chobe River. The goal was to 'restore' the flow of water from the two rivers into the interior as it once was. The water from the Kunene would thus flow along the beds of the Oshana Olushandja/Etaka, push through the Omiramba Owambo and Omatako after flooding the Etosha Pan and then dewater into the Okavango River. In the Okavango swamps, excess water would naturally reach the Makgadikgadi Pans. The flooding of the Makgadikgadi Pans would have forced water to enter the Letiahau River, which would result in water percolating through the sandy country and then into the Nossob and Auob rivers. The net result would have been, in his words, a rise of settlements and agriculture in the Kalahari, which would be "better than those of Egypt". His approach was thus applied in nature, with a grain of utopian motive and paid little details to the origin of the Etosha Pan. Nevertheless, the work of Schwarz (1920) left a template of Etosha Pan as a desiccated lakebed.

Although largely implicit, the basic premise for this group is the continuous existence of an endorheic basin dating to the Tertiary. In other words, the initial thin surface cover that was purported by the camp of the pan as an erosional landform has never been in place on the surface of the Etosha Pan. That stance removed any expectation for pan incision into the underlying layer(s). Thus, instead of net sediment removal from the surface of the pan, they anticipated sediment deposition into the basin. Although they presented no comprehensive details, the ubiquitous olive green clay found in the pan was considered at face value as the lake sediments (Schwarz 1920; Jaeger 1926). The lack of in-depth coverage of these sediments, such as its depth, exposed this group to criticism. It was discussed in the previous section how this lack of conclusive field evidence was exploited by the aeolian camp, which with the benefit of hindsight, has unsuccessfully attempted to demonstrate the absence of allochthonous sediments in the pan.

Other landscape elements that were assessed and evaluated by proponents of Etosha Pan as a desiccated lake were also almost identical to those examined by the antagonists. Central among these elements is the double ridge located along the western margin of Etosha Pan. Jaeger (1926), for example, viewed the so-called lunette dunes in a different light. Instead of dunes, he described five different levels of lacustrine terraces around the pan. These terraces are found at 1 m, 3 m, 12 m (1090 m a.s.l.), and 17-20 m (1,100 m a.s.l.) above the present floor of the pan. With exception of the first two, these terraces are not continuous. They are nevertheless interpreted to have resulted from lowering of the lake's water level (e.g. Jaeger 1926; Stuart-Williams 1992a).

However, Wellington (1938) acknowledged that since aeolian processes have played a role in the evolution of the area, it is likely that these terraces are not entirely fluvial or lacustrine in origin and that they are not necessarily representing former lake levels. His argument was partially based on the findings of Jaeger (1926), who appraised part of the Okondeka 'dune' complex and concluded that it accumulated and encroached towards the pan to the extent of 500 m in 40 years.

Stuart-Williams (1992a, 1992b) variously projected that if the lake existed today, at full capacity it would have covered a surface area greater than 82 000 km^2 or larger than Lake Victoria (69 464 km^2). This would have ranked Etosha Lake as a distant second largest inland water body in the world, behind the Caspian Sea at 370 992 km^2. In her interpretation, approximately 3 million years ago this saline body of water was at least 45 m deep, reaching up to the modern 1,120 m contour line (Figure 5.19a). According to Stuart-Williams (1992a, 1992b), this lake system was altered as recently as the last major wet period approximately 35,000 years ago. As the Kunene drainage assumed its modern form, it affected and removed many of the headwaters of the Etosha Lake system. The 17 - 20 m relict shorelines seen at the eastern side of the Andoni Flats and in the west at Okondeka have thus been generated from recent geological stabilization of the Etosha Lake system at the 1,100 m level. Lower shorelines were subsequently generated when the water level dropped. Interestingly, Stuart-Williams (1992a) believed that during the time when the 1,100 m level developed, Lake Etosha was draining to the sea via the Oshana Etaka and into the Kunene River (Figure 5.19b). No tangible explanation was offered as to how the reverse of present gradient, especially of the Oshana Etaka, might have come into place.

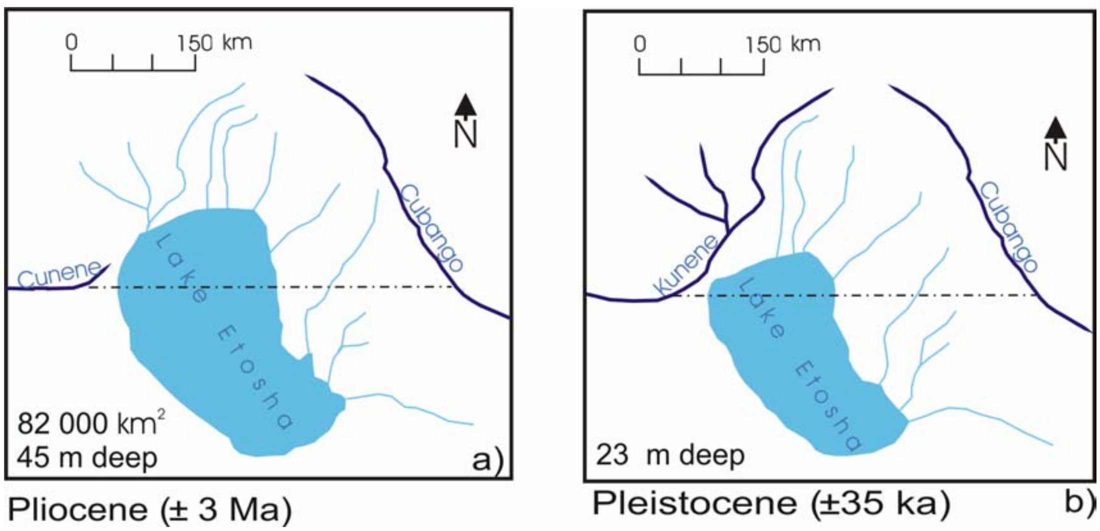


Figure 5.19: The Evolution of the Etosha Pan according to Stuart-Williams (1992a; 1992b). a) During the Pliocene the lake is purported to have been saline; b) during the Pleistocene, it was thought to drain to the sea via the Oshana Etaka and then the Kunene River. (Redrawn from Stuart-Williams 1992a).

The above evidence were considered by among others, Rust (1984, 1985), Buch *et al.* (1992) and Buch (1993), as inadequate to sustain the pronouncement made about Etosha Pan as a desiccated palaeolake. Admittedly, they assembled little field evidence to withstand a technical and substantial body of evidence formulated against their conclusion. In substance and methods, therefore, the dispute was technically in favour of the Etosha Pan as a wind-deflated pan. The fact remains, however, that the role of the Kunene River in the evolution of the Etosha Pan can no longer be denied. The following chapter will attempt to contextualize and synthesise the chronology of the evolution of the Etosha Pan.

CHAPTER 6

The Development and Evolution of Etosha Pan

Introduction

In essence, the subliminal content of the previous chapter might have set the impression that the controversy surrounding the origin of the Etosha Pan was virtually superfluous. On the contrary, that controversy spilled over even to this study. Like in detective investigations, where the product is typically known and the formative processes are the mysteries, solving a puzzle is always marred with innumerable questions, with answers to some destined to remain elusive. Evidence that might have been assembled may also appear to contradict each other. It is thus in that light that this study too, is confronted with a similar dilemma.

Answers to some of the questions concerning Etosha and its environs have emerged, however. Prominent among them is the settling of the debate as to whether wet phases occurred at Etosha, which was ruled out by Buch (1993, 1997) for the last 1.8 Ma. The discussion has thus shifted to when the Kunene River ended its course into the Owambo Basin, particularly in taking into consideration that evidence for a lacustrine environment at Etosha Pan goes beyond the middle Pleistocene and swampy conditions are recorded to have existed as recently as the middle Holocene. It is also important to explain the origin of near-vertical walls that rim the pan at a number of sections in the southern and eastern pan margins, while at the very same time other segments in similar localities consists of loose sand sediments. In other words, why are some of the sections in the eastern and southern margin of the pan characterized by laterally discontinuous calcrete walls, while the rest of the edge is made up of loose sediments?

This chapter is intended to delve into some of these pertinent issues. In so doing, it will also attempt to foster a model for the development and evolution of the Etosha Pan. The wealth of data at our disposal, including those used to interpret the pan as an aeolian landform, will be harnessed accordingly towards that endeavor. Instead of communicating a paradox as it might be perceived in some quarters, the consulting and utilization of all relevant data, regardless of their origin, should rather convey the integrity of and value attached to the amassed data. In other words, opinion is liable to change, unlike reproducible data.

6.1 The timing of the upper Kunene River's deflection to the Atlantic ocean

In section 4.2, we looked at the geomorphic connection between the Etosha Pan and the Kunene River. It was established from field evidence that the upper Kunene had once continued flowing southward at almost the same width as the floodplain upstream of the narrow valley, near Calueque. The Oshana Olushandja/Etaka is considered to occupy the eastern section of that former valley, most of it now filled with sediments. Owing to the fact that traces of this infilled valley and the effects of neo-tectonic in the Oshana Olushandja/Etaka are still discernable, it is probable that the deflection of the Kunene from the Owambo Basin to the sea took place fairly recently.

It is generally believed that some of the current geomorphic characters and configurations of major rivers in the subcontinent, such as the deflection of the upper Zambezi, owe their occurrences to the uplift that took place in the Pliocene. This uplift is considered to have been the most intensive during the Cenozoic in southern Africa (e.g. Bond 1975; Maud & Partridge 1987; Thomas & Shaw 1988; Allanson *et al.* 1990; de Wit *et al.* 2000; Partridge & Scott 2000). Moreover, Brunotte & Spönemann (1997), Kempf (2000), Kempf & Busche (2002) and Sander (2002) recognized in parts of Namibia remnants of saprolithic weathering profiles attributed to ancient intensive chemical weathering estimated to have started in the Upper Cretaceous and lasted until the middle Tertiary. This protracted chemical destabilization might have contributed to the weakening of rocks and made them vulnerable to subsequent denudational processes.

In central Namibia, downcutting of deep canyons of the Kuiseb were numerically dated by means of cosmogenic isotope measurements, which revealed that the incision there started at around 2.81 ± 0.11 Ma ago (Van der Wateren & Dunai 2001). The river-cut surfaces close to the present canyon floor yielded an age of 1.28 ± 0.09 Ma at Oswater and 0.44 ± 0.04 Ma at Gaub Bridge. These dates are regarded by these authors to represent the minimum ages. Although uniform validity for the whole region is not necessarily assumed, it is likely that gorge cutting in the western-flowing rivers, such as the Hoanib and the lower Cunene took place at around the same time. The timing of these events would therefore suggest that there was a tendency towards an increasingly moist climatic regime in the region during that period. For that reason and in combination with the reported neo-tectonics, the deflection of the upper Kunene to the coast is cautiously assigned here to the Late Pliocene, while reliable age-dating is awaited to shed more light. This tentative date is in total agreement with Buch (1997) (albeit he also designated the Miocene/Pliocene age to the uppermost sequence of the Andoni Formation) and falls within the time range of Pliocene and Early Pleistocene that was suggested by Wellington (1938).

Like in previous studies, the controversy surrounding the development of the Etosha Pan starts after the deflection of the Kunene. The most critical period with respect to the Etosha Pan is the time between the Early to Middle Pleistocene. In the available geomorphological and geological data, that period is essentially blank. It is for that reason that diverging theories regarding the pan's development are being formulated to fill that gap in the data. Specifically, the most precarious aspect at Etosha Pan is a regional, calcrete stratigraphy that is believed to underlie the entire Kalahari Basin surface. The presence or absence of this stratigraphy at Etosha has far-reaching ramifications. Its presence would call for the Etosha Pan as an erosional landform, whereas its absence would imply that the basin evolved directly from a smaller lake which, due to the absence of the Kunene's water and sediment contributions, developed successively after the Kunene River was deflected to the coast. The reminder of this chapter will pursue the former scenario.

6.2 Etosha Pan as an erosional Landform

In principle, the scenario for the development of Etosha Pan as an erosional landform presented below follows very closely the model proposed by Rust (1984, 1985) and Buch (1993, 1997). The only major parameter excluded from the earlier model is the role of aeolian action. In this model the eroding and transporting power of sediment from what would later become the surface of the Etosha Pan is shifted wholly to fluvial and littoral actions.

Like Rust (1984, 1985) and Buch (1993) this top-down approach assumes that after the Kunene River started its new course to the South Atlantic in the Late Pliocene, lacustrine conditions over the Owambo Basin surface came to a halt. This came about through the new cycle of relief development that was ushered in through the tectonic uplift and the deteriorating climatic conditions which characterized the Late Pliocene/Early Pleistocene. The abandoned lake experienced a complete desiccation and during that process it left behind a thick layer of calcrete deposited on top of the Andoni Formation. This calcrete layer, called Etosha Limestone by Buch (1993), marked the terminal phase of the fluvial and fluvio-limnic depositional history of the Kunene River. However, this contrasts markedly with popular accounts that regard the Andoni Formation as the uppermost sequence of that depositional history.

Climatic deterioration that followed during the Pleistocene has also facilitated the initial emplacement of the extensive dune systems of the Kalahari (e.g. Heine 1989, 1990; Kempf 2000). In the case of the Owambo Basin surface, this sediment accumulation was particularly laid over the Etosha Limestone, which marked the floor of the former lake and the rest of the basin (e.g. Buch 1993, 1997). The spatial extent of the Etosha Limestone was reconstructed by Buch (1997) to coincide with the modern 1,110 and 1,130 m a.s.l. contour line for the areas lying north and south of the Etosha Pan, respectively (Figure 6.1, 6.3). The same author has also reported that the thickness of the Etosha Limestone reached the 1,110 m a.s.l level (Figure 6.3). At the regional scale, Eitel (1994) came to a similar conclusion by recognizing such a lithostratigraphic sequence which resulted from the deposition of the endorheic drainage systems ending in the Kalahari Basin.

Attempts at attaching an age to the formation of this calcrete generation revealed a complete lack of coherence and correlation at Etosha Pan and indeed across the region. At the regional level, for example, Eitel (1994) derived an Upper Miocene age for what he called the 'Kalahari calcrete'. The basis for assigning that age-range was that climatic shift towards drier conditions in Namibia are connected to the existence of the Benguela upwelling system, which dawned at the end of the Miocene. The very same Kalahari calcretes were previously studied by Watts (1980), who concluded that their formation occurred episodically since the Pliocene. Kempf (2000) working in central and southern Namibia, assigned an age of 600 ka to the terminal formation of the calcrete beneath the Namib Dunes. Such an age derivation was based on fossils of *Elephas recki* and Acheulean artefacts that were found at a site 'Namib IV' situated some 30 km SE of Gobabeb (Shackley 1982). At a manganese mine of Hotazel in the southwestern Kalahari, Bateman *et al.* (2003) applied the OSL dating method and obtained an age of 108 ±24 – 132 ±29 ka from a blocky and massive calcrete, approximately 2 m thick. The sample was taken at a depth of 5.9 m below the surface. Similarly, Bertram (2003) presented an OSL age of 106.1 ±7.9 for a sand ramp layer situated at a depth of 3 m and overlain by a 50 cm layer of calcrete at Aus, southern Namibia.

At Serondela Terraces, near Lake Caprivi, mollusc assemblages contained in calcrete matrixes showed an age of 15,380 ±140 ka, while the calcrete itself registered an age of 11,550 ±140 ka (Shaw & Thomas 1993). These authors are of the opinion that the age variations between the paired materials tend to highlight the difference between the 'real-time' indicator and the subsequent development of the matrix. More recently, Ringrose *et al.* (2005) obtained a TL-derived age of 108,555 ±7,621 ka from calcrete suites of the Makgadikgadi upper shorelines (943-945 m a.s.l.).

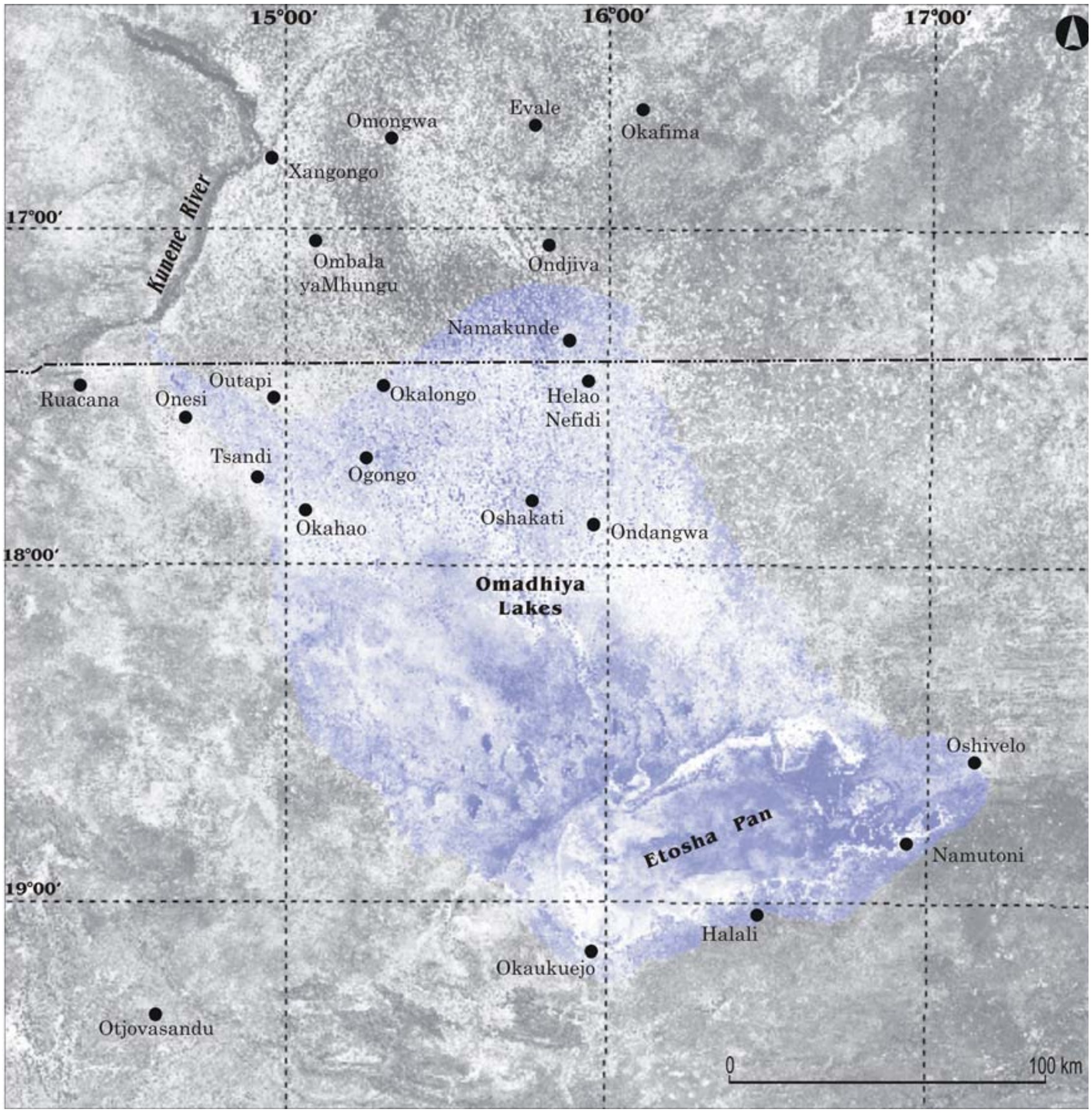


Figure 6.1: Areas (in blue shade) falling within the limit of the 1,110 m a.s.l. contour, which according to Buch (1997) coincides with the spatial extent of the Etosha Limestone sequence (nomenclature after Buch 1993). Etosha Limestone, with a thickness reaching also the 1,110 m a.s.l. in the north of Etosha Pan, is believed to have marked the terminal phase of the depositional history of the Kunene River. Today, natural exposures of calcrete (Etosha Limestone) within that delineated region are limited to pan margins and river flanks, like in many other parts of the Kalahari.

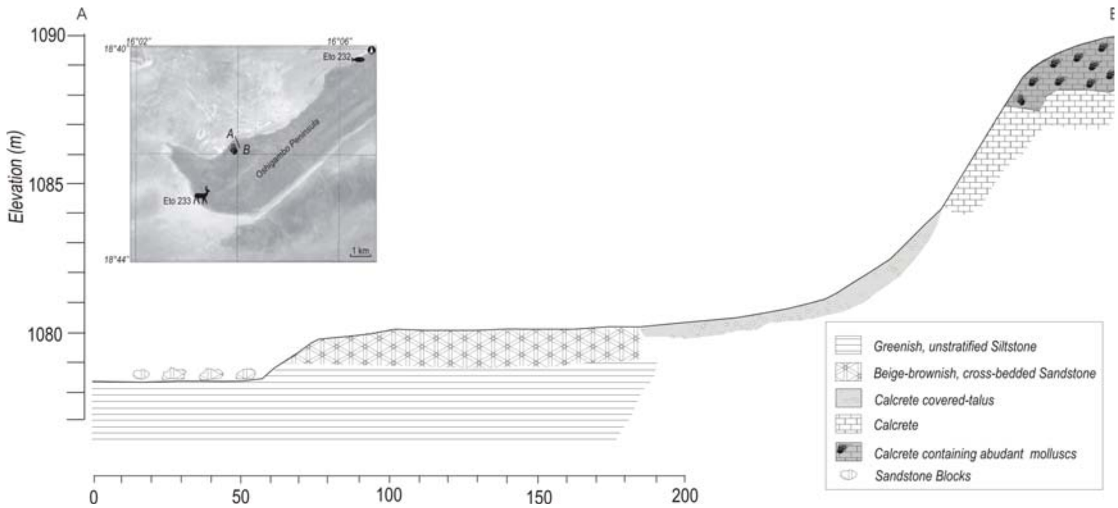


Figure 6.2: Cross-section at the northern flank of Oshigambo Peninsula showing the location of shoreline molluscs embedded in the calcrete (Etosha Limestone). Dating of similar molluscs in the surroundings gave an age of 10,670 ka (Heine 1982), while the host facies is believed to be of Miocene/Pliocene (Buch 1993). The Etosha Limestone is understood to have covered the entire surface before it was eroded to make way for the Pan. Eto 232 (terrace, made up of similar lithology as the sandstone in the transect) and 233 (beach ridge) in the inset yielded, respectively, fragments of freshwater fish and a variety of mammals, including sitatunga (section 5.2). Fish fossils are comparable to modern species, whilst a mammal fragment produced a date of 4.5 - 5.0 ka. From field observations, it appears that the beige-brownish fluvial sediments are younger than the calcrete rim, which implies that they were deposited and then cemented after the cliff had been formed. Except for a 50 cm limestone layer directly overlaying the Andoni Formation at Poacher's Point, calcrete in the area lacks a (horizontal) bedding signature. (Redrawn after Buch 1993).

Almost identical to Eitel (1994), Buch (1993, 1997) and Buch & Trippner (1997) assigned the Miocene/Pliocene age to the formation of the Etosha Limestone at Etosha. They also pointed out that the oldest layer was deposited simultaneously with the Andoni Formation. Smith & Mason (1991), while studying the lacustrine oncoids at the Poacher's Point exposure, inferred an Early to Middle Pleistocene age for the stromatolite-bearing succession which they called Poacher's Point Formation limestone. Rust (1984) reported radiocarbon dates ranging from 42,400 to 39,300 ka and 32,200 to 30,300 ka for the said oncoids and calcrete, respectively. Smith & Mason (1991), for example, have already reserved their judgment with respect to these dates and opted to consider them as minimum ages for that formation due to the ambiguous C14-age determination from calcrete suites and the resultant age's proximity to the 40 ka limits.

Interestingly, Buch (1993, 1997) constructed a number of cross-sections at several locations at Etosha, including one each at the Oshigambo (Figure 6.2) and Poacher's Point Peninsula where the uppermost section of the so-called Etosha Limestone contains abundant molluscs embedded in the sequence. These molluscs are pervasive in the shorelines of Etosha Pan; they are also found in loose and semi-consolidated sediments (section 5.2). The most common species in the area is the *Xerocerastus burchelli*. Samples of these gastropods were obtained from the western margin of the pan and dated by Heine (1982). They yielded an age of 10,670 ±465 ka. Moreover, lower terraces cut into the Etosha Limestone at the Ekuma River host fossils of *Clariidae* species, a freshwater fish believed to be of recent age. As detailed in section 5.2, fragments of these fossils are comparable to modern species measuring approximately 90 cm.

Essentially, the inclusion of the molluscs in these calcrete beds and the resulted age appear to jeopardize the assumption of the pre-existence of an extensive limestone sequence older than the Middle Pleistocene over the former surface of Etosha Pan as it was to be endorsed and incorporated in this model of the pan as an erosional landform. All results from absolute dating of calcrete in the region and at Etosha showed consistently an age younger than the oldest recorded date (140 ka) of the Okondeka shorelines (Buch & Zöller 1992; Buch *et al.* 1992). This suggests that the presumed 'remnants' of the so-called Etosha Limestone along the shorelines of Etosha Pan are most likely nothing else than groundwater calcrete formed in the littoral zone. In fact, the occurrence of calcrete rims, particularly along valleys flanks, lakes and pans margins are a common phenomenon in the Kalahari (e.g. Lancaster 1979; Nash *et al.* 1994a, 1994b, 1994c; McCarthy & Ellery 1995; Nash 1997; Ringrose *et al.* 1999, 2002; Mees 2002; Nash & McLaren 2003). Initially, the calcrete exposure in this setting was used to construct a lithostratigraphic sequence in the region, based on the assumption that the relict rivers of the Kalahari, in particular, have originated purely from channel incision into a pre-existing calcrete sequence (e.g. Grove). Recent studies in the region and beyond tend to point to the role of groundwater and structural control as some of the main agents at work in shaping the morphology of these valleys (e.g. Bowler 1986; Laity & Malin 1985; Osterkamp & Wood 1987; Shaw & de Vries 1988; Nash *et al.* 1994a, 1994b; Nash 1995, 1997). It is therefore likely that if the pre-existing Etosha Limestone has indeed occurred within the realm of the Etosha Pan boundaries, then there may be an alternative explanation for the absence of calcrete rims at most of the pan's segments. The only logical explanation is fairly remote, which is that the pre-existing Etosha Limestone must have been subjected to considerable alteration during the development phase(s) of the pan, then must have gone into solution and have disappeared as a result. In spite of this major setback in pursuing the model of Etosha Pan as an erosional landform, the age of this alleged stratum is arbitrarily adjusted to the Middle Pleistocene.

The volume of sediment removed from the surface of the basin to make way for the Etosha Pan is best appreciated from Figure 6.3, where a cross-section of the Pan from Adamax Pan to the shoreline of the Andoni flats, via Poacher's Point Peninsula is presented. As stated earlier, at the terminal phase of Lake Kunene the surface of Etosha Limestone was presumed to follow the 1,110 m a.s.l. level in the areas lying north of the Pan, while in the south the 1,130 m a.s.l. level was proposed (Buch 1997). Along the cross-section in Figure 6.3(ii), Etosha Limestone is exposed at the eastern rim of Adamax Pan and the flanks of Poacher's Point. The calcrete layer shown in Figure 6.3(i) between the pan and Ombika belongs to the pediment zone that intermediates between the Otavi Dolomite Hills and the plains. The presence and development of calcrete in this area is largely attributed to groundwater by Miller (2001) (section 4.4, Figure 4.20).

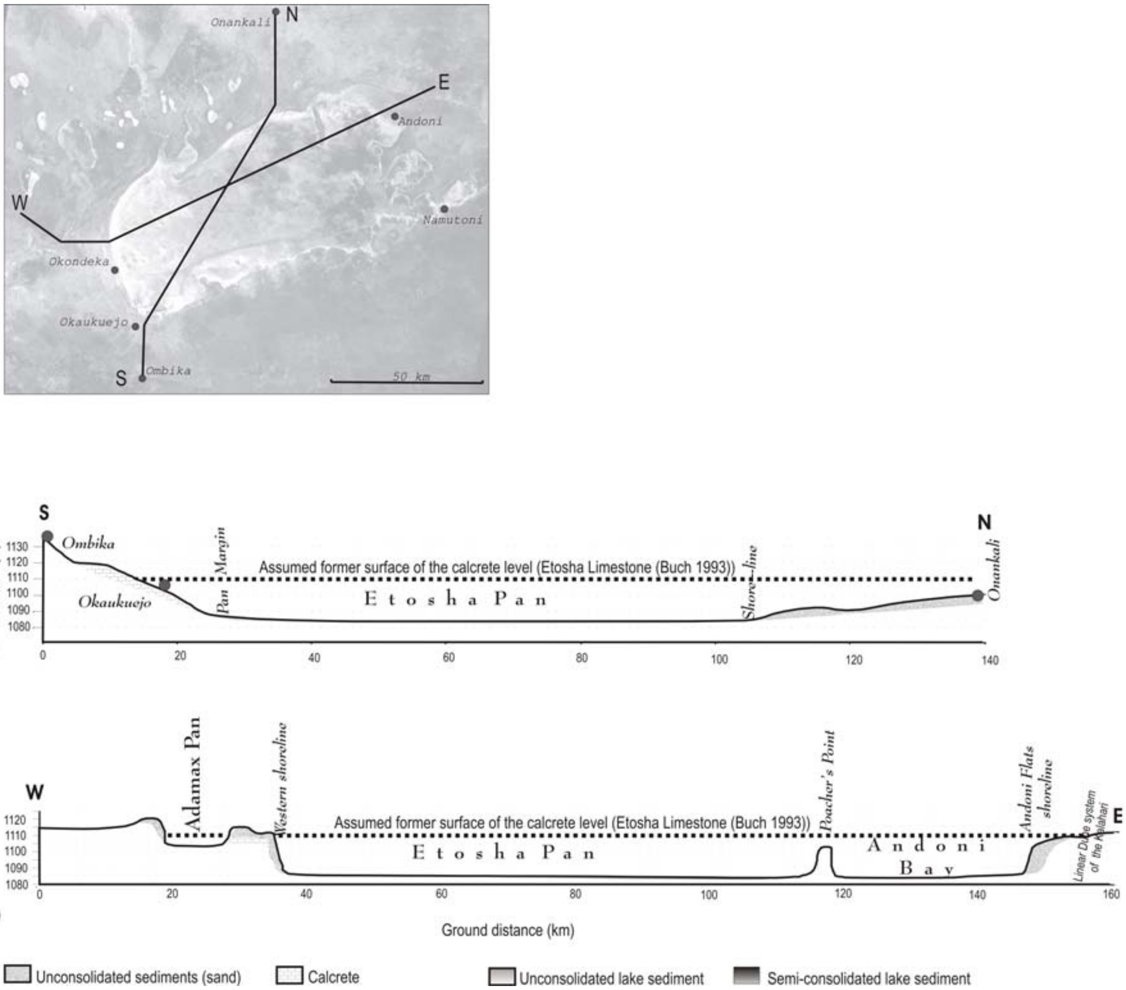


Figure 6.3: Cross-section of the Etosha Pan illustrating the presumed volume of sediments removed from the surface of the Etosha Pan, the bulk of it belonging to the Etosha Limestone. i) Ombika to Onankali and ii) Adamax Pan to the shoreline at Andoni flats, via Poacher's Point Peninsula. Poacher's Point is regarded as a remnant of the Etosha Limestone. Note that there is a lack of lateral continuity between calcrete exposures in opposing flanks. The uppermost section of the limestone at Poacher's Point contains shoreline molluscs; a sample of similar molluscs was obtained from elsewhere in the surroundings and yielded an age of approximately 10,670 ka when dated (Heine 1982). The thickness of unconsolidated lake sediment is unknown. A maximum depth of 3.9 m was attained without reaching the bedrock (section 5.2). Note also the subdued relative relief towards Onankali and a lack of sharp, erosional margins from both Okaukuejo and Onankali. Vertical exaggeration – 1:280. (Data source NASA)

As mentioned in section 5.1, the shoreline at Andoni flats is composed of loose Kalahari sediments. The area east of that shoreline belongs to an extensive linear dune system of the middle Kalahari, in which the shoreline at Andoni flats grades. Such a spatial arrangement stimulates suspicion as to why the Etosha Limestone is missing in that area and in the north towards Onankali, while the dune systems are presumed to have been incised up to 30 m below their base level, and only then had the denudational processes continued further into the Etosha Limestone (Buch 1997). Despite the reservation expressed above, the figure illustrates an enormous volume of sediments that was removed to make way for the Etosha Pan.

Owing to the discounted aeolian deflation that was suggested by previous researchers, fluvial action is implicated here as the principal agent responsible for the removal of sediments from the surface of the Pan. Predisposing conditions for fluvial action to take root are the presence of a deeper base level and an outlet to help whisking away the eroded materials. In this case, it is presumed that the effective base level resulted from the work of neo-tectonics or by virtue of epeirogenetic movements as Rust (1984, 1985) suggested.

A considerable volume of water is also essential for the denudational processes to be effective. This would imply that either perennial flow or low-frequency high-magnitude flood events were observed in the area. Since the Cuvelai was the only significant fluvial system that remained in the catchment area of Etosha after the Kunene was deflected, it is imperative that the inflow over what soon was to be created as the Etosha basin was supplied from that system. Thus, the Cuvelai System flowed southward like today and poured its collective load into a transient shallow lake Etosha. To prevent the lake basin from filling up and to enable the erosion of the Etosha Limestone in particular to continue, outlets in the northeast are thus suggested. This is because the eastern section of the pan is slightly deeper than the western side. The Omiramba Owambo and Omuthiya, although they flow into the pan today, are considered as the likely outlets. Omuramba Owambo would connect with Omuramba Omatako as suggested earlier by Schwarz (1920) (section 4.4) and then feed into the Okavango River. However, as discussed in section 4.4, Omuramba Owambo has the second steepest longitudinal profile (averaging 1:1795 and four times steeper than Oshana Etaka) in the current major channels that feed into or towards the Pan. Alternatively, Omuramba Omuthiya could have communicated with the Okavango River via the Omuramba Namungundo.

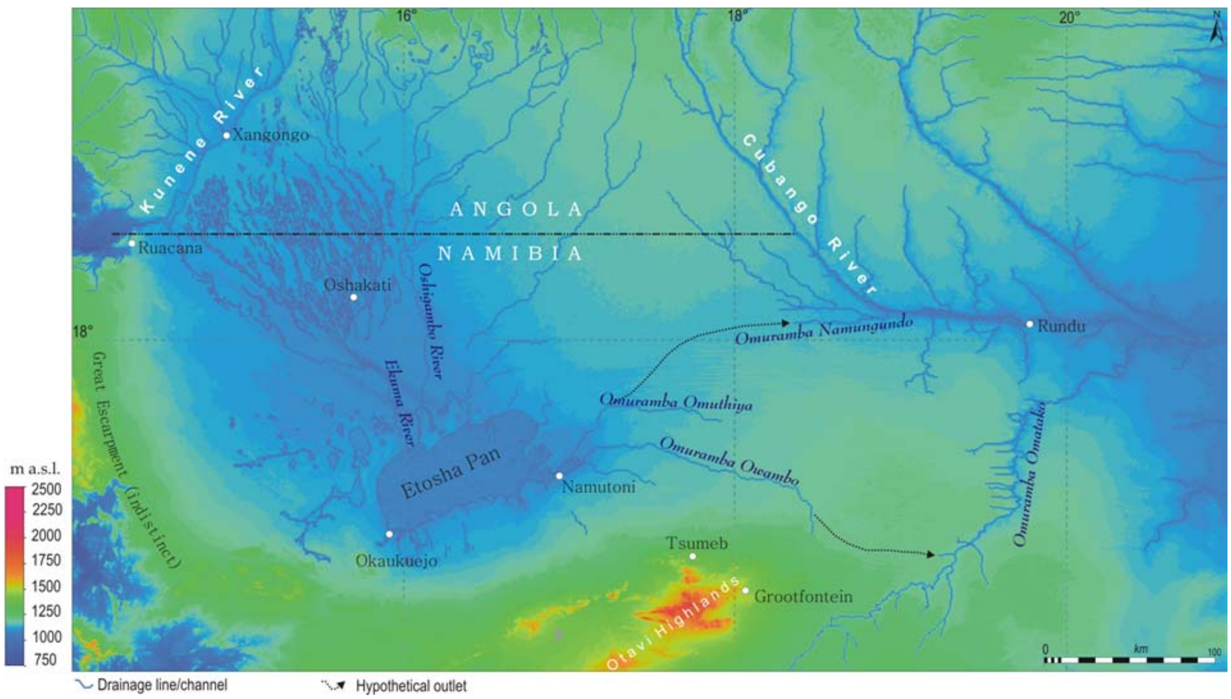


Figure 6.4: Deciphering of possible outlet(s) from the Owambo Basin to help transport materials from the surface of Etosha and create the basin in the process. Currently, both the Omuramba Owambo and Omuthiya feed into the Etosha Pan. Omuramba Owambo is comparatively larger than Omuthiya and both channels drain from an area mantled with an extensive linear dune system (section 4.4). The distance represented by the dotted arrow situated between Omuramba Omuthiya and Namungundo is approximately 140 km, while for the one between Omuramba Owambo and Omatako is 80 km. Likewise, the difference in elevation between Fisher's Pan, into which the Omuthiya and Owambo feed, and the respective water divide of their catchment from those of the Okavango system is approximately 80 m and 150 m, respectively. (Source of topographic data: NASA)

The exorheic setting of Lake Etosha must have to come to an end, however, to characterize the current endorheic situation. Again, the only possible explanation is neo-tectonics that could have affected and reversed this exorheic configuration at a regional scale. However, it is also possible that neo-tectonics might have remained as active before the shift to endorheic conditions, except that for a long time fluvial erosion could compensate uplift and thus continue exorheic drainage. Additionally, when the climate shifted to drier conditions, less water meant less erosion and a change of the balance to neo-tectonics thus become more dominant in this contest. Similar conditions were also reported in central Iranian Uplands, where internal drainage is known to have developed after the Tertiary (Busche *et al.* 2002). The shift of drainage at Etosha would then have translated into the net sediment balance in the pan, from degradation to aggradation. Lake sediments that are found in the Etosha Pan are thus a product of the aggradational process that followed.

The timing of when the system changed from being predominantly one of degradation can only be speculated on. Buch (1997), for example, suggested that incision reached the boundary surface between the hanging Etosha Limestone and the underlying Andoni Formation at around 20 ka. Since the Andoni Formation is only exposed at two localities (at Pelican Island and Poacher's Point) in the Pan, it is likely that further incision into that layer was prevented as a result of system change in favor of endorheic conditions. From there on, the evolution of the pan was largely governed by climatic forcing, and witnessed episodes of high and low lake levels (discussed in section 6.3). The deterioration of climate had subsequently led Lake Etosha to dry-up completely which ultimately reached the current situation.

Against the background of the events highlighted above, distinct morphological processes have also been taking place within the basin. Amongst the processes and results worth mentioning here is the curved coastline, beach alignment, and the sorting of beach ridges of the northwestern and western Etosha Pan. The characteristics of these features are considered to have been largely determined by the waves generated by a dominant northeasterly wind, which in Buch's (1993) analysis is thought to have been in place since at least the last 140 ka (Figure 6.5). Furthermore, the elliptical form and the long axis of the Etosha Pan, the latter being strikingly oriented in the ENE-WSW (Figure 5.1), are also in general agreement with the northeasterly wind.[i] The assumed cycles of the monsoonal system during the Pleistocene, however, tend to denote that such an alternating shift would also have an impact on the prevailing dominant wind from northeast to southeast, as Stuut *et al.* (2002) remarked for such a change on the basis of sediments obtained from the Walvis Ridge. From that perspective, the shift to and dominance of the southeasterly wind is not reflected in the configuration of the pan.

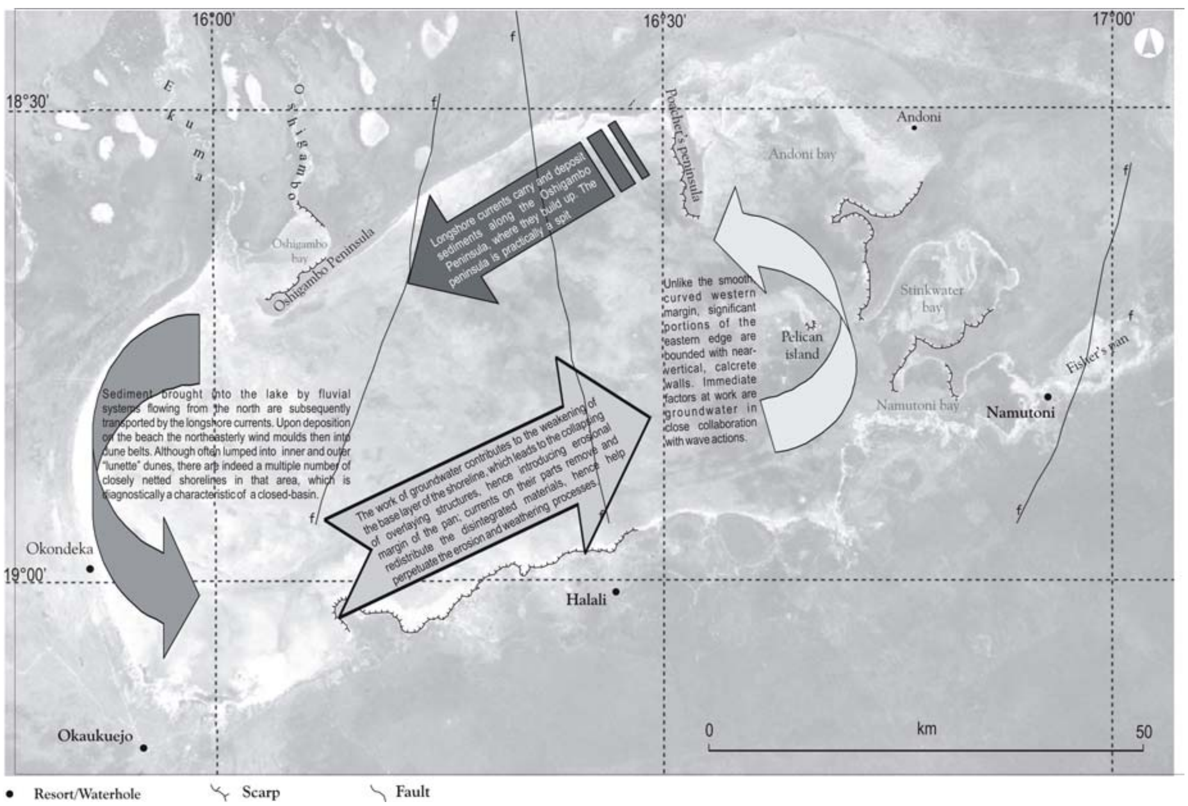


Figure 6.5: Inferred currents of Lake Etosha under the influence of the prevailing northeasterly wind. The long axis of the pan is also thought to owe its orientation to that wind vector.

To recap, the plausibility of Etosha Pan as an erosional landform as reconstructed above rests on whether or not there existed a calcrete stratigraphy over the pan's surface. The available data send a confounded signal in substantiating the pre-existence of such an alleged formation, partly because the

[i] Until recently, wind data at Okaukuejo, the only first weather station in the Park, were collected from 8 compass sectors.

occurrence of calcrete walls is laterally discontinuous along the pan margin and that its patchy occurrence is almost exclusively restricted to the southern and eastern segments, where carbonate-rich pan sediments are also found. Such a spatial distribution gives a hallmark of groundwater calcrete formed opportunistically in the littoral zone. An additional mixed-signal in the data are the shoreline molluscs that are contained in the uppermost layer of the Etosha Limestone, whilst fragments of a large freshwater fish comparable to modern species have been found in the lower terraces of the Ekuma River and Oshigambo Peninsula and then in the pan sediments (section 5.2). This is compounded by the fact that, for the molluscs, similar species from elsewhere in the surrounding that were subjected to absolute dating, yielded a late Pleistocene age. Thus an unambiguous ascertaining of the pre-existence of the Etosha Limestone post-dating at least the middle Pleistocene is of utmost importance for a sound reconstruction of the development of the Etosha Pan.

The absolute age determination of the so-called Etosha Limestone is in progress, where the molluscs found at the flanks of Poacher's Point (Eto 236) and those unearthed from the outer shoreline of the northwestern pan margin (Eto 212), and then the sediments containing fragments of a freshwater fish obtained from the lower terraces of the Ekuma River (Eto 230) are being employed. Whilst those results are awaited to shed more light, it is certain, however, that the role of wind that was advocated by previous researchers as being the primary geomorphological agent responsible for the development of the Etosha Pan is a complete fallacy. The next section will thus explore the palaeoclimate of the area, which, contrary to Buch (1993, 1997), intercepted windows of humid phases during the last 140 ka at Etosha Pan.

6.3 Palaeoclimate of the study area since the Late Quaternary

The palaeoclimate of the Etosha region that is presented here is vouchsafed from dates generated by a number of studies carried out in the area. At present there are at least 30 quantitatively dated sediments and fossils from within the periphery of Etosha Pan (Table 6.1). However, the bulk of this information was derived mainly from calcretes, the material renowned for yielding ambiguous dates and for being notoriously a poor indicator for palaeoenvironment in which they were formed (e.g. Kempf 2000). Shaw & Thomas (1993), for example, paired the ages obtained from calcrete matrices against the dating of shoreline mollusc assemblages contained in that facies. In so doing, these authors concluded that calcrete was generally some 15 - 25 % younger over the late glacial range. However, it is also likely that the age variation between the paired materials might be correctly reflecting the difference between the 'real-time' indicator and the subsequent development of the matrix (Shaw & Thomas 1993). Conversely, Candy *et al.* (2004) employed a U-series disequilibria dating method to derive chronological information of calcrete sampled from alluvial terraces in Spain. Their results have shown that 'mature' calcrete may take 121 to 61 ka to form. These examples serve to illustrate that for the time being the usage of the dates obtained from calcrete should be done provisionally and with utmost care.

Analysis of the available absolute dated materials derived from the Etosha Pan and its immediate surroundings suggests that there were at least seven pluvial phases in the area during the last 140 ka years (Figure 6.6). The interception of high water levels is obviously skewed as they were derived largely from radiocarbon measurements. Thus, most of the assumed wetter phases fall within the last 40 ka, in which the radiocarbon dating method is applicable for suitable materials.

Inferred episodes of high lake levels are comparable with those obtained from neighboring Makgadikgadi Pans (Figure 6.6). Corresponding lake phases are particularly discernable at 33 - 28 ka, 15 - 12 ka and at about 5 ka. This suggests that similar climatic conditions were in force almost across the region. Conversely, notable contrasts between dune sands mobilization and assumed high lake levels have also emerged particularly at 19 - 17 ka. At the same time, marked overlaps between dune accumulation and high lake levels at the Etosha Pan are apparent at 34 - 28 ka, 17 - 15 ka as well as at around 5 ka.

The overlap that punctuated the period around 5 ka is particularly interesting because of the unequivocal evidence for perennial conditions at Etosha. Fossils of an 'amphibious' antelope, sitatunga that this study presented were dated to that period. Enhanced lacustrine activities during that period is also corroborated by the shoreline accumulation in the western side of Etosha Pan, as re-interpreted from the studies of Buch & Zöller (1992) and Buch *et al.* (1992). Moreover, the environmental chronology of northwestern Namibia over the last 22 ka compiled by Eitel *et al.* (2002) reflected a lone 'more humid' phase during the time under review, which is overlapping with the recorded period of when sitatunga roamed at Etosha Pan. In contrast, Brook *et al.* (1999) dated submerged cave speleothems from a nearby Aikab Cenote and derived an equivalent radiocarbon age of 6.4 ka, but argued against evidence for a significant increase in moisture during the middle Holocene from about 6 ka to as late as 1 ka BP. It is thus likely that if the data resolution reflects the true complexity of local records, then these overlaps suggest that local precipitation at Etosha was subdued, whilst the lake levels (most likely low) were sustained by inflow from the Angolan Highlands. Alternatively the obtained overlaps between wet and dry evidence are merely due to a lack of resolution in the current data set.

Table 6.1: Age of absolute dated materials from Etosha Pan and the surroundings

Age	Dated material	General Location	Depth	Author's interpretation	Dating method	Source
2400 ±500	Sand	Okondeka (outer dune)	20 cm	Dry*	TL	(Buch & Zöller 1992; Buch *et al.* 1992).
3510 ±120	Evaporitic (powder) calcrete	Andoni Bay (barchan)	Unknown	Dry	C-14	(Rust 1984)
4750 ±80	Mammal fossil (sitatunga)	Oshigambo Peninsula	50 cm	Humid	C-14	This study
5600 ±2200	Sand	Okondeka (inner dune)	120 cm	Dry*	TL	(Buch & Zöller 1992; Buch *et al.* 1992).
9310 ±90	Sinter	Namutoni	Unknown	Humid	C-14	(Rust 1984; Rust et al. 1984)
10000 ±2400	Sand	Okondeka (outer dune)	95 cm	Dry*	TL	(Buch & Zöller 1992; Buch *et al.* 1992).
10400 ±90	Calcrete	Chudop	unknown	Morphodynamic stability	C-14	(Rust *et al.* 1984)
10670 ±465	Molluscs (Xerocerastus sp.)	Western margin	Unknown	Pluvial	*C-14*	(Heine, 1982)
11900 ±120	Lacustrine chalk	Kapupuhedi	surface	Pluvial	C-14	(Rust 1984; Rust et al. 1984)
12720 ±165	Lacustrine chalk	Unknown	Unknown	Pluvial	*C-14*	(Heine, 1982)
13000 ±2700	Sand	Okondeka (inner dune)	390 cm	Dry*	TL	(Buch & Zöller 1992; Buch *et al.* 1992).
13680 ±175	Lacustrine chalk	Unknown	Unknown	Pluvial	*C-14*	(Heine, 1979; 1982)
14800 ±2800	Sand	Okondeka (inner dune)	250 cm	Dry*	TL	(Buch & Zöller 1992; Buch *et al.* 1992).
18100 ±190	Calcrete	Kapupuhedi	unknown	Morphodynamic stability	C-14	(Rust et al. 1984)
19700 ±4100	Sand	Okondeka (inner dune, base)	>400 cm	Dry*	TL	(Buch & Zöller 1992; Buch *et al.* 1992).
19800 ±180	Pedogenic calcrete	Pelican Island	Unknown	Humid	C-14	(Rust 1984)
21400 ±230	Lacustrine chalk	Pan surface, Ondongab	surface	Pluvial	C-14	(Rust 1984; Rust et al. 1984)
22250 ± 330	Mollusc-rich tufa	Southern margin (near Homob)	Unknown	Higher water-level	C-14	(Heine, 1982)
28200 ±890	Pedogenic calcrete	Oshigambo	Unknown	Humid	C-14	(Rust 1984)
28700 ±550	Pedogenic calcrete	Oshigambo	Unknown	Humid	C-14	(Rust 1984)
30300 ±720	Pedogenic calcrete	Pelican Island	Unknown	Humid	C-14	(Rust 1984)
31000 ±710	Pedogenic calcrete	Ondongab	Unknown	Humid	C-14	(Rust 1984)
32200 ±770	Pedogenic calcrete	Pelican Island	Unknown	Humid	C-14	(Rust 1984)
33900 ±730	Pedogenic calcrete	Ondongab	Unknown	Humid	C-14	(Rust 1984)
37900 ±100	Pedogenic calcrete	Logan's Island	80 cm		C-14	(Rust 1984)
39300 ±1470	Stromatolite	Pelican Island	Unknown		C-14	(Rust 1984)
>40000	Stromatolite	Pelican Island	Unknown		C-14	(Rust 1984)
>41000	Stromatolite	Poacher's Point	Unknown		C-14	(Rust 1984)
>42000	Stromatolite	Poacher's Point	Unknown		C-14	(Rust 1984)
42400 ±1950	Stromatolite	Pelican Island	Unknown		C-14	(Rust 1984)
46300 +4060/-2680	Evaporitic calcrete	Oshigambo	Unknown		C-14	(Rust 1984)
70400 ±1620	Okondeka III soil	Okondeka Dune	450 cm	Dry*	TL	(Buch & Zöller 1992; Buch *et al.* 1992).
140000 +312/-45 ka *(sic)*	Sand	Okondeka (outer dune, base)	>500 cm	Dry*	TL	(Buch & Zöller 1992; Buch *et al.* 1992).

* Data re-interpreted in this study to high lake levels, on account that dated samples were obtained from shorelines and not 'lunette dunes' as previously thought by original author(s).

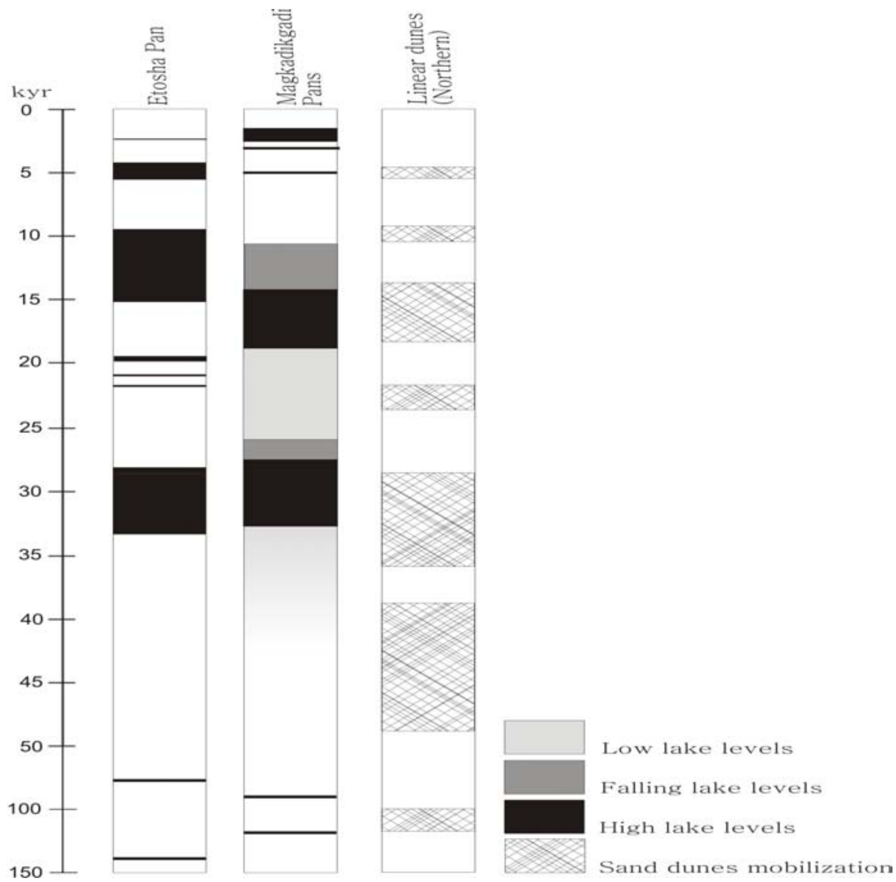


Figure 6.6: Time-slices of palaeoclimatic episodes for the Etosha Pan as compared with neighboring Makgadikgadi Pans and the phases of major linear dunes construction in the northern part of the middle Kalahari (part of the dune system covering the area drained by the Omuramba Owambo) within the last 150 ka. Note the break in scale at 50 ka, and also that qualitative ranking of lake levels at Etosha Pan utilized only one class. (Sources: Heine 1979, 1982; Cooke & Verstappen 1984; Rust 1984; Rust et al. *1984; Buch & Zöller 1992; Buch* et al. *1992; Shaw* et al. *1997; Stokes* et al. *1997b, 1998; Thomas* et al. *2000; Thomas & Shaw 2002; Ringrose* et al. *2005; this study).*

The timing regarding the terminal desiccation of Lake Etosha and the subsequent attainment of its modern status is indetermined. However, the youngest absolute date testifying to a humid phase was approximately 2,400 years before the present (re-interpreted from Buch & Zöller 1992; Buch *et al.* 1992). Although sample material that yielded the youngest date was taken near the surface, it was spatially obtained from the outer shoreline. Hence it may not reliably capture an insightful date with respect to the time when the lake was effectively transformed into a playa. Nonetheless it is probable that after 2,400 years before the present, subsequent fluctuations have been of relatively low amplitude and duration (e.g. Tyson 1986; Cockroft *et al.* 1987; Cohen & Tyson 1995) as the pulse of environmental conditions similar to today was shifting into place.

The historical record on the other hand is rather patchy. However, some early missionary travelers such as H. Hahn (from his diary quoted in Kempf 2000) and individual members of the Hei//kum community, C.J. Andersson and the game wardens at Etosha reported to Schwarz (1920) at least semi-perennial conditions for the Lake Onandova (now Fisher's Pan) during the middle of the 19^{th} century. In the same vein, the prehistoric record is presently not fully understood, although there are many Later Stone Age and Middle Stone Age sites surrounding the pan and on the islands (e.g. Pelican Island), suggesting that at times this was a favorable habitat for hunting man.

Kempf (2000) derived a preliminary relative chronology for the Namibian Pleistocene and Upper Tertiary from geomorphological analysis and proxy data of both the terrestrial and shelf regions. The succession of relief generations in Etosha and adjoining north-central Namibia does fit into this proposed geo-chronology (Table 6.2) on the supposition that soil formation on nearby sand deposits reflects more or less perennial lake phases in the basin. The extension of those lake phases is varying throughout the Pleistocene. Especially on the present lowest level there was obviously a smaller lake with beach ridge formation represented in Logan's Island and some foreshore islands of the northern rim, followed by a more extensive lake phase that eroded much of Logan's beach deposits.

Table 6.2: Relative chronology for the Namibian Pleistocene and upper Tertiary from geomorphological analysis and proxy data of both the terrestrial and shelf regions as modified from Kempf (2000).

Main Generations	**Processes in Central Namibia**	**Morphological Indicators**	**Cultural / Fossil Documents**	**Relief Genera-tions in Etosha**	**Sea Level**	**Timing after Kempf (2000)**	**Possible Corelation**
30	Holocene relief generations	Gravels, kollu-vial sediments	Namibian Later Stone Age	Gully formation in lower beach ridge sediments			Holocene
29	Minor gravel re-activation	Young gravels	LSA, transported MSA	Lake phase, Okondeka-I soil			
28	Soil formation, minor incision	Terrace formation					
27	Silty-clayey depo-sits (river end dep./ slackwater dep. ?	Fine terraces, colluvial bodies	MSA	Okondeka-II soil	Post-Walvis Regressions	**Ugab-Era**	Gamblian/ Wisconsin (?)
26	Aeolian sand sedimentation	Dunes, gypcrete and desert pave-ment with rock coatings	HIATUS	Longitudinal dune reactivation in north-central Namibia			
25	Week minor incision	Valleys, terraces	Early Middle Stone Age (MSA)				
24	Swamp sediments, calcareous tufa at stillwater reaches	Tufa with root casts and fossil leaves	„Mousterian" „Sangoan" (?)	Lake phases Okondeka-III soil	Walvis-Transgression		Monastir/ Eemian (?)
23	Aeolian sand sedimentation	Sand and dune bodies	HIATUS (?)	Longitudinal dune formation	Post-Vineta Regression	**Tsondab-Era**	Kanjeran/ Saale
22	Minor gravel deposition	Loose gravel layers		Lake phase	Vineta-Transgression		Tyrrhen/ Holstein
21	2^{nd} phase of deep valley incision	Valleys, Terraces	„Acheulian Complex"				
20	Calcification	„Middle Cal-crete Surface"		Calcrete, calcified talus	Post-Rooikop Regression	**Swakop-Era**	Kamasian/ Elster
19	Gravel and sand sedimentation („Nagelfluh")	„Terrazzo"-Terrace					
18	Deep valley ero-sion, Karstification (peneplains)	Deep valleys, sharp piedmont angles (?)		Lake phase	Rooikop-Transgression		Milazzo/ Cromerian
17	Intensive Calcification	Main gravel cal-crete, „Main Calcrete"		Main calcrete cap	Post-Nonidas-Regression	**Omaruru-Era**	Kageran/ Menap (?)
16	Gravel sedimentation	Gravel layer	Equus sandwithi, Elephas recki, Early Acheulian	Lake phase	Nonidas-Transgression		Sizil/ Waalian (?)
15	Calcification (bigger nodules)	Nodular cemen-ted calcrete			Post-Goanikontes-Regression	**Messum-Era**	Eburon (?)
14	Erosion	3^{rd} unconformity					
13	Soil formation	Fe-Concrete, root casts		Lake phases (?)			
12	Sandy to silty sedimentation	Sand layer, minor gravels			Goanikontes-Transgressions		Calabrium/ Tegelen (?)
11	Calcification (nodular)	Nodular calcrete					
10	Intensive soil formation, wet (?)	Fe-concrete, soil colour					
9	Fluvial sand sedimentation	Sandy layer					
8	Slight calcification	Nodular Crusts					
7	Gravel deposition	Gravel layer					
6	Calcification and calcrete formation	Calcrete		„Etosha Limestone" ?	Postpliozäne Regression	**Hoanib-Era**	Pre-Tegelen
5	Erosion	2^{nd} unconformity					
4	Main gravels (dia-meter > 80 cm)	Gravel layer	*„Notohipparion namaquense"*	Deflection of the Kunene	Anichab-Transgression		
3	Erosion and karstification	Base unconfor-mity (1^{st} unc.)		Upper Andoni Formation ?			
2	Calcification	Regolith-Calcrete			End-Miocene Regression		
1	Saprolitization	Fossil Kaolinites and Ferralsols			Up to Middle Miocene		

6.4 Conclusions

Etosha Pan, a large endorheic depression situated at the heart of the Etosha National Park was investigated from a geomorphological perspective. At stake was the origin of this depression. Remote Sensing and auxiliary data were collated and systematically interpreted, analyzed and synthesized to that effect. In essence, the use of relevant Remote Sensing data enhanced the induction and deduction processes that led to locating appropriate investigation sites. The data obtained effectively demonstrated that the Etosha Pan is presently an aggradational landform, contrary to conclusions reached by recent studies. Aggradational phases alternated with erosive phases and phases of relative stability and soil formation.

Along with a surplus of field evidence, the results demonstrated that Etosha Pan is a vestige of a large lake, termed Lake Kunene, which existed most likely until the Late Pliocene. The intervening period was marked by a smaller and sporadic Lake Etosha, which owes its development to the diversion of the Kunene River from the Owambo Basin. This diversion was sparked by intense pluvial phases which eroded deeply cut gorges and the passive, mild but widespread tectonic upheaval that affected much of the region during the Pliocene. Since then, Lake Etosha evolved as an independent entity and culminated into what we call Etosha Pan today. Conclusions reached here inevitably dismissed previous assertions that Etosha Pan is essentially a consequent of relentless morpho-dynamic processes of aeolian deflation, coupled with weathering of sediments induced by local seasonal inflow under a climatic regime similar to today.

The final desiccation of Lake Etosha is ascribed to climatic change. The timing of this last desiccation was not fully established. However, the last recorded episode of high lake levels was as recent as the middle Holocene (Holocene climatic optimum), while historical account acknowledged the occurrence of semi-permanent inundations at least at Fisher's Pan. A possible future climatic change towards humid phases will probably cause a perennial lake again, but due to the progressive tectonic situation a very high lake level is necessary to produce erosive conditions with an outflow. Tectonics at present seems to overprint the situation of the pan on the base level, so that hardly any fluvial output of material will be possible, unlike the situation in the Early and Middle Pleistocene.

In summary,

- The Etosha Region presents a chain of geomorphological events typical for a Cainozoic climatic history with humid tropical conditions until the Middle Miocene, a pronounced change at the change of Miocene to Pliocene, a shorter rebound into tropoid conditions with climatic deterioration at the end of the Pliocene and frequent Pleistocene changes.
- The classical Pleistocene record is substantially modified by neo-tectonic activity that led to shifts in the fluvial system as well as to the local base level of erosion and sedimentation. Due to the levelled topography, even little tectonic activity was able to cause severe changes in the relevant catchments.
- Etosha Pan shows a lake history with perennial wet ("pluvial"), perennial dry and episodical wet phases of non-uniform extension, as evidenced from terrace galleries around the depression and along mayor fluvial feeders.
- The only well-dated perennial lake phase is probably the youngest and coincides with the youngest soil formation on beach ridges and bars of Okondeka type. Assuming that similar soil formations of different ages on other beach ridges are also coinciding with perennial wet phases, there are at least three such phases during the last 140 ka. It is highly likely that further back in the Pleistocene this record continues, although many indicators are overprinted by later processes.

The role of karstic processes is not fully understood at present. From the bordering Otavi Mountain area, a steady inflow of carbonate-rich water on certain groundwater levels occurs. Only the upper level interflow appears as contact springs on the southern rim of the present pan. From the northern regions, water of high salinity content is brought into the system, which is reminiscent of periods when the greater Lake Kunene dried up and the present catchment developed under progressively terrestrial conditions. The centre of Etosha Pan is also the base level for northern and southern groundwater streams as evidenced by a chain of artesian springs.

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Appendix A: *Selected sedimentological properties of soil samples*

Sample No	Course Sand (%)	Medium Sand (%)	Fine Sand (%)	Course Silt (%)	Medium Silt (%)	Fine silt (%)	Clay (%)	All Sand (%)	pH	$CaCO_3$
L 1 (c)	2.21	11.37	41.86	10.34	7.90	18.38	6.99	*55.44*	7.67	10.34
L 2 (c)	2.66	7.29	38.52	17.37	26.55	11.33	0.51	*48.47*	7.77	36.44
L3 (2m) (c)	1.26	19.38	24.61	25.11	9.84	12.96	7.65	*45.25*	10.17	56.89
L3 (6m)	1.44	8.17	32.80	30.23	3.80	9.25	4.67	*42.41*	9.07	71.03
L3 (7m)	1.88	14.06	35.53	22.95	5.94	10.31	7.41	*51.48*	8.30	42.49
L3 (7.5 m)	1.04	6.96	19.50	23.79	13.61	3.17	23.85	*27.50*	8.85	72.04
L4 (120 cm) (c)	4.92	36.91	45.49	2.59	7.36	7.10	10.70	*87.32*	8.89	14.10
L4 (20-30 cm)	1.69	34.73	47.54	6.15	3.65	7.83	7.08	*83.96*	8.13	13.33
L 5(c)	0.20	34.64	48.03	4.49	3.79	10.11	4.45	*82.87*	8.14	54.33
L 6	0.85	27.20	42.49	4.54	6.25	6.74	15.49	*70.54*	9.51	32.81
L 7(c)	0.44	27.64	41.83	6.20	5.92	12.34	4.96	*69.91*	9.15	31.63
L 8	5.28	44.00	35.83	5.41	3.25	9.62	0.00	*85.12*	8.52	37.42
L 9	1.33	35.63	45.00	1.32	1.98	7.02	6.24	*81.96*	8.33	24.72
L 10	5.57	61.77	21.58	3.97	6.15	1.03	5.42	*88.92*	9.38	1.45
L 11	3.20	30.30	42.10	8.25	3.58	8.19	8.11	*75.60*	7.92	17.09
L 12(c)	2.50	41.36	32.97	4.35	5.71	7.23	5.11	*76.83*	8.28	10.99
L 13(c)	4.52	36.45	35.89	2.88	4.24	8.35	7.17	*76.85*	8.51	17.70
L 30	0.28	36.69	52.82	0.78	0.00	7.21	1.80	*89.79*	8.27	64.60
K1	4.47	54.21	30.98	1.84	0.00	5.07	6.64	*89.66*	4.01	0.33
Eto5	2.90	35.59	55.25	0.10	0.70	5.53	3.96	*93.74*	9.85	0.42
Eto202	0.76	2.24	18.95	19.23	13.02	15.26	33.60	*21.95*		
Eto204	1.06	8.27	16.69	9.40	21.53	17.02	33.73	*26.02*		
Eto206	0.19	22.51	40.49	7.80	4.10	13.34	17.36	*63.19*		
Eto207	0.25	33.79	37.65	8.00	5.00	11.64	6.76	*71.69*		
Eto208	0.47	23.24	53.86	7.72	3.81	8.26	6.37	*77.57*		
Eto209	4.37	32.70	31.81	8.37	7.86	13.52	8.02	*68.89*		
Eto211	1.53	18.14	62.87	7.40	2.50	7.54	5.96	*82.54*		
Eto234	0.12	21.77	46.03	7.00	4.10	14.34	10.66	*67.91*		

Appendix B: *Thickness measurements of loose sediments in the Etosha Pan*

Site	Easting	Northing	Total depth reached (cm)	Bedrock reached?	Remarks
Eto 202	596423	7886759	70	N	Sediment's compaction prevented further augering; profile made up of olive green clay.
Eto 27	594498	7893916	70	Y	Greyish clay up to 20 cm, followed by olive green clay for the rest of depth.
Eto 21	665050	7912896	80	Y	Very compacted olive green clay; aardvark hole that led us to site.
Eto 05	608429	7934451	90	Y	Fluvial sand (0-50 cm); olive sandy clay for rest of the depth (Ekuma Delta).
Eto 224	654798	7926997	90	N	Retrieval of soil auger becoming impossible due to heavy clay; profile composed of olive green clay.
Eto 33	592897	7903482	90	N	Clay too heavy for further augering; dark clay up to 20 cm, olive green clay for the rest of the depth.
Eto 13	703079	7921654	100	Y	Sediment succession similar to Eto 12, except that light grey sandy soil appears at 60 cm depth with 10 cm thickness, and cemented rock was hit at 100 cm.
Eto 205	602182	7897405	100	N	Augering aborted; 20 cm greyish clay, and olive green clay for the rest of the depth
Eto 215	606696	7935782	100	Y	Cemented layer penetrated for about 10 cm (to 110 cm); fish fossils found at 70 to 90 cm depth.
Eto 01	604433	7885386	110	N	Dark grey sandy clay layer (0-30 cm); medium grey sandy clay for the rest of the depth (alluvial fan)
Eto 216	671911	7930494	110	N	Augering aborted due to heavy clay; olive green clay from 0-70 cm, dark clay from 70 - 100 cm, then olive green clay for the remaining 10 cm.
Eto 10	680119	7930268	130	N	Sediment compaction prevented further soil augering; sediment succession similar to Eto 9, except that the light sandy layer was encountered at about 50 cm, with similar thickness
Eto 218	640318	7924418	130	N	Retrieval of soil auger becoming impossible due to heavy clay; profile composed of olive green clay; fish fossils fragments encountered at 70 to 100 cm depth; grass patches in the area.
Eto 220	621992	7916421	130	N	Sediment's compaction prevented further augering; profile composed of olive green clay.
Eto 06	615652	7941604	140	Y	Fluvial sand (0-50 cm); olive sandy clay for rest of the depth
Eto 23	636202	7898528	150	N	Clay too heavy for further augering; dark clay up to 30 cm, olive green clay for the rest of the depth; groundwater seeped in at 70 cm, but moisture decreased with depth after that depth.
Eto 219	631362	7920035	150	N	Compaction of sediment making it harder for soil auger to penetrate; profile composed of olive green clay.
Eto 221	611372	7913413	150	N	Sediment's compaction prevented further augering; profile composed of olive green clay.
Eto 09	677808	7929356	160	N	Olive green clay up to 90 cm followed by light sandy layer 15-20 cm in thickness, then back to olive green clay till to the end of the depth.
Eto 28	591546	7903005	160	Y	Cemented outcrop on one side of the hole prevented further augering; greyish clay for the first 20 cm, followed by olive green clay for the rest of the depth.
Eto 222	600093	7908608	195	Y	Olive green clay till 170 cm, then unusually sandy clay layer at 170 - 190 cm, which also contained fish fossils, before hitting a cemented layer at 195 cm.
Eto 25	620169	7897398	230	N	Augering aborted; medium grey clay up to 90 cm, then olive green clay for the rest of the depth
Eto 32	596836	7924015	230	Y	Profile composed of olive green clay.
Eto 12	702020	7920516	240	Y	Dark clay up to 50 cm, 20 cm thick of light grey sandy soil, then greenish sandy soil grading into greenish sandy-clay sediments; rocky layer at 240 cm prevented further augering.
Eto 31	714280	7925275	370	N	Augering aborted; clayey sand, greenish in colour but fading with depth, while pH (>9) increases with depth.
Eto 217	659370	7930636	390	N	Augering aborted; olive green clay till 3.80, where is darkened slightly; platy black (organic) material found at 90 cm.