THE SUB-KALAHARI GEOLOGY AND TECTONIC EVOLUTION OF THE KALAHARI BASIN, SOUTHERN AFRICA.

by

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ABSTRACT

Geophysical, borehole and mapped data from the Kalahari Basin were used to create maps of the sub-Kalahari geology, isopachs of the Kalahari Group and basal gravels and a sub-Kalahari topographical surface. These are the first basin-wide maps of this type to be produced. These new data were interpreted with the aid of an extensive literature review as well as data gathered at three localities in the southern part of the Kalahari Basin and enabled several conclusions to be made regarding the tectonic evolution of the area.

The sub-Kalahari Geological Map shows that rocks dating from the Archaean to present are exposed on the edges of the basin as well as covered by the Kalahari Group sedimentary rocks. Many of the rocks shown on the sub-Kalahari geological map record a history of rifting and subsequent collision, with the NE and SW trending structures appearing to have been reactivated at various times in the geological past. The extent of Karoo Supergroup rocks is greater than previously thought and Karoo sedimentary and volcanic rocks cover a large percentage of the sub-Kalahari surface. The Karoo Supergroup lithologies have been intruded by dolerite dykes and sills and the massive Botswana Dyke Swarm is shown on the sub-Kalahari map extending in a northwest direction across Botswana.

The subtraction of the thicknesses of Kalahari Group sediments from the current topographical digital elevation model (DEM) of Africa in order to prepare a DEM of the sub-Kalahari topographical surface and the preparation of an isopach map of the basal gravels gives some indication of the courses followed by Mid-Cretaceous rivers. Topographic profiles along the proposed courses of these rivers show that the floor of the Kalahari Basin has a particularly low elevation in certain areas suggesting that downwarp of the interior of the basin rather than adjacent uplift was the driving force behind Kalahari Group sedimentation. When down-warp of the Kalahari Basin began in the Late Cretaceous these rivers were back-tilted into the newly formed basin and deposition of the Kalahari Group sediments began. The basal unit of the Kalahari Group consists of gravels deposited by the Cretaceous rivers as well as on scree slopes. As down-warp of the basin continued, so more gravels were deposited as well as the sand and

finer sediment carried by the rivers. Thick clay beds accumulated in the lakes that formed by the back-tilted rivers, with sandstone being deposited in braided streams interfingering with the clays and covering them in some areas as the shallow lakes filled up with sediment.

During the Mid-Miocene there was a period of tectonic stability that saw the silcretisation and calcretisation of older Kalahari Group lithologies. At the end of the Miocene there was some uplift along the eastern side of southern Africa as well as along certain epeirogenic axes in the interior. In general this uplift was fairly gentle. Later more significant uplift in the Pliocene possibly elevated Kalahari Group and Karoo Supergroup sedimentary rocks above the basin floor and exposed many of them to erosion. The eroded sand was washed into the basin and reworked into dunes during drier periods. This uplift occurred along epeirogenic axes and was greater than the Miocene uplift.

The development of the East African Rift System (EARS) in the Late Eocene or Oligocene has had a significant influence on the Kalahari Basin. Reactivation of older NE-SW trends by SW-propagating rifts extending from the main EARS is evident by recent movement along faults along the Damara Belt and those that were associated with Karoo sedimentation and post-Karoo graben formation. The propagating rifts have resulted in uplifting, faulting and in some cases, graben formation. In some cases lakes have formed in the grabens or half-grabens themselves and in other cases they have been formed between the uplifted arches related to parallel rifts. The propagating rifts have had a strong influence on the drainage patterns and shape of the Kalahari Basin, in particular in the middle parts of the basin where they have controlled the formation of the Okavango Delta and the Makgadikgadi pans.

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CHAPTER 1: INTRODUCTION

While much research has been conducted on the Kalahari Group in the last one hundred years, from Passarge's (1904) detailed account of the geology of Botswana, *Die Kalahari*, to the application of modern dating techniques (e.g. Thomas *et al.*, 2000; Stokes *et al.*, 1997a,b) and the more recent syntheses of data (e.g. Thomas and Shaw, 1990, 1991a, 2002; Haddon, 2000; Haddon and McCarthy, in press), inadequate attention has been paid to understanding the tectonic evolution of the region in general and the development of the Kalahari Basin in particular. An understanding of the geology of the Kalahari Basin, as well as the sub-Kalahari geology and structure, is vital to the understanding of the geology and geomorphology of southern Africa, as Kalahari Group sedimentary rocks cover large portions of seven southern African countries (Fig. 1.1), stretching some 2200 km from South Africa in the south, northwards through Botswana, and up into Angola. In northern Angola, the Kalahari Basin appears to narrow before extending about 200 km into the Democratic Republic of the Congo, but Kalahari Group sediments were probably originally deposited over a large part of the Democratic Republic of the Congo as well as into countries even further to the north. A large percentage of the Kalahari Group sediments deposited in the Congo Basin appear to have subsequently been eroded by the Congo River and its tributaries, but some of the Cretaceous deposits found in the Congo Basin may be the equivalent of basal Kalahari Group deposits found further south. The upper unit of the Kalahari Group, the unconsolidated sands, cover an area of some 2.5 million km^2 (Thomas and Shaw, 1991a), which according to Baillieul (1975) makes it the largest continuous erg on earth. A country like Angola, with its vast potential wealth and largely unexplored interior, has more than 55% of its land area covered by the unconsolidated Kalahari sands; 30% of Namibia and 25% of Zambia are covered and Botswana, with its massive diamond wealth, has a cover of Kalahari sedimentary rocks over some 75% of the country.

The name Kalahari itself has many connotations and it is important to establish the terminology to be used in this work. "Kalahari" (derived from the word Kgalagadi) means "the great thirst" and the Kalahari Desert is perhaps best known for its arid climate, the Okavango swamps and the indigenous San (otherwise known as Bushmen or Basarwa) people living in the region.

Fig 1.1

This semi-arid area stretches from the northern Cape Province of South Africa to northern Botswana and Namibia, and is therefore far smaller than the area covered by the Kalahari Group sedimentary rocks. In a geological sense, the term "Kalahari Basin" is also misleading. In several publications, of which Johnson *et al.* (1996) is one of the most recent and notable, the Kalahari Basin is the name given to the Karoo Supergroup-filled sub-basin that trends roughly northeast-southwest across central Botswana. This obviously results in the unsatisfactory situation whereby the Kalahari Basin used in the Karoo context has nothing to do with the Kalahari Group itself. It is proposed after consultation with Dr M.R. Johnson, the secretary of the South Africa Committee for Stratigraphy (SACS), that the Karoo sub-basin centred on Botswana be referred to as the Botswana Basin (e.g. R.A. Smith, 1984; Partridge, 1998; Partridge and Maud, 2000). In this study, the term Kalahari Basin will be used in reference to the depositional basin into which Kalahari Group sedimentary rocks were deposited.

Climate

The southern part of the Kalahari Basin falls within the borders of Namibia, Botswana, Zimbabwe and South Africa and is characterised by semi-arid to arid conditions. The region is, however, characterised by a strong NE-SW precipitation gradient which is due to the contrasts in sea surface temperature between the South Atlantic and Indian Oceans off the coasts of southern Africa (Stokes *et al.*, 1997b). Mean precipitation is around 150 to 250 mm per annum in the southwest (Tyson, 1986; van Rooyen and Bredenkamp, 1996a), about 600 mm per annum in northern Botswana, 972 mm per annum in western Zambia (Thomas, 1984a; Thomas *et al.*, 2003) and 1400 mm in northern Angola and southern Democratic Republic of Congo (Vetter, 2004). Evaporation rates are very high, and in the summer months can be up to six times greater than the precipitation (Meyer *et al.*, 1985). Daily ranges of temperature can be extreme, with temperatures in the southwestern basin ranging from -10°C in winter and up to 45°C in summer (van Rooyen and Bredenkamp, 1996a).

Vegetation

More than 400 species of plant are found in the Kalahari with a gradation from the drier open shrublands of the extreme south, through the savannas of the central basin, to the broad leafed deciduous forests of the north. Most of the area falls within the Savannah biome, which is predominantly characterised by grasses of mainly the C_4-type, as well as shrubs and trees. In the southwestern parts of the basin the deep sandy soils of the Kgalagadi Transfrontier Park are characterised by Camel Thorn *(Acacia erioloba*) trees, Belly Thorn (*Acacia luederitzii*), Shepard's Tree *(Boscia albitrunca*), as well as Silver Clusterleaf (*Terminalia sericea*) occurring on some dune crests (van Rooyen and Bredenkamp, 1996a; Weare and Yalala, 1971). To the east and southwest of this area Grey Camel Thorn (*Acacia haematoxylon*) becomes more common as well as shrubs of Black Thorn (*Acacia mellifera*), Weeping Candle Thorn (*Acacia hebeclada*), Karee Thorn (*Lycium hirsutum*) and scattered individual plants of Bastard Roughleaf Raisin (*Greia retinervis*) and Kalahari Currant (*Rhus tenuinervis*)(van Rooyen and Bredenkamp, 1996b,c). A surface creeping plant, the Tsamma (*Citrullus lanatus*) is found throughout the drier portions of the Kalahari Basin and along with gemsbok cucumbers (*Acanthosicyos naudinianus*), and wild cucumbers (*Cucumis africanus*) are important sources of water for various antelope species. The Wilde Okkerneut tree (*Ricinodendron Rautennii*) is also known for an edible fruit (the Manketti nut). Eastern Namibia and central and southeastern Botswana are characterised by Silver Clusterleaf, Camel Thorn, Wild Seringa (*Burkea africana*), African wattle (*Peltophorum africanum*) and Knob Thorn (*Acacia nigrescens*)(Weare and Yalala, 1971), with areas of shallower bedrock in western Botswana additionally characterised by the Red Umbrella thorn *(Acacia reficiens)* and Russet Bushwillow *(Combretum hereoense*)(Cole and Le Roex, 1978). Kalahari Sand Apple (*Lonchocarpus nelsii*), stretches across the central part of the basin from Namibia into western Zimbabwe (Weare and Yalala, 1971) and the trees and shrubs and areas to the south of N'Gwaku Pan are characterised by woodlands of Umbrella Thorn (*Acacia tortilis*) and the Blue Thorn (*Acacia erubescens)*(Cole and Le Roex, 1978). Areas of calcrete development in the entire area from Windhoek in Namibia through to the southwestern parts of the Okavango Delta are characterised by Trumpet Thorns *(Catophractes alexandrii*)(Cole and Le Roex, 1978) and Baobab trees (*Adansonia digitata*) are found around the Makgadikgadi pans. Towards the north and northeast of the basin, in Zimbabwe, Zambia, Angola and into the Democratic Republic of Congo, the climate becomes progressively wetter, with the dune ridges characterised by woodlands (Van Zinderen Bakker and Clark, 1962). Mopane (*Colophospermum mopane*), Rhodesian teak (*Baikiaea plurijuga*) and Marula (*Sclerocarya birrea*) trees are found in the dry deciduous forest of northeastern Namibia, southern Angola, northern Botswana, northwestern Zimbabwe and southern Zambia, with the dunes of western Zimbabwe additionally characterised by Musasa/Mtundu (*Brachystegia spiciformis*)and Mukwa (*Pterocarpus angolensis*) (Sutton, 1979). Western Zambia and central and eastern Angola are characterised by woodland dominated by the African Blackwood (*Erythrophleum Africanum*), Ebony (*Diospyros batocana*) and trees belonging to the *Brachystegia* genera, while the dense woodlands of northwestern Zambia are also characterised by Mukwe (*Cryptosepalum exfoliatum pseudotaxus)* (Vetter, 2004). The Okavango Delta and adjacent wetland areas are characterised by swamps filled with reeds and other aquatic plants.

Human activity

Evidence of human activity in the Kalahari extends as far back as the Early Stone Age with artefacts from this period being found at various sites in and around the basin (see review in Thomas and Shaw, 1991a). Until as recently as 2000-1500 years ago the Kalahari was inhabited exclusively by the San people who survived as hunters and gatherers. The arrival of Bantu people and European settlers changed the way of life in the region, with permanent farms becoming dominant. Today large game, cattle, sheep and goat farms cover the majority of the area along with several reserves set aside for the preservation of flora and fauna. These reserves cater for a large tourist industry attracted by the high concentration of wildlife as well as the beauty of the Kalahari Desert and the Okavango Delta and Makgadikgadi and Etosha pans. Today, major settlements are predominantly found around the edges of the basin although some small towns have developed around the sites of mines (e.g. Orapa) while others have seen growth along with a boom in the tourist industry (e.g. Maun). The regional road network becomes markedly less dense in the basin itself, although a trans-Kalahari highway now links South Africa and Gaborone with Namibia (Fig.1.2).

Aims

This study concentrates on the area covered by Kalahari Group deposits to the south of the Democratic Republic of the Congo. It is the aim of this study not only to present a synthesis of available information for the entire Kalahari Basin, but also to generate new data, model and combine existing data and present the resultant interpretations of Kalahari Basin evolution. Previously published and unpublished as well as new borehole, geophysical, and geological data are used to produce maps showing the thickness of Kalahari Group sedimentary rocks, the geology underlying the Kalahari Group rocks, and the topography of the sub-Kalahari surface. The large size of the area covered and the complexity of the geology does not enable all aspects of the Kalahari Basin to be covered in equal detail, but an attempt has been made to describe the

sub-Kalahari geology as well as the Kalahari Group stratigraphy and to identify those factors deemed to have played an important part in the basin evolution. The southern African rivers and their distribution both during and after basin formation provide important evidence for basin evolution and an effort has been made to understand the factors which controlled the drainage patterns. The timing of basin formation and the deposition of Kalahari Group sedimentary rocks is discussed and the tectonic influences that have acted on the basin subsequently are described. A brief chapter on the mineral potential of the area is included in order to illustrate the potential the region has for mineral exploration and it is hoped that this work will help to promote interest in furthering research, development and mineral exploration in the area.

CHAPTER 2: METHODOLOGY

2.1 Introduction

The Southern African Development Community (SADC) Kalahari Working Group provided the initial impetus for a Kalahari Basin study by identifying that there was a need to produce maps showing the thickness of Kalahari Group sedimentary rocks as well as the distribution and nature of the geological units beneath the Kalahari Group rocks for the SADC region. The "Isopach Map of the Kalahari Group" and "Sub-Kalahari Geological Map" (see Appendix B and C) to be produced were to show the thicknesses of Kalahari Group rocks and the geology underlying the Kalahari Group respectively for South Africa, Namibia, Botswana, Zimbabwe, Zambia, and Angola. The Democratic Republic of Congo (DRC) was not included at the time of the project initiation as at that stage it was not a member of SADC. All data compilation and digital capture as well as construction of databases and GIS manipulation of the data was carried out by the author who was assisted with some of the more complicated GIS modelling by various staff at the Council for Geoscience.

Following the preparation of the two maps it was clear that further research on the Kalahari Basin needed to be conducted, firstly in order to complement and provide an explanation for the two maps and secondly to identify the sequence of tectonic and climatic events that influenced the formation and evolution of the Kalahari Basin. The substantial amount of literature and data that exists for the Kalahari Basin provided the major source of information for this study and the compilation and synthesis of this information was seen as one of the major objectives of the project. Field and laboratory descriptions of Kalahari Group rocks as well as the interpretation of the maps provided additional data and insight into the evolution of the Kalahari Basin. The production of a map showing the thicknesses of the basal gravels of the Kalahari Group in South Africa (Appendix D) was attempted in order to better understand where the main palaeo-channels lie as this is not only of importance to understanding basin evolution but also has economic significance for diamond and water exploration. While the isopach maps provide valuable information, it was also decided to model the sub-Kalahari Group topography in order to better understand basin evolution. The production of the Sub-Kalahari topographical surface (Appendix E) has provided valuable new information regarding basin development.

2.2 The compilation of the Kalahari Group Isopach, Basement Gravel Isopach, Sub-Kalahari Geological and Sub-Kalahari Topographical Maps

2.2.1 Borehole database compilation

Information for the Kalahari isopach and sub-Kalahari geology maps was obtained from national representatives on the Kalahari Working Group, directly from the various national Geological Surveys, from several mining and mineral exploration companies and from private geological and groundwater consultancies. The data used for the compilation of the maps came in four main forms:

- borehole data (Kalahari Group lithologies and thicknesses and basement rock types)
- geological maps
- geological reports, journal papers, published books
- geophysical data

In order to gather this information visits were made to the Geological Survey Departments of Botswana, Namibia, Zambia, Angola and Zimbabwe, as well as to the Department of Water Affairs offices in Namibia. In addition to this, private consultancies and mining companies that had borehole information were approached in each of these countries during each visit. On average a period of one and a half to two weeks was spent in each country gathering relevant data.

In excess of 30 000 boreholes drilled into the Kalahari Group sediments were analysed, evaluated and, depending on their reliability and detail, it was decided which of these would eventually be used for the production of the map. Each borehole log was individually assessed according to the following criteria:

- accuracy of locality data
- accuracy and detail of lithological descriptions

The borehole logs that were considered to contain data with high levels of accuracy according to these criteria were then entered into a new Paradox-based database. The borehole data was then sorted, edited, evaluated, assigned a unique number and individually coded to enable data such as thickness of the Kalahari Group sedimentary rocks and sub-Kalahari lithology to be easily queried and extracted. Those boreholes which did not reach the base of the Kalahari were assigned different codes and their logs were kept separately for later manual interpretation of sediment thicknesses for those areas with sparse data coverage. Where possible, borehole core and percussion chips from various countries were examined in order to more accurately ascertain the depth of the base of the Kalahari Group, and what the underlying lithologies consist of. Where the lithologies described in the logs appeared confused or unreliable, the data from these logs was discarded.

Of the initial 30 000 boreholes only 7 074 were thought to be detailed and reliable enough for use in the production of the isopach and sub-Kalahari geology maps. The majority of the data used falls within the borders of South Africa, Botswana and Namibia, and very little data exists for Angola and Zambia (see "Isopach Map of the Kalahari Group" Appendix B). From these 7 074 borehole logs two separate data files were created in Paradox: the first (co-ordinate file) containing unique number, longitude and latitude and the second (data file) containing unique number, borehole number, thickness of Kalahari Group rocks and sub-Kalahari lithology. The borehole data was then exported as ASCII comma delimited files and imported into the ArcINFO 6 GIS system. The gathering and preparation of the borehole data took about one year.

2.2.2 Borehole data processing

In ArcINFO 6 the borehole data co-ordinate files were generated and then built as point coverages. This created a point attribute table (PAT). INFO was then used to define the structure of an attributes table into which the data from the data file would be placed. Data was then added to the newly defined attribute table, and this in turn was joined to the points table. The result of this was to create a coverage showing the positions of the boreholes with attached lithological and thickness data.

2.2.3 Addition of mapped and published data

2.2.3.1 Kalahari isopach map

The areas on the isopach map where the sub-Kalahari rocks are exposed form an important additional source of thickness data that needed to be included on the map. Published and unpublished 1: 250 000 and 1: 1000 000 geological maps of the region were used to delineate the furthest most extent of the Kalahari Group sedimentary rocks as well as to show where pre-Kalahari rocks outcrop. Twenty 1: 250 000 maps were used as a source of data for Namibia, and 12 for South Africa. For the other countries, 1: 1000 000 scale maps were mainly used. Overlays showing the outcrop areas of sub-Kalahari rocks were drawn on a stable base in preparation for digitising.

Kalahari Group isopachs shown on earlier maps in the Morokweng area of South Africa (Smit, 1977) were digitised and brought into ArcINFO as a separate coverage. Where new data was available, it was used to modify these isopachs. The Morokweng isopachs were added to the borehole data in order to improve data density in areas of limited borehole coverage.

Isopachs from the published 1: 1000 000 geological map of Angola (Serviço Geológico de Angola, 1988) were included as borehole data from Angola was sparse and in many cases was considered to be inaccurate. The source of the data used for the Angolan isopachs and the accuracy of the isopachs is not known and it is possible that some pre-Kalahari Group Cretaceous sedimentary rocks may have been included with the Kalahari Group. Unfortunately it is extremely difficult to obtain accurate and reliable borehole logs from the areas of Angola covered by Kalahari Group rocks.

2.2.3.2 Basal gravels isopach map

The same base data used for the isopach map of the Kalahari Group was used for the construction of the isopachs of the basal gravels.

2.2.3.3 Sub-Kalahari geology map

In order to prepare a base map onto which borehole data could be added and interpreted, all available geological maps for the area were consulted. In the case of some of the countries, 1: 250 000 scale geological maps exist, and these were used as a base map onto which other data could be added. For Namibia, 20 1: 250 000 maps exist for the area covered by the Kalahari Group rocks and for some of these areas sub-Kalahari geology had been interpreted by Dr W. Hegenberger of the Geological Survey of Namibia. The latest edition of the 1: 1000 000 geological map of Botswana already shows the sub-Kalahari geology and as a result was added with only minor changes where new data had become available. For Angola, Zambia and Zimbabwe the national 1: 1000 000 scale geological maps were used as base maps, and for South Africa 12 separate 1: 250 000 maps were used for the base maps, none of which showed any sub-Kalahari geology.

Borehole data containing sub-Kalahari lithological information was then overlain on the geological maps of all the countries. The sub-Kalahari data retrieved from these boreholes was directly used to determine the distribution of geological units underlying the Kalahari Group rocks. Geological detail down to formation level is shown where possible. Where there were large gaps in borehole distribution and the geological maps did not indicate the nature of sub-Kalahari lithologies, all published and unpublished geological reports from the area were studied for information. Geophysical data, which included previously interpreted and raw data, was used particularly in areas where other data was absent. In particular, aeromagnetic and gravity data enabled further interpolations to be made by highlighting some of the most important pre-Kalahari structures and magnetic units.

A “buffer zone” of outcropping pre-Kalahari units was included on the map around the edge of the Kalahari Basin in order to provide continuity between the interpolated geology beneath the Kalahari sedimentary rocks and the surface geology that has been mapped previously.

2.2.3.4 Sub-Kalahari Group topographical map

The same base data used for the isopach map of the Kalahari Group was used for the construction of the sub-Kalahari topography. This involved selecting areas of sub-Kalahari outcrop as well as adding in some of the previously mapped isopachs from areas where borehole data was insufficient. The DEM used in the GIS processing of the data is the USGS SRTM30 DEM, which is the original GTOPO30 DEM enhanced by data from the Shuttle Radar Topography Mission flown in February 2000 (USGS, 2004).

2.2.4 GIS manipulation of data

The necessary information from each of the maps showing pre-Kalahari outcrops and sub-Kalahari geology was digitised using AutoCAD, with these digitised maps then being exported as DXF files and then imported into ArcINFO. TIC coverages for each map were created in ArcINFO and the maps were then “transformed” to these TIC coverages. Once the maps were geographically referenced and were in the Arc Info system the following steps were followed:

1. All maps were projected to a common projection.
2. The ArcEdit command “Get” was used to bring individual maps into a common coverage for each country.
3. Polygons and arcs were edited in ArcEdit in order to produce composite maps of individual countries, and the individual map coverages, now geo-referenced, were then edited and finally were joined together using ArcEdit.

2.2.4.1 Isopach maps

ArcINFO 6 software was chosen to model the borehole data and create the isopach maps of the entire Kalahari Group and of the basal gravels. Other contouring programs were felt to be inadequate for the manipulation of the amounts of data to be used.

The digitised pre-Kalahari outcrops needed to be contoured along with the borehole data so that both would contribute to the isopach map. As the digitised data was in the form of polygons or lines and the borehole data was point data, it was first necessary to convert the digitised data to point data. In order to do this, the digitised pre-Kalahari polygons were given a value for thickness of zero, and a “TIN” was then created using both the borehole and the zero thickness coverages. This process created a three-dimensional surface which was then further manipulated using “Tinlattice” and a “focal mean”, before the data was contoured using “lattice contour”.

In order to produce a map that was a size that was convenient to handle, it was decided to prepare the map at a scale of 1: 2 500 000.

2.2.4.2 Sub-Kalahari geology map

Once the digitised geological maps had been joined together and edited, they were built as a polygon coverage and coded, with labels assigned to each polygon and these in turn linked to a look up table which defines the colours to be allocated to each polygon.

The correlation of lithostratigraphic units across international boundaries and the preparation of a common legend for the map involved an extensive literature survey and a critical analysis of previous stratigraphic correlations. In some cases correlation across borders was fairly simple, particularly with the Karoo Supergroup formations for example, but in other cases, the stratigraphic sub-divisions between each country differed significantly, making the correlation process more difficult. In these instances published and unpublished age determinations, recorded field observations as well as lithological descriptions were used as a basis for the correlations. In addition to this, several meetings were organised in Windhoek, Lobatse, Pretoria and Luanda between geologists of the countries involved in order to sort out some of the stratigraphic correlation problems that arose. The legend for the sub-Kalahari geological map was created entirely for this project as no previous common legend for the stratigraphy of Southern Africa down to formation level existed. The legend and the map itself therefore reflect the correlations attempted as part of this research.

2.2.4.3 Sub-Kalahari Group topography map

The principle behind the construction of a sub-Kalahari topographical surface is quite simple. The thicknesses of the Kalahari Group, as extracted from boreholes and from additional isopachs where data is absent or sparse, is subtracted from a digital elevation model (DEM) in order to get a new height above sea-level.

In practice this procedure involves several steps:

1. Each of the four sections of the Africa 1km DEM developed by the USGS (SRTM30) was converted to a grid using the ArcINFO command IMAGEGRID. In order to represent

the negative 16-bit DEM values correctly the following formula was then run in Grid: out_grid = con(in_grid >= 32768, in_grid - 65536, in_grid)

2. The four individual grids were then brought together using the ArcINFO command <GRIDINSERT>
3. The sub-Kalahari outcrop (zero-thickness) data and any additional previously isopached data were gridded along with the borehole data using ArcINFO. The ARC command <TOPOGRID> was used to produce the grid.
4. The grid of thickness values was then subtracted from the DEM grid using ArcINFO GRID. This resulted in a new grid for the sub-Kalahari topographical surface.
5. The new DEM grid was contoured at an interval of 30 m and colours were assigned which would best highlight the changes in topography within and immediately adjacent to the Kalahari Basin.
6. A hill-shading was applied to accentuate the topography.

The surrounding regions not covered by Kalahari Group sedimentary rocks are included in order to provide a more regional idea of the topography of southern Africa. As the paucity of thickness data for some parts of the basin resulted in averaged thicknesses over large areas, deep river channels on the current surface are replicated on the sub-Kalahari topographical surface. These rivers almost certainly followed different courses during pre-Kalahari times (if indeed they existed at all) and as a result should strictly not be shown on a sub-Kalahari surface. It was, however, decided to keep them rather than manually manipulate the DEM.

2.3 Kalahari Group stratigraphy

Due to the limited number of exposures of Kalahari Group sedimentary rocks in the south of the basin it is difficult to study the lithologies in the field. This is compounded by the easily weathered nature of the sediments themselves as well as the high degree of calcretisation and silcretisation of the sediments. As a result of these factors, much of the literature on the Kalahari Group covers the upper duricrust horizons, and the unconsolidated sands which form the uppermost formation of the Kalahari Group and cover the majority of the area.

Time constraints on the project prevented extensive field work being undertaken, but field visits lasting a total of about six weeks were undertaken as part of this study to three localities in the southern Kalahari Basin (see map in Chapter 4). The first locality is at the Sishen Iron Ore Mine to the west of the town of Kathu where a massive open pit cut through the Kalahari Group sedimentary rocks provided opportunity for closer examination of the basal gravels of the Wessels Formation and overlying clays of the Budin Formation. The sequence exposed in the pit at Sishen was measured using a tape measure and Jacobs Staff and lithologies were described from hand specimen and thin section. Samples from a borehole drilled immediately to the west of the open pit were submitted to University of the Witwatersrand for analysis, with the aim of determining the degree of calcretisation of the sediments and a clay sample from inside the pit was submitted for XRD analysis at the Council for Geoscience, in order to determine the type of clay and quartz component of the sample. The results are discussed in Chapter 4. The second locality where Kalahari Group rocks were studied is along the Moshaweng River and the third locality lies within the Kalahari Gemsbok Park. In both of these two localities, calcretised sandstones of the Eden Formation exposed along the dry river valleys were described. Field observations were recorded and hand specimens were taken back to the laboratory where thin sections were made. An exposure of Kalahari Group rocks along the Moshaweng River was also measured using a tape measure and Jacobs Staff.

2.4 Palaeo-drainage reconstruction and model of basin development

Various existing models for the drainage patterns of the Cretaceous and Tertiary were critically examined in the light of the new data generated by this research as well as the combination of this new data with existing data gathered during an extensive literature survey. The lithological descriptions, isopach, sub-Kalahari geological and sub-Kalahari topographical maps produced during this study were added to palaeodrainage direction, geophysical and various other data from previous research in order to better understand the timing and nature of the Kalahari Basin formation.

Using ArcINFO 8.3 and the newly created sub-Kalahari topographical surface, topographic profiles were created along the courses of rivers which flowed across southern Africa prior to the formation of the Kalahari Basin in the Late Cretaceous. These profiles help to understand the amount of downwarp and to see areas of later uplift. All figures and maps in this thesis have been drawn by the author using Corel Draw, ArcView and ArcMap software.

CHAPTER 3: SUB-KALAHARI GEOLOGY

3.1 Introduction

Figures 3.1 and 3.2 show the aeromagnetic and gravity coverages respectively for the study area. The north-south trending Archaean Kraaipan belt partially crops out near the southern margin of the basin and its highly magnetic nature has enabled its extent to be mapped beneath the Kalahari Group sedimentary rocks (Richards, 1979). The highly magnetic Xade and Tshane (section 3.4.1) and Molopo Farms Complexes (section 3.3.9.2), all of which are completely covered by Kalahari Group sedimentary rocks, are particularly prominent on these images. Further to the south, the highly magnetic iron-formations of the Asbestos Hills Subgroup (section 3.37) extend north towards the Botswana border where they flank the southern side of the Molopo Farms Complex and western side of the Morokweng Impact structure (section 3.7), the latter of which shows up clearly on the gravity coverage (Fig. 3.2). To the west of this, north-south trending ridges of Olifantshoek Supergroup rocks forming part of the Kheis Belt (section 3.4.1) in South Africa extend into Botswana. The Kheis Belt flanks a north trending magnetic feature, the Kalahari Line, which marks an east to west transition from average level, shallow magnetic features, to deeper features with higher magnetic levels (Reeves, 1979). The magnetic signature west of the Kalahari Line suggests deeply buried highly magnetic rocks (see Fig. 3.3), while gravity fields for this area are also much higher than those over the Kheis Belt and the intermediate levels further east (Reeves, 1979). Shallow level features in this area are attributed to post-Karoo dolerite sills (Eberle *et al.*, 2002). The northeast-trending Makgadikgadi Line possibly marks the northwestern boundary of the Kaapvaal Craton. The metamorphic and granitic rocks of the Namaqua-Natal Belt run parallel to the southwestern edge of the Kalahari Basin, while the Poffadder (Pl) and Excelsior (El) Lineaments encompass a zone known as the Trans-Gondwana Konkiep Structural Zone (Corner and Swart, 1997), perhaps representing the southern African continuation of the tectonised belt separating the Alto Paraguay and Rio de la Plata cratons of South America (Corner and Swart, 1997, in: Corner, 2000). The Ghanzi-Chobe and Damara Belts (Fig. 3.4; section 3.5.2) strike eastwards from Namibia into Botswana, where the high magnetic signatures in northern Botswana can be attributed to various intrusives of mainly

Mesoproterozoic age. The

fig 3.1

Fig 3.2

Fig 3.3

Fig 3.4

Pan-African Omaruru (OL) and Autseib (AUL) lineaments are interpreted as associated features, bordering the Omaruru Lineament Zone (OMLZ) (Corner, 2000). The Grootfontein Complex (see section 3.4.6) lies along the OMLZ, while evidence from the Waterberg Fault in the Waterberg Basin of Namibia suggests that the Omaruru and Autseib Lineaments were reactivated as recently as the Cretaceous (Corner, 2000). According to Corner (2000) the Nama Lineament (NL) represents the locus of a series of parallel faults along which down-throw to the northeast into the Nama Basin has occurred. The Nama Lineament zone seems to be the direct northwest extension of the fault zone near the town of Prieska in South Africa. Dykes are highly visible on the aeromagnetic coverage, with the northwest-southeast trending dyke swarm extending across Botswana particularly prominent (see section 3.6.8).

In figure 3.3 the estimated depths to magnetic basement for Botswana highlights several sub-Kalahari Group structural features. Much of the area, including the northwestern part around Okavango, as well as the southeast where Archaean rocks outcrop is characterised by fairly shallow magnetic basement, mostly less than 250 m beneath the surface. The deepest parts occur between the Kalahari Line and Namibian border, where the magnetic basement is depressed to more than 15000 m in the Nosop and Ncojane Basins. South of the Ghanzi ridge (Ghanzi-Chobe Belt), in central Botswana, the base of the Passarge Basin (see section 3.5.4) also has an anomalous depth of over 15000 m.

3.2 Evolution of continental crust

The regional structure of the study area incorporates three main cratonic blocks, Angola, Kaapvaal and Zimbabwe, which form the core of the basement geology of the region (Fig. 3.4). The oldest rocks present in the area covered by the sub-Kalahari geological map form the basis of these cratonic blocks, and are found around the edge of the Kalahari Basin, in northern Angola, eastern Botswana and in northwestern South Africa. Kaapvaal craton assembly took place as a result of the collision between the Kraaipan arc and the continental margin. The cratons comprise mostly tonalitic and trondhjemitic gneisses and granitoids, as well as subordinate volumes of metamorphosed volcano-sedimentary rocks forming the so-called greenstone belts.

The Kaapvaal Craton extends westwards underneath the Kalahari rocks probably up to the northerly trending magnetic feature, the Kalahari Line. A borehole drilled near the Mabuasehube Game Reserve in Botswana, just to the east of the Kalahari Line intersected granite gneisses beneath the Karoo dated at 2928 ± 4 Ma (Kamo *et al.*, 1995). If the granite-gneisses are present to the west of the Kalahari Line, they appear to be much deeper (Fig. 3.3) and form the base to the Nosop-Ncojane Basin. Although Reeves and Hutchins (1982) discount the presence of the Kaapvaal west of the Kalahari Line as they postulate the presence of younger oceanic crust in this direction (see section 3.2.3.3), deep seismic data from southwest Botswana suggests that the granite gneisses of the Kaapvaal Craton extend to the west of the Kalahari Line, where the craton was thinned by stretching, due to rifting of the Kaapvaal margin (Wright and Hall, 1990). The northern boundary of the Kaapvaal Craton was previously thought to be the Zoetfontein Fault (e.g. Reeves, 1979; Hutchins and Reeves, 1980), but later work by Meixner and Peart (1984) suggests that the craton edge may continue up to the northeast trending Makgadikgadi Line (see 3.2.3.3; Fig. 3.4). The Zimbabwe Craton, which extends into eastern Botswana, is separated from the Kaapvaal Craton by the Limpopo Belt.

The Congo Craton is separated from the Kaapvaal/Zimbabwe Cratons by a broad band of Neoproterozoic reactivated crust, of the Damaran and Ghanzi-Chobe Belts. The southern boundary of the Congo Craton (which includes the older Angola and Kasai Cratons) is by no means clearly defined, and various craton margins have been proposed. From an economic point of view the occurrence of underlying cratonic material is important as diamondiferous kimberlites are primarily found only on cratonic areas. In the Tsumkwe area of northeastern Namibia garnets thought to have been derived from a mantle with a cratonic geotherm are present, which might be indicative of a source of diamondiferous kimberlite (Hoal *et al.*, 2000). Numerous kimberlites have, however, been drilled in the Angola-Namibia-Botswana border areas without diamonds being found. Although this suggests that these areas lie off the craton, the discovery of a narrow zone of relatively low electric resistivity running from Otjiwarongo in Namibia, around to the south of Okavango in Botswana, and into the Middle Zambezi Valley may indicate a zone of lithospheric weakness that probably marks the transition from the Congo Craton to the Neoproterozoic metamorphic belt to the south (de Beer, 1979). Metasedimentary rocks of uncertain age in the northwestern corner of Botswana, with a north-northwest to north structural grain, have been interpreted as possible cover sequences on the southern part of the

Congo Craton (Key and Ayres, 2000).

3.2.1 Archaean Greenstones

Archaean greenstones are present along the eastern edge of the Kalahari Basin in Zimbabwe, as well as further to the south of the basin in Botswana and South Africa where four greenstone belts make up the Kraaipan Greenstone Belt or Kraaipan Terrane (eg. Kiefer and Viljoen, 2004). The greenstone belts are usually narrow and elongated, lie on a granite base and may have developed along lineaments in thinner granitic crust or along ancient continental margins (de Wit *et al.*, 1992). The rocks in the belts comprise greenschist facies mafic to ultra-mafic volcanic or magmatic rocks as well as metasedimentary rocks (de Wit *et al.*, 2000). The earliest greenstone belts occurring in the Zimbabwe Craton formed about 3500 Ma ago (Carney *et al.*, 1994), with the Pietersberg Greenstone Belt and possibly the western Kraaipan (Stella) greenstone belt forming at around 3200 Ma (Poujol *et al.*, 2003). A phyllite from the Madibe Greenstone Belt, to the southwest of Mmabatho, yielded a SHRIMP age of 3098.4 ± 7.6 Ma (Hirner *et al.*, 2004).

3.2.2 Archaean granites and gneisses

Archaean granites and gneisses are found largely along the edge of the Kalahari Basin, and occur in all of the countries. These rocks have been subdivided and named uniquely in the different areas, but for convenience have been grouped together on the sub-Kalahari geological map (Appendix C) and are probably related to the original collision events responsible for the formation of the volcano-sedimentary greenstone belts, and craton assembly. The granitoids associated with the Kraaipan greenstone belts yielded ages between 3250 and 2735 Ma (Anhaeusser *et al.,* 1997) and recent SHRIMP U-Pb ages of 2929 ± 9 Ma and 2943 ± 9 Ma for granites adjacent to the Kraaipan Belt of southeastern Botswana are believed to date the assembly of the Kaapvaal Craton (Mapeo *et al.*, 2004a,c).

3.3 Developments on the cratons

The Witwatersrand Basin possibly developed either as a foreland basin in response to the collision of the Kraaipan arc, or as a response to the development of an Andean arc on the northern margin of the Kaapvaal Craton (Burke *et al.*, 2003). The suturing between the Kaapvaal and Zimbabwe Cratons in the Archaean, was followed by basin development in the area now spanning the boundary between the North West and Northern Cape Provinces of South Africa. The Ventersdorp Supergroup volcanic and sedimentary rocks were followed by the basal Transvaal Supergroup sedimentary deposits of the Black Reef (main Transvaal Basin) or Vryburg (Griqualand West Basin) Formations in the Neoarchaean.

3.3.1 Dominion Group and Witwatersrand Supergroup

Lava of the Dominion Group covers a small area on the sub-Kalahari geological map and their rift-related extrusion is probably related to a period of extensional tectonics on the Kaapvaal Craton (Tankard *et al.*, 1982; Poujol *et al.*, 2003). The accumulation of the Dominion Group at 3120-3070 Ma (Armstrong *et al.*, 1991) was followed by the development of the Witwatersrand Basin and the deposition of the Witwatersrand Supergroup. Deposition of the West Rand Group took place between 2970 and around 2914 ± 8 Ma (Hartzer *et al.*, 1998), with deposition of the Central Rand Group corresponding with a period of magmatic activity in the western and northern parts of the Kaapvaal Craton between 2880-2820 Ma (Poujol *et al.*, 2003).

3.3.2 Mabuasehube Group

The Mesoarchaean Mabuasehube Group, consisting of arkosic sandstone, limestone, shale, mudstone and iron formation, occurs in southern Botswana in windows through younger Olifantshoek and Karoo Supergroup sediments. The Mabuasehube Group was correlated with the Transvaal Supergroup of South Africa on the basis of lithological similarites (Key and Ayres, 2000) but it is now thought to have an age of ± 2900 Ma (T.Majaule, *pers. comm.*).

3.3.3 Lobatse Group and Kanye Formation

The Kanye Formation consists of felsites and some volcanoclastic deposits. The age of the Kanye is constrained by the Gaborone Granite which intruded into it, and is therefore older than about 2781 Ma (Carney *et al.*, 1994; Key and Ayres, 2000).

3.3.4 The Gaborone, Mmathethe, and Mosita Granites

The Gaborone Granite occurs in southeastern Botswana where it outcrops over an area of 140km east to west and 100 km from north to south (Carney *et al.*, 1994), while the Mmathethe Granite occurs approximately 20km to the southeast of Lobatse. The Gaborone Granite has given ID-TIMS ages of 2784.9 ± 1.9, 2783.1 ± 2 and 2781 Ma (Moore *et al.*, 1993) and Pb-Pb zircon evaporation ages of about 2783 Ma (Grobler and Walraven, 1993) and the Mmathethe Granite gives an age of 2775 ± 7 Ma (Key and Ayres, 2000). The Mosita Granite cropping out on the northwestern edge of the Kraaipan Belt in South Africa has an age of 2791 ± 8 Ma (Poujol *et al.*, 2000), which is similar to that of the Gaborone Granite suggesting that the two bodies may be genetically related (Poujol *et al.*, 2000). Other felsic intrusives shown on the sub-Kalahari geological map as occurring in southern Botswana possibly have a similar age to the Gaborone and Mmathethe granites (Key and Ayres, 2000).

3.3.5 Limpopo Belt

While the Limpopo Belt may not extend very far beneath the Kalahari Basin, appearing to die out at around 27°E (Key and Hutton, 1974; Reeves and Hutchins, 1975; Reeves, 1979)(Fig. 3.4), it remains an important structural feature of the region as a whole, is a very well defined zone of weakness and separates the Zimbabwe and Kaapvaal cratons. While the lithologies associated with the Limpopo Belt do not appear on the sub-Kalahari geological map, and are therefore not described in this study, the tectonic development related to the collision phase of the event is mentioned in order to further understand the tectonic evolution of the area. The Limpopo Belt is subdivided into a Central Zone, Northern Marginal Zone, and Southern Marginal Zone (Cox *et al.*, 1965). The Central Zone (CZ) was thought to have been metamorphosed at about 3150 Ma (Barton, 1983), but the syntectonic Bulai pluton suggests that the uplift occurred around 2572 Ma (Barton and Doig, 1993). The Northern Marginal Zone (NMZ) was possibly intensely deformed and metamorphosed about 2870 Ma (Cahen *et al.*, 1984), although the intrusion of syntectonic granites at 2583 Ma suggests a younger age (Kamber *et al.*, 1992; in McCourt *et al.*, 1995). Satellite bodies of the Great Dyke (~2460 Ma) are undeformed and discordant to the regional structure of the NMZ, suggesting that deformation occurred prior to its intrusion (McCourt *et al.*, 1995). The Matok Pluton in the Southern Marginal Zone (SMZ) constrains uplift in that region to ~2671 Ma (Barton and Van Reenen, 1992), with deformation completed by 2450 Ma which is the age of the post-orogenic Palmietfontein Granite (McCourt *et al.*, 1995). The peak of metamorphism probably occurred between 2700 and 2650 Ma (Barton and Van Reenen, 1992).

At around 2000 Ma there appears to have been reactivation of shear zones and high grade metamorphism in the Limpopo Belt (Eglington and Armstrong, 2004). The rocks of the Mahalapye Complex are found in the western extension of the Central zone of the Limpopo Belt and were included into the Limpopo Belt by Carney *et al.* (1994) largely on the basis of their geographical position. It is possible that the rocks of the Mahalapye Complex are related to a later event suggested by McCourt *et al.* (1995) involving uplift and crustal thinning, possibly related to the emplacement of the Bushveld Complex and development of the Soutpansberg Basin.

Rocks of the Mahalapye Complex of Botswana comprise the Mokgwane Granite, Mahalapye Migmatite and Mahalapye Granite (Carney *et al.*, 1994). The Mahalapye Migmatite is believed to be the oldest of the rocks in the complex (McCourt *et al.*, 2004a) and was intruded by the Mahalapye Granite about 2023 ± 7 Ma ago (McCourt and Armstrong, 1998). The Mahalapye Migmatite, on the eastern edge of the complex, consists of three types of migmatite (Key, 1979), of largely a granodioritic composition (Skinner, 1978). The Mahalapye Granite contains a leucocratic quartz monzanite, a granodiorite in the northwest (Ermanovics, 1980), as well as a porphyroblastic diorite in the southeast (Skinner, 1978).

3.3.6 Ventersdorp Supergroup

The Ventersdorp rifting lasted about 50 million years, beginning with the eruption of the Klipriviersberg Group basalts at about 2714 ± 8 Ma. The Makwassie Formation has a U-Pb (SHRIMP) age of 2709 ± 4 Ma (Armstrong *et al.*, 1991).

3.3.7 Transvaal Supergroup

Rocks of the Transvaal Supergroup occur in two separate areas, the Transvaal Basin which occurs around the edge of the Bushveld Complex and the Griqualand West Basin extending from Prieska in South Africa into Botswana where it is known as the Kanye Basin (Fig. 3.5).

The basal unit of the Transvaal Supergroup in the Griqualand West Basin, the Vryburg Formation, is the equivalent of the Kanye and Transvaal Basin's Black Reef Formation. It consists of shales, siltstones, quartzites, subordinate carbonates and basaltic to amygdaloidal lavas (Altermann and Siegfried, 1997) and in Botswana has an age of about 2650 Ma (Key and Ayres, 2000). As the extent of the Black Reef Formation in Botswana is not shown on the sub-Kalahari geological map, the unit was intersected in boreholes east and north of the Molopo Farms Complex in Botswana (Gould *et al.*, 1987) suggesting that it is contiguous with the Vryburg Formation of South Africa. Although inferred on the sub-Kalahari geological map as part of Botswana's Taupone Group, Aldiss *et al.* (1989) exclude the Black Reef Formation from the Taupone Group in their stratigraphy, following the practice of the South African Committee for Stratigraphy (SACS, 1980).

Overlying the Vryburg Formation, the Schmidtsdrif Subgroup consists largely of a lower unit of platform carbonates, often characterised by stromatolites, carbonate sands and oolites (Altermann and Siegfried, 1997) as well as fluvial quartz arenites typical of shallow marine and intertidal environments (Beukes, 1986). A Pb-Pb age for the Schmidtsdrif Formation stromatolitic limestones of 2557 ± 49 Ma has been obtained by Jahn *et al.* (1990).

The Campbell Rand Subgroup, covering a very large area on the sub-Kalahari geological map, consists of dolomites, limestones and cherts, with characteristic wavy laminated stromatolites (Grobbelaar *et al.*, 1995; Altermann and Siegfried, 1997).

The Asbestos Hills Subgroup conformably overlying the Campbellrand Subgroup is divided into the basal Kuruman and overlying Griquatown Formations. At Sishen Iron Ore Mine, near the town of Kathu, the Asbestos Hills Subgroup comprises banded iron-formation, jaspillite, shale and

fig 3.5

siltstone, which are ferruginised in places forming irregular bodies of high- to low-grade iron ore (Hälbich *et al.*, 1993). South of the Griquatown fault zone the Asbestos Hills Subgroup rocks grade upwards into the mudstone, iron-formation, riebeckite and amphibolite of the Koegas Subgroup (Beukes, 1980).

The Gamagara Formation consists of shale, quartzites and conglomerates, with ferruginous shales and siltstones found at the Sishen Iron Ore Mine (Hälbich *et al.*, 1993). Debate, however, still exists as to whether the Formation belongs to the Transvaal Supergroup (e.g. De Villiers, 1967, SACS, 1980) or is the equivalent of the Mapedi Formation (e.g. Visser, 1944; Beukes and Smit, 1987). For the sub-Kalahari geological map, SACS (1980) and the most recent version of the 1:1000 000 scale geological map of South Africa (Keyser, 1997) were followed, which place the Gamagara Formation in the Transvaal Supergroup.

The diamictites of the Makganyene Formation, the overlying mafic lavas of the Ongeluk Formation and the manganiferous rocks of the Voëlwater Subgroup together constitute the Postmasburg Group. The Ongeluk and Makganyene Formations can be correlated with the Segwagwa Group of Botswana (according to Eriksson *et al.*, 1993), and for the sub-Kalahari geological map the Voëlwater Subgroup was also included in this correlation on the basis of lithological similarities between the upper part of the Segwagwa Group and the Voëlwater Subgroup (Carney *et al.*, 1994). The maximum age of the Segwagwa Group is constrained by a 2193 ± 20 Ma detrital zircon age (Mapeo *et al.*, 2004b,c). The nature of the Makganyene diamictite is still uncertain, although Visser (1971) and de Villiers and Visser (1977) consider it to be of glacial origin and that it was deposited on an unconformity which developed after the uplift and erosion of the Postmasburg Group (Visser, 1971; Beukes, 1986). This may be true for the Kanye Basin, since a distinct unconformity is present between the Segwagwa and underlying Taupone Groups (Eriksson *et al.*, 1995). In the Griqualand West Basin, however, such an unconformity is not observable (Moore *et al.*, 2001) and the Makganyene Formation overlies both the Koegas and Asbestos Hills Subgroups in what has been suggested as a conformable contact (Polteau and Moore, 1999; Moore *et al.*, 2001). The iron formations interbedded with basal Makganyene diamictites in the area between Sishen and Rooinekke in the west of the basin, are compositionally and isotopically similar to those of the Koegas and Asbestos Hills Formation (Polteau and Moore, 1999) and the Makganyene diamictites are transgressive across the Asbestos

Hills Subgroup in the east, on the flanks of the Vryburg arch (Moore *et al.*, 2001). On the basis of these relationships, Moore *et al.* (2001) suggest a continuous succession between the Ghaap and Postmasburg Groups in the Griqualand West Basin. At the time of writing, however, the stratigraphy shown on the sub-Kalahari map is recognised by the South African Committee for Stratigraphy as correct (M.R. Johnson, *pers. comm.*). The Makganyene Formation possibly provides evidence of a global glaciation event, a "snowball earth", with evidence of contemporaneous glaciation events having been recognised in central Canada, Wyoming and Finland (summarised in Hoffmann and Schrag, 2002).

The Ongeluk Formation was deposited unconformably over the Makganyene Formation (Altermann and Hälbich, 1991) and is characterised by massive andesites, hyaloclastics and pillow lavas (Grobler and Botha, 1976; Grobelaar *et al.*, 1995). At Wessels Manganese Mine to the northwest of Kuruman, the Ongeluk Formation comprises a thick basaltic lava package hosting prominent jaspillite horizons, lenticular altered banded iron-formation layers, pillow lavas and occasional tuffaceous layers (Wessels Manganese Mine, visitors guidebook, 1996). The Ongeluk Formation has a Pb-Pb age of 2222 ± 13 Ma (Cornell *et al.*, 1996), but Moore *et al.* (2001) suggest that this age may have been influenced by subsequent metamorphism or metasomatic processes and the lava is in fact older. The Pb-Pb age of 2394 ± 26 Ma for the Mooidraai Formation (Voëlwater Subgroup) carbonates (Bau *et al.*, 1999) provides evidence for this.

The overlying Hotazel Formation, part of the Voëlwater Subgroup, consists of volcanogenic sedimentary jaspilites with manganiferous layers (Beukes, 1980, 1984). At the Wessels Manganese Mine, it comprises three strataform manganese bodies within a banded ironstone succession (Wessels Manganese Mine, *pers. comm.*).

3.3.8 Cassinga Supergroup

The Cassinga Supergroup of southwest Angola (see sub-Kalahari geology map) consists of the basal Jamba Group, comprising amphibolites, greenstones, felsic lavas and tuffs, argillite, graywacke and banded hematite quartzite, and the Cuandja Group comprising only sandstone and various types of greenstone (Hood and Korpershoek, 1968). Although shown as Eo- to

Palaeoproterozoic in age on the sub-Kalahari geological map , the deposition of the Cassinga Supergroup spanned a much longer period, stretching from the Neoarchaean to about 1835 Ma (Rb/Sr age) in the Palaeoproterozoic era (Hartzer, 1998).

3.3.9 Lower Proterozoic intrusives

3.3.9.1 The Molopo Farms Complex

The Molopo Farms Complex is a layered intrusion of approximately 12000 km^2 in area covered by Kalahari Group sedimentary rocks along the border between Botswana and South Africa (Gould *et al.*, 1987)(see sub-Kalahari geological map, Figs 3.1,3.2). A Pb-Pb age of 2055 Ma (Hartzer, 1998) puts it at the same time zone as the more famous Bushveld Complex of South Africa.

As shown on the sub-Kalahari geological map, the complex is divided into the upper Complex consisting of norites and gabbros, and the lower Complex consisting of diorite, harzburgite and pyroxenite. In South Africa Wilhelm *et al.* (1988) subdivided the complex into four units and Gould *et al.* (1987) differentiated an Ultrabasic Series, Layered Basic Series and a Minor Intrusive Suite in Botswana. Basic dykes intruded the Waterberg Group sedimentary rocks which were deposited subsequent to the emplacement of the Minor Intrusive Suite (Carney *et al.*, 1994).

3.3.9.2 The Segwagwa Complex

The Segwagwa Complex consists of syenites, granites, hornblende-diorites, pyroxene-mica-diorites, gabbros and norites (Carney *et al.*, 1994). The Segwagwa granite has a U-Pb zircon crystallisation age of 2054 ± 9 Ma (Mapeo and Wingate, 2004).

3.3.9.3 The Kubu Island Granite

The Kubu Island Granite forms conspicuous 'islands' in the Makgadikgadi Pans of Botswana. It yielded a U-Pb zircon age of 2039 ± 1.4 Ma interpreted as the crystallisation age (Majaule *et al.*, 2001).

3.4 Assembly of supercontinents

It is believed that during the Proterozoic the collision of cratonic blocks occurred along several tectonic belts. Although the exact configuration of these blocks both prior to and at the time of collision is still debatable, the amalgamation of these blocks had a significant influence on the geological evolution of the region.

The supercontinent of Rodinia was created as a result of collision of cratonic blocks in the Mesoproterozoic. It is not clear if the continent of Rodinia included the amalgamated Congo-Tanzania-Bangweulu Craton, which was joined together by the Kibaran orogeny and was flanked by the Zambezi and Irumide Belts, or if this was a separate entity at this time (Johnson and Rivers, 2004).

3.4.1 The Okwa Complex, Kheis and Magondi Belts and the Kalahari Suture Zone

3.4.1.1 The Okwa Complex

The Okwa Complex covers a relatively small area on the sub-Kalahari geological map and is found just to the south of the Ghanzi Group rocks of Botswana, where it is bounded by the Tsau Fault Zone and the Kalahari Line. The Okwa Complex consists largely of porphyritic felsites, augen granites and gneisses, leucogranites, sericitic quartzites (Aldiss and Carney, 1992) and metadolerite (T.Majaule, *pers. comm.*). The nature and position of the Okwa Complex has led some researchers to believe that it and the Magondi Belt may originally have been laterally continuous (e.g. Stowe, 1986; Munyanyiwa and Maaskant, 1998; Ramokate *et al.*, 2000; Brett *et al.*, 2000) or that the Okwa complex granitic rocks indicate a continuation of the Kheis Belt (Key and Rundle, 1981). Dates of ± 2050 Ma for basal metamagmatic rocks, of 2055 ± 4 Ma for felsic rocks and 2101 ± 4 Ma from a xenocrystic zircon in a porphyritic rhyolitic felsite (Ramokate *et al.*,1996; Ramokate *et al.*, 2000) compare well with the ages of the Magondi Supergroup (see section 3.4.1.2). The latter age probably represents the older rocks into which the Okwa Complex igneous rocks intruded (Ramokate *et al.*, 2000). It is postulated that after emplacement and deposition, the Okwa Complex was tectonically transported eastwards and accreted onto the Kaapvaal Craton at ± 1813 Ma, during the same collision event thought to have imposed the north-south structural grain on the Kheis Belt (Carney *et al.*, 1994).

3.4.1.2 The Kheis Belt

The Kheis Belt or Kheis Tectonic Subprovince (Moen, 1999) is a zone of low-grade metasedimentary and metavolcanic rocks stretching from the Northern Cape Province of South Africa, northwards into Botswana, forming a tectono-metamorphic transition zone between the stable Kaapvaal Craton to the east and high grade metamorphic rocks of the Namaqua-Natal Belt to the southwest and west. The Kheis Belt mainly contains rocks of the Olifantshoek Supergroup, as well as rocks of the Brulpan, Vaalkoppies and Wilgenhoudtsdrif Groups.

The Olifantshoek Supergroup is divided into the basal Mapedi, Lucknow, Hartley, Matsap, and Brulsand Formations in South Africa. The metasedimentary rocks and andesites (Hartley Formation) of the Olifantshoek Supergroup were, according to Stowe (1990), deposited in rift related fluvial, deltaic and near shore tidal environments. The Hartley Formation dated at 1928 ± 4 Ma (Cornell *et al.*, 1998; Hartzer *et al.*, 1998), while an age of 1881 ± 56 Ma (Armstrong, 1987) possibly represents isotopic resetting during the Kheis orogeny (Master, 1991,1994).

The quartz-muscovite schists and quartzites (Moen, 1999) of the Groblershoop Formation, were formerly considered part of the Olifantshoek Supergroup (SACS, 1980) but lithological, stratigraphic and structural evidence suggested it should be separated (Moen, 1999), and it is now assigned to the Brulpan Group (M.R. Johnson, *pers. comm.*). The Groblershoop Formation has yielded an ^{40}Ar-^{39}Ar age of ± 1750 Ma (Master, 1994). The Olifantshoek Supergroup extends about 270km into Botswana where it is separated from the Waterberg Supergroup by a basement high (Meixner and Peart, 1984). The similarities between the Olifantshoek Supergroup and Waterberg Supergroup rocks, as well as their close proximity to each other, led DuToit (1926) to propose a correlation between them. It has, however, subsequently been decided to separate them (e.g. SACS, 1980), with the Waterberg Supergroup rocks appearing above the Olifantshoek Supergroup on the geological legend of the sub-Kalahari geological map.

The quartzites, schists, gneisses and migmatites of the Vaalkoppies Group are found to the east of Upington in South Africa. A maximum age for the Vaalkoppies Group is indicated by ages from detrital zircons from the high-grade paragneisses to the east of Upington. These dates of 1800-2100 Ma probably represent a minimum age for the precursor (Barton and Burger, 1983; Moen, 1999). A minimum age for the Vaalkoppies can be ascertained by intrusive granites of the Keimoes Suite (SACS,1980; Moen, 1999) and by dates from the schists, phyllites, quartzites and volcanic rocks of the overlying Wilgenhoudtsdrif Group, which gives a U-Pb (SHRIMP) age of 1290 ± 8 Ma (Hartzer *et al.*, 1998) and a depositional age of ~1330 Ma (Barton and Burger, 1983). The Vaalkoppies sequences are thought to represent back-arc or fore-arc deposits of an island arc that developed during Middle Proterozoic convergence prior to its accretion onto the Kaapvaal Craton (Moen, 1999; Gutzmer *et al.*, 2000).

3.4.1.3 The Magondi Belt

The early Proterozoic Magondi Belt (Fig. 3.4) is composed mainly of sedimentary as well as minor mafic and intermediate to felsic metavolcanic rocks of the Magondi Supergroup, which comprises the Deweras, Lomagundi and Piriwiri Groups, deposited between 2160 and 2000 Ma ago (Master, 1991;1994). Magmatic zircons from the syntectonic Hurungwe Granite, intrusive into the Piriwiri Group were dated and a U-Th-Pb (SHRIMP) concordia age of 1997.5 ± 2.6 Ma was obtained (McCourt *et al.*, 2001). This date is interpreted as the time of crystallisation and gives a minimum age for the Piriwiri Formation (McCourt *et al.*, 2001) as well as the main phase of deformation of the Magondi Orogeny (Treloar, 1988; McCourt *et al.*, 2001).

The tectonic setting of the Magondi Belt has been proposed as (1) developing in an ensialic geosyncline along the western margin of the Zimbabwe Craton (Leyshon and Tennick, 1988); as (2) depicting the transition of a passive-margin setting into geosynclinal flysch-type deposits (Stowe, 1989), as (3) deposition in a back-arc continental basin (as with the Kheis Belt) which resulted from subduction in an easterly-dipping direction (Master, 1991; 1994), and (4) initial sedimentation and volcanism related to northerly trending rifting of continental crust (Campbell *et al.*, 1991).

Magondi Supergroup rocks were initially deformed around 2000-1800 Ma (Stowe, 1986; Treloar and Kramers, 1989; Munyanyiwa and Maaskant, 1998), 1900-1700 Ma (Olson, 2000), or 1800-1700 Ma (Stowe *et al.*, 1984) and again at ~820 Ma (Loney, 1969; Munyanyiwa and Blenkinsop, 1993). The Magondi belt rocks underwent low-grade greenschist facies metamorphism in the south, middle to upper amphibolite facies in the north, and granulite grades in the extreme north of the belt (Munyanyiwa and Maaskant, 1998).

3.4.1.4 A possible correlation between the Magondi and Kheis Belts?

The southern continuation of the Magondi Belt is covered by Karoo Supergroup and Kalahari Group deposits in Botswana and western Zimbabwe, but it has been proposed that the Magondi Belt is laterally continuous with the Kheis Belt (Coward and Potgieter, 1983). The correlation between the Magondi and Kheis belts was based on a carbonate isotope anomaly in carbonates of the Lucknow Formation (Master *et al.*, 1993, in: Master, 1994) and the Magondi Belt (the "Lomagundi carbon isotope anomaly") (Schidlowski *et al.*, 1976; Master *et al.*, 1990). It was suggested by Master (1994) that if the Kheis and Magondi Belts are continuous and coeval, then the Kheis-Magondi Orogeny may have resulted from closure of continental back-arc basins because of the suturing of two Archaean-age continents. P-T-path data from the granulite facies terranes of the Magondi Belt suggests, however, that the granulites were not formed by continent-continent collision, but in a region of high heat flow, with the heat possibly being supplied by deep-seated plutons (Munyanyiwa and Maaskant, 1998).

The ~1997 Ma age of the syn-kinematic Hurungwe Granite (McCourt *et al.*, 2001), indicates that the main contractional deformation of the Magondi Orogeny was concluded prior to the extrusion of the Hartley basalts, ruling out the suggestion that the Magondi and Kheis Belts are part of the same orogen (McCourt *et al.*, 2001). It is possible, therefore, that the orogeny in the Kheis Belt was related to the Namaqua-Natal orogeny (eg. Armstrong, 2004). The similar ages of the Magondi Supergroup rocks and those of the Segwagwa Group of the Transvaal Supergroup led Mapeo *et al.* (2004c) to suggest that they should perhaps be combined into the same Supergroup and Eglington and Armstrong (2004) suggest a correlation can be made between the Magondi and Limpopo Belts based on the metamorphism occurring in both belts at around 2000 Ma.

Unfortunately, as the sub-Kalahari geology map shows, little is known about the rocks underlying the Karoo Supergroup sedimentary and volcanic rocks between the Kheis and the Magondi Belts, and rocks of the Magondi Belt are not shown on the map. A possible extension of the Magondi Belt towards the southwest may, however, be represented by the undifferentiated metamorphic rocks (A_2Mm) shown on the sub-Kalahari geological map to the west of the Makgadikgadi Basin. If these undifferentiated metamorphic rocks are indeed Magondi equivalents, they could be moved to a late Eoproterozoic or early Palaeoproterozoic position on the legend of the sub-Kalahari geological map.

3.4.1.4 The Kalahari Suture Zone

Adjacent to the Kheis Belt lies a prominent northerly trending magnetic feature of between 10-40 km in width, the Kalahari Line (Figs 3.1, 3.4). The Kalahari Line extends north from the Northern Cape Province of South Africa into southern Botswana but its northerly extent is unclear, and from central Botswana the Kalahari Line may swing northeastwards where it becomes known as the Makgadikgadi Line. Geophysical evidence suggests, however, that the Kalahari Line may branch to the north with a NNW-trending branch extending from the Makgadikgadi Line up through an area of exposed basement and a north-south branch continuing up as far as Angola (Corner, 2003; B. Corner, *pers. comm.*). The Makgadikgadi Line extends around the Okwa basement block and as far east as the Magondi Belt of Zimbabwe, and defines the southeastern edge of the Ghanzi-Chobe Belt (or Koras-Sinclair-Ghanzi Rift)(see 3.3.3). Collectively, the Kalahari and Makgadikgadi Lines are known as the Kalahari Suture Zone (KSZ) (Reeves, 1978c; Mason,1998) and are shown schematically in Figure 3.4.

The Kalahari Line is believed to be a tectonic front marking the western edge of the Kaapvaal Craton, separating a zone of deeply buried, highly magnetic basement to the west from shallow, less magnetic features to the east (Reeves, 1979; Hutchins and Reeves, 1980; Reeves and Hutchins, 1982).

The Makgadikgadi Line, with its northeasterly trend probably marks the northern edge of the Kaapvaal craton (see section 3.2), where it separates the relatively stable cratonic areas to its south from the more recently tectonised rocks (e.g. Ghanzi-Chobe Belt) to its north and northwest. This is reflected in the differing aeromagnetic and gravity signatures (Carney *et al.*, 1994; Figs 3.1, 3.2), and is seen as a major front of Proterozoic metamorphism and tectonism (Hutchins and Reeves, 1980; Reeves and Hutchins, 1982).

The KSZ is thought to have been a major thrust zone associated with the orogenesis in the Kheis and Magondi Belts (Mason, 1998; Key and Mapeo, 1999) and Hutchins and Reeves (1980) believe that tectonic activity associated with this thrust zone accreted Kheis schists against the edge of the Kalahari Craton. To the west of the Kalahari Line, in an area called the Nosop Basin, lies an area of deeply buried strongly magnetic basement (Reeves, 1979) and interactive forward modelling in the basin revealed a deeply buried slab characterised by alternately normal and reversed magnetisation. The magnetic basement is interpreted as representing oceanic crust with the magnetic stripes indicating ocean floor spreading, with the Makgadikgadi Line coinciding with the spreading centre (Zhou, 1988). If the interpretation by Zhou (1988) is correct, then the presence of ocean floor immediately west of the Kalahari Line discounts the possible extension of the Kaapvaal Craton in this direction. The obduction of oceanic crust (ophiolite) colliding and being overthrust onto the continental craton to the east is thought to have occurred at about 1700 Ma (Reeves, 1978 b,c; Zhou, 1988). The KSZ is thought to have been subsequently reactivated as a rift zone during the late Mesoproterozoic and Neoproterozoic with down throw to the northwest (Jones, 1979; Mason, 1998), and the presence of the Passarge (see 3.5.4) and the Nosop Basins in-filled with great thicknesses (over 10 km) of Neoproterozoic Ghanzi and Nama Group sedimentary rock suggests that the down throw was significant (Key and Mapeo, 1999). This explanation for the formation of the Passarge Basin has, however, been disputed by Pretorius (1979, 1984) as discussed in section 3.5.4.

Several anomalous magnetic and gravity features along the KSZ have been identified and described, with the Tshane and Xade Complexes (Fig. 3.1) being the most prominent. Although these complexes are covered by Karoo Supergroup rocks and are not shown on the Sub-Kalahari map, they are thought to be an important part of the sub-Kalahari geology of the area. The Tshane Complex is between 10-40 km wide and over 300 km in length and has been dated at ±1000-1100 Ma (Key and Ayres, 2000). The Tshane Complex was interpreted as an ophiolite sheet obducted onto a continental block at a convergent plate boundary (Reeves and Hutchins, 1982), or as a series of basic or ultrabasic intrusives emplaced along the Kalahari Line at various stages (Reeves, 1978c). A borehole drilled into the complex intersected a gabbro at 774m below the surface (Meixner and Peart, 1984). The Tshane Complex was divided into four sections based on aeromagnetic interpretation by Brett *et al.* (2000). The northern section can be divided into a north-striking easterly body, and a northeast-striking westerly body. Both of these bodies lie in a northeast-trending graben which may contain over 1000m of Karoo sedimentary and volcanic rocks (Brett *et al.*, 2000). The main section of the Tshane Complex extends from 23°15'S to 25°S and has been modelled as a continuous vertical intrusive body, offset by northwest and northeast-trending faults. The third section from 25°S to 26°20'S consists of three discrete, north-south elongated bodies, while the southernmost section (from 26°20'S to 27°S) consists mainly of two north-south elongated bodies (Brett *et al.*, 2000).

Intruded along the Makgadikgadi Line is the Xade Complex, and boreholes drilled into the complex as part of the Kalahari Drilling Project suggests that it is a layered basic/ultrabasic complex resulting from a series of intrusions (Meixner and Peart, 1984), of Mid-late Proterozoic in age, and possibly containing basic intrusions of Karoo age (Meixner and Peart, 1984; Carney *et al.*, 1994). The Xade Complex has a distinctive "Y" shape, with a southern arm parallel to the Kalahari Line, the northeast-trending arm following the Makgadikgadi Line, and the northwest-trending arm following a trend similar to the faulting in the Okwa Complex (Meixner and Peart, 1984).

3.4.2 Waterberg Supergroup

The Waterberg Supergroup consists of purple to khaki sandstones and quartzites with some shale and conglomerate layers. In Botswana it occurs to the south of the Zoetfontein Fault in a west-trending belt that continues into South Africa. Faulting influenced the deposition of Waterberg sediments in Botswana, with the thickest, or more complete successions occurring in fault-bounded troughs, with syn-depositional vertical movements along the faults suggested by occasional layers of coarser sediments (Crockett and Jones, 1975; Green *et al.*, 1980).

A maximum age for the Waterberg Supergroup rocks in Botswana is provided by the U-Pb crystallisation age of 2054 ± 2 Ma for the unconformably underlying Moshaneng Complex and a minimum age for the basal Mannyelanong Formation is provided by a U-Pb age of ~1927 Ma from an intrusive dolerite (Mapeo *et al.*, 2004d).

3.4.3 Palapye Group

The Palapye Group comprises largely clastic sedimentary rocks outcropping in eastern Botswana, which are probably the equivalents of the Soutpansberg Group of South Africa (Cheney *et al.*, 1990; Carney *et al.*, 1994). The Sibisa Formation of the Soutpansberg Group has an age of 1769 ± 17 Ma (Barton, 1979) and 1750 ± 100 Ma (Hartzer *et al.*, 1998), although these ages are disputed by Cheney *et al.* (1990) who favour an older age for the Soutpansberg Group. A ~1600 Ma age has been yielded by zircons retrieved from tuffs near the top of the Palapye Group in Botswana (Mapeo *et al.*, 2000a).

3.4.4 Khoabendus Group

The Khoabendus Group of Namibia consists of tuffs, metamorphosed basaltic to rhyolitic lavas, volcanoclastic material, quartzites, and dolomitic and carbonaceous mudstones/siltstones (Miller, 1992b; Steven and Armstrong, 2002). Northwest of Kamanjab, the Khoabendus Group contains a > 50m thick banded iron formation that extends over 60 km (Miller, 1992c). The Khoabendus Group possibly formed in an ocean arc or active continental margin setting (Miller, 1992c). The lavas were initially dated at 1860 -1760 Ma by Burger and Coertze (1975) but zircons from a rhyolite yielded a more precise SHRIMP U-Pb age of 1862 ± 6 Ma (Steven and Armstrong, 2002).

3.4.5 The Kibaran Belt

The Kibaran Belt separates the Congo craton from the Tanzanian craton and trends northwest over a distance of some 700 km. The southwestern portion of the Kibaran Belt is obscured by a thick cover of Karoo, Cretaceous and Kalahari sedimentary and volcanic material, and the lack of boreholes and detailed geophysical data from the southeastern part of Angola makes interpretations difficult.

The usage of the term Kibaran will be limited to the Kibaran Belt only, and not as a synonymous term for the Mesoproterozoic erathem. This is in line with the decision made in IGCP 418 (Key and Mapeo, 1999).

The earliest sediments deposited were quartzites and phyllites, with some volcanic material also being extruded (Cahen *et al.*, 1984). In the Eastern Domain of the Northeastern Kibaran Belt the age of basin formation and the onset of volcanism has been constrained by a tuff which yielded magmatic age of 1.78 Ga (Cutten *et al.*, 2004). Magmatic zircons of around 1.41 Ga from the Western Domain of the Northeastern Kibaran Belt may indicate either that sedimentation began much later than that in the Eastern Domain, or that sedimentation began at around 1.8 Ga and continued until at least 1.41 Ga (Cutten *et al.*, 2004). Following the initiation of a subduction stage starting at around 1400 Ma (Kokonyangi *et al.*, 2002), various granites were intruded, with an age of ~1375 Ma (SHRIMP) coming from the northeastern end of the belt (Tack *et al.*, 1999), and other U-Pb SHRIMP ages of ~1372-1386 Ma from Mitwaba in the Democratic Republic of the Congo (Kokonyangi *et al.*, 2001; 2002). Recent ages suggest all granitoids may, however, have been emplaced during a short period around 1381 ± 8 Ma (Kokonyangi *et al.*, 2004). The compressional event lasted until about 1180-1100 Ma (Olson, 2000; Kokonyangi *et al.*,2002) when continental collision occurred, probably between the Congo and East African (Tanzania-Bangweulu) Cratons (Kampunza, 2001). Samples taken from the pre-Katangan basement in the Domes area of Zambia yielded zircon U-Pb ages of 1400 to 1200 Ma and a whole rock Rb-Sr age of 1220 Ma (Cosi *et al.*, 1992), and these ages were interpreted as indicating the age of metamorphism of the basement rocks.

The southwestern extent of the Kibaran Belt is unclear. The Red Granites of southwestern Angola (see sub-Kalahari geological map) provided Rb/Sr ages of between 1411 and 1302 Ma (Hartzer, 1998) and Olson (2000) suggests that the southern Angola granitoids can be related to a period of increased magmatism because of a southwestward extension of the Kibaran Rift System, as can the presence of the Cunene Complex (see following section). SHRIMP ages of 1234 -1019 Ma from northwestern Botswana were also interpreted as evidence of an extension of the Kibaran Belt of central Africa all the way down into Botswana (Kampunza *et al.*, 1999), but these ages could rather be related to the Irumide Orogeny (see section 3.4.8). Key *et al.* (2001) suggest that the Lufilian Arc is in direct contact with the Congo Craton in northwestern Zambia, which does not leave room for any extension of the Kibaran Belt into Zambia. The presence of rocks of a similar age to the south and southwest does not provide sufficient grounds for suggesting an extension of the Kibaran Belt in this direction. Until compelling evidence of an extension of the Kibaran Belt into Zambia and Angola is found, it should be assumed that it does not extend beyond the Democratic Republic of Congo (Hanson *et al.*, 1988a; Carney *et al.*, 1994).

3.4.6 The Cunene (Kunene) Complex

The Cunene Complex occurs in southern Angola and northern Namibia (see sub-Kalahari geological map), and is one of the largest massif-type gabbro-anorthosite bodies in the world

(Ashwal and Twist, 1994). It covers about 15000 km^2 (Morais *et al.*, 1998), and consists of a succession some 15000 m thick, comprising 25 subzones (Menge, 1998). The main lithologies of the Complex include anorthosite, troctolite, marginal noritic anorthosite, anorthositic gneiss, syenite and titaniferous magnetite.

K-Ar dates from an anorthosite in the Cunene Complex yielded ages of 2098 ± 51 and 2151 ± 43 Ma (Silva *et al.*, 1973), and subsequent isotopic ages of between 2700 and 2160 Ma (Carvalho and Alves, 1993) have been recorded. Menge (1998) believes that the Cunene Complex has a maximum age of 1500 Ma, however, and Nd isotopic data suggests a Mesoproterozoic age for the emplacement of the Complex (Morais *et al.*, 1997). Ages of 1470-1408 Ma are reported for a gneiss intruded by the Cunene Complex in northern Namibia (de Carvalho *et al.*, 2000). A U-Pb single zircon age of 1385 ± 25 Ma was retrieved from the Namibian part of the complex (Druppel *et al.*, 2000), a zircon from a vein in the Anorthosite gave an age of 1370 ± 4 Ma and an intrusive mangerite dyke gives concordant ages of 1371 ± 2.5 Ma (Mayer *et al.*, 2004) and 1385 ± 7.6 Ma (McCourt *et al.*, 2004b) with the latter age giving a minimum age for the complex.

The variety of ages retrieved from the Cunene Complex probably indicates the emplacement of the rocks over a period of time, as suggested by Ashwall and Twist (1994) who interpret the Cunene as a composite, massif-type intrusive complex.

3.4.6 The Namaqua-Natal accretion

The Namaqua Orogeny affected rocks of the Korannaland Supergroup and Areachap Group (as well as the rocks of the Kheis Belt), and caused the intrusion of various granitoids along the southern edge of what is now the Kalahari Basin. In southern Namibia metasedimentary and igneous rocks were reworked by the Namaqualand tectonothermal event of ± 1200-900 Ma (Miller, 1992d). The Namaqualand metasediments, gneisses and granitoids shown on the Sub-Kalahari Geology map in Namibia are thought to belong to the Gordonia Subprovince of the

Namaqua Province (e.g. Joubert, 1986) and have been grouped together following the 1: 1000 000 scale Geological map of Namibia (Geological Survey of SWA/Namibia, 1980). In South Africa rocks of the Eendoorn Suite were emplaced during this tectonic event (SACS, 1980).

The Korannaland Supergroup is represented on the Sub-Kalahari map by the Biesje Poort Formation, consisting of calc-silicate rocks with layers of marble, amphibolite and granulite (SACS, 1980). The Korannaland Supergroup rocks were deposited in a basin on the southern edge of the Kaapvaal Craton at the same time as the amphibolites, calc-silicates, gneisses and calc-silicate and pelitic schists of the Areachap Group (SACS, 1980; Geringer *et al.*, 1994) were being formed in a volcanic arc setting (Geringer *et al.*, 1994). A minimum age for the formation of the Areachap Group comes from a late tectonic quartz diorite which intruded into the amphibolites of the Areachap Group (Cilliers, 1987), and ages of 1285 ± 14 Ma for volcanic rocks (Cornell *et al.*, 1990a) and Pb-Pb ages of 1660-1350 Ma for the formation of amphibolites have been obtained (Theart, 1985, in: Geringer *et al.*, 1994; Cilliers, 1987). Amphibolite-facies regional metamorphism is thought to have affected the Areachap Group at about 1210 Ma, probably caused by the pre-Namaqua orogenic subduction of oceanic crust underneath the Kaapvaal Craton (Geringer *et al.*, 1988; Cornell *et al.*, 1990b). At around 1200 Ma a collision stage started and calc-alkaline granites of the Keimoes Suite were intruded into the Korannaland Supergroup and Areachap group (Stowe, 1986).

The Namaqua-Natal belt is characterised by thrusting followed by transcurrent shearing (Dalziel *et al.*, 2000), and has been interpreted as part of a collision zone between the Laurentian and Kalahari Cratons between 1150 and 950 Ma (Dalziel *et al.*, 2000), with R.J. Thomas *et al.* (1993, 1994) putting the peak of collisional orogenesis at ~1150 Ma.

3.4.7 The Quangwadum Group and the Grootfontein, Huab and Kwando Complexes

Granites and gneisses in the Damara Belt of northwestern Botswana are referred to as the Quangwadum Group (Botswana Geological Survey, *pers. comm.*), or Quangwadum Complex (Carney *et al.*, 1994). The Quangwadum Group rocks appear to be laterally continuous with the granites and amphibolites of the Grootfontein Complex of Namibia (see sub-Kalahari geological map). The Grootfontein complex is thought to be Palaeoproterozoic in age (Hartzer *et al.*, 1998; Hoal *et al.*, 2000), and dates from the Quangwadum Group include an age of 2050 Ma for an augen gneiss, which was intruded by granites at about 1020-1000 Ma (Singletary *et al.*, 2003). The relative proportions of the Palaeoproterozoic and the younger granites in the Quangwadum Group are not known (Singletary *et al.*, 2003).

The Huab Complex of Namibia, to the west of the Grootfontein Complex is considered to be of similar age to the Grootfontein and Quangwadum complexes (F.J. Hartzer, *pers. comm.*; Miller, 1992b). The complex consists of metabasite dykes, sills and stocks, leucocratic granitic gneiss, banded paragneiss, quartzite, schist, amphibolite, metavolcanic rocks, conglomerate, and orthogneiss (Miller, 1992b).

The Kwando Complex is the name given to a geophysically distinct area in northwestern Botswana (Carney *et al.*, 1994). A borehole drilled into this feature intersected granite gneisses, that grade down into layers of granite and with an intrusive contact against migmatitic amphibolite and biotite-hornblende schist (Singletary *et al.*, 2003). $^{207}Pb/^{206}Pb$ zircon ages from the borehole suggest that emplacement of the Kwando Complex occurred at 1200-1150 Ma (Singletary *et al.*, 2003).

3.4.8 The Irumide Belt

The Irumide Belt may once have been a continuous belt, stretching from Namibia (Ghanzi-Chobe Belt), through southern Zambia and northern Malawi into southern Tanzania, although Johnson and Rivers (2004) believe that there is no correlation between the Irumide Belt and the Choma-Kalomo Block on the southwestern side of the Zambezi Belt. It consists of granites and gneisses, as well as metasedimentary rocks (Ackermann and Forster, 1960; de Waele *et al.*, 2000, 2002).

According to de Waele *et al.* (2000) the sedimentary rocks of the Muva Supergroup were deposited in what appears to have been an unstable intra cratonic basin. A detrital zircon age of

~1941 Ma was recovered from a Muva Supergroup quartzite (Rainaud *et al.*, 2002), and de Waele and Fitzsimons (2004) report that lavas near the base of the Mporokoso Group give an age of 1860 Ma and basalts and rhyolites within the Manshya River Group give an age of 1879-1856 Ma. From the present outcrop appearance it seems as if the basin was narrow in the northwest, where initial rifting and extrusion of bimodal volcanics occurred. Towards the southeast, the basin deepened and widened, with the thick succession of volcanic rocks found here extruded in an oceanic environment. The presence of oceanic crust is substantiated by an ophiolite complex in the Zambezi Valley of Zimbabwe with an age of 1393 ± 22 Ma (Oliver *et al.*, 1998). This might be the equivalent in age and genesis to the Mpanshya Zone of Zambia (Oliver *et al.*, 1998), which according to Daly (1986) represents an Irumide Suture Zone. The ophiolite complex is thought to have formed along an oceanic convergent plate margin (Johnson and Oliver, 2000), suggesting that the Irumide Belt cannot be ensialic (Oliver *et al.*, 1998).

On structural grounds the Irumide Belt is subdivided into northern, central and southern regions (de Waele *et al.*, 2000). The age of the Irumide Belt is poorly constrained, with collision between the Congo Craton and an unknown terrane around 1100 Ma being most probable (Dirks *et al.*, 1999). NW-directed thrusting was accompanied by intrusion of granites. The granites were dated with the SHRIMP at 1020 Ma (Tembo and de Waele, 2001) as well as between 1038 ± 10 Ma and 1018 ± 6 Ma (de Waele *et al.*, 2002) and 1050-970 Ma (de Waele *et al.*, 2004) constraining the magmatic event. Peak metamorphism in the belt probably occurred around 1000 Ma (de Waele *et al.*, 2002).

3.4.9 The Koras-Sinclair-Ghanzi Rift

The Koras Group and Sinclair Supergroup rocks are believed to be contemporaneous depositions (e.g. SACS, 1980). Through a window in younger Nama Group and Karoo Supergroup rocks in southern Namibia, possible correlatives of Sinclair and Koras rocks have been recognised by Borg (1988) (Fig. 3.6). On the sub-Kalahari geological map the Sinclair Supergroup is represented by the Marienhof and Nückopf Formations. The Nückopf Formation consists of acid, basic and pyroclastic volcanic rocks, as well as quartzite and conglomerate with an age of about 1210 Ma (Hartzer, 1998). The overlying Marienhof Formation is about 1170 Ma old (Hartzer, 1998) and comprises quartzitic phyllite, conglomerate and basalt (SACS, 1980; Miller, 1992b).

The Koras Group rocks are shown on the southern edge of the sub-Kalahari geological map, where they are exposed to the north of the Orange River. The Koras Group consists of sandstone, quartz porphyry and basalt (SACS, 1980) and has U-Pb zircon (SHRIMP) ages of 1123 ± 12 Ma (Hartzer *et al.*, 1998), and 1171 ± 7 Ma for a quartz porphyry near its base (Gutzmer *et al.*, 2000). The Blauwbosch Granite, Betadam Gabbronorite and Rooiputs Granophyre are considered as equivalent to the Koras volcanism (SACS, 1980), since the Blauwbosch Granite was dated at 1054 Ma (Geringer and Botha, 1976). In Namibia, the Gamsberg Granite provided an age of 1078 ± 30 Ma (Burger and Coertze, 1973-74) providing the timing to the rifting associated with the Sinclair Supergroup deposition.

The Koras-Sinclair-Ghanzi succession defines a rift which can be identified by a number of basins (interpreted as grabens and yoked basins) that developed in two linear zones which parallel the southeast as well as the northeast margins of the Kaapvaal and Zimbabwe Cratons respectively (Borg, 1988)(Figs 3.6, 3.7). Borg (1988) believes that the northeast branch of the rift may represent the third arm of a triple junction, and can be regarded as a failed proto-rift separating the Congo from the Kalahari Craton. The rift-basins were filled with thick sequences of both sedimentary and volcanic rocks, which include rocks of the Koras Group (Koras Basin), Sinclair Supergroup (Sinclair Basin), and the Ghanzi Group (Ghanzi/Lake Ngami Basin)(Borg,1988), and the southeastern part of the rift is covered by the younger, largely undeformed sediments of the Neoproterozoic-Lower Palaeozoic age Nama Group (Key and Mapeo, 1999). Radiometric data suggests a younging of volcanic rocks and their contemporaneous intrusions along the rift from

south to north and northeastwards, which is attributed to rift propagation along older zones of crustal weakness (Fig. 3.7)(Borg,1988). The block faults, which initiated the volcano-sedimentary deposits in the grabens, intensified during graben evolution, and probably were reactivated locally during the Damara orogeny (Borg,1988).

According to Aldiss and Carney (1992) there was rifting in Botswana in the Mesoproterozoic, with rifting along the northeast trending Botswana Rift (Key and Ayres, 2000) continuing into the Neoproterozoic (950-850) (Modie, 1996). The intra-continental rifting and bimodal volcanism were accompanied by the accumulation of sandstones, volcanoclastics and tuffitic siltstones (Modie, 1995,1996,1998,2000; Modie *et al.*, 1998) which led to the development of the Kgwebe Formation. According to Kampunza *et al.* (1998) the volcanic rocks were extruded late during the orogenic collision and the associated extensional collapse phases, with the Kgwebe metarhyolites produced by melting of Mesoproterozoic calcalkaline rocks underplated in the middle and/or lower crust. A U-Pb zircon age from a Kgwebe Formation rhyolite in the Mabeleapodi Hills dated the volcanism in the KSG rift at about 1106 ± 2 Ma (Schwartz *et al.*, 1996), with an almost identical crystallisation age of 1106.2 ± 3.6 Ma being obtained recently from zircon analyses of a rhyolite lava further to the northeast (Singletary *et al.*, 2003). The Goha Hills Formation of Botswana is lithologically similar to the Kgwebe Formation, (Carney *et al.*, 1994; Key and Ayres, 2000), suggesting that they can be correlated (Key and Ayres, 2000).

The Ghanzi Group is correlated with the Nosib Group of Namibia and the Chinamba Hills Formation of Botswana (Key and Ayres, 2000) and is included in the Damara Supergroup. This means that if the Ghanzi Group was deposited in the KSG rift of Borg (1988), it was part of a different tectonic episode to the rest of the Damara Supergroup, which was only deposited when Rodinia began to break apart.

Deposition of the lower Ghanzi Group is thought to have occurred as a result of rifting and uplift and erosion of surrounding areas, followed by thermal sag and the development of a shallow marine basin into which the middle and upper Ghanzi Group beds were deposited (Modie, 1995). Ion microprobe U-Pb dates obtained from detrital zircons, extracted from sediments in the northern part of the rift revealed ages of 1104 ±16 Ma, 1363 ±1 Ma, and 1748 ±13 Ma (Kampunza *et al.*, 2000). These sediments were interpreted by Kampunza *et al.* (2000) as belonging to the Ghanzi Group, but it is possible that these sediments may, however, represent a sedimentary sequence within the Kgwebe Formation as suggested by Meixner and Peart (1984). If the rocks do indeed belong to the Ghanzi Group, the date of 1104 ±16 Ma may reflect the Kgwebe Formation as being the primary source of Ghanzi Group sedimentary rocks, while the decreasing content of feldspar and rock fragments of volcanic origin from the lower to upper Ghanzi Group was interpreted as showing that subsequent to the extrusion of the Kgwebe lava, there were no further magmatic events that have provided fresh material to the sedimentary deposits (Akanyang and Schwartz, 1994). The age of the Kgwebe volcanic rocks therefore sets a maximum age for the Ghanzi Group. Subsequent to the Kgwebe volcanism, a large sedimentary basin with up to 5000 m of Ghanzi Group sediments developed (Modie, 2000). The other ages of 1363 ±1 Ma, and 1748 ±13 Ma are thought to represent source rocks belonging to the Mesoproterozoic rocks and Palaeoproterozoic granitoids in Namibia and southern Angola respectively. Sedimentary rocks of the Okwa Group, unconformably overlying the Ghanzi Group rocks, yielded a detrital zircon U-Pb age of 579 ±12 Ma (Ramokate *et al.*, 2000) and provide a minimum age for the deposition of the Ghanzi (and adjacent Nosib Groups). A hiatus of at least 200 Ma between the Okwa and Ghanzi rocks is postulated by Ramokate *et al.* (2000), putting the deposition of the Ghanzi Group prior to 750 Ma. If photo geological interpretations by Akanyong and Schwartz (1994), suggesting that there is no unconformity present between the Kgwebe and Ghanzi units, are correct, it could put the depositional age of the Ghanzi and adjacent Nosib Group at around 1100 Ma. Sandstones, quartzites, siltstones and conglomerates of the Ghanzi (Botswana) and Nosib (Namibia) Groups are laterally continuous across the border, and have been correlated following work by various researchers (e.g. Litherland, 1982; Germs, 1995).

There is unfortunately little age control over the Ghanzi Group sediments themselves (B. Modie, *pers. comm.*) and if correlated with the Nosib Group sedimentary rocks of northern Namibia (about 800 Ma in age), it means that they were not deposited in the same rift structures as the Koras and Sinclair rocks as proposed by Borg (1988) and instead may be related to a later rifting event that resulted in the break up of Rodinia.

Borg (1988) proposed that the migration of the African Plate over a mantle plume or “hotspot” between 1050-950 Ma led to a “rift-jump” from the KSG Rift to the early Damara Rift, although new ages for early Damara rifting of ~800 Ma (de Kock *et al.*, 2000) means that an age gap of some 150 Ma make a movement of the hotspot unlikely.

3.5 The break-up of Rodinia and the Pan-African event

Rocks of the Damara, West Congolian and Katanga Supergroups (Fig. 3.8) were deposited over large areas of Rodinia in the Neoproterozoic, as a result of the continental stretching and subsequent rifting related to the break up of Rodinia. By the end of the Neoproterozoic Rodinia had been broken up and largely reconfigured as the new supercontinent Gondwana (Key, 2002) with collision between the Congo and Kalahari Cratons occurring in the Neoproterozoic, final suturing between the Australian and Antarctic segments occurring in the Lower to Middle Cambrian (Meert and van der Voo, 1996) and Middle Cambrian collision recognised in southeastern Brazil (Schmitt *et al.*, 2004). The *Pan African* event that resulted in the assembly of Gondwana saw the deformation of the rocks deposited over Rodinia in several well-defined orogenic belts.

3.5.1 The Lufilian and Zambezi Belts

Katanga Supergroup sedimentary rocks were deposited unconformably over the rocks of the Muva Supergroup in Zambia following the Irumide Orogeny (see section 3.4.7). The distribution of the Katanga Supergroup rocks is shown on figure 3.8 and on the sub-Kalahari geological map. The early rifting associated with deposition of the early Katanga sediments was thought to be rapid (Unrug, 1988) and related to the fragmentation of Rodinia at about 880-850 Ma (Porada and Berhorst, 2000). The first Katanga formation to be deposited, the Lower Roan Formation, comprises basal conglomerates and aeolian sandstones, followed by siliciclastic material, and finally by shales, dolostones and arenites (Binda, 1994), typical of a continental rift setting which later evolved to a proto-oceanic stage (Kampunza *et al.*, 1991; Kampunza and Cailteux, 1999). The succeeding Upper Roan Formation, which hosts the Mine Series, is predominantly dolomite and shale (Binda, 1994). Both the Roan Group and the uppermost unit of the Katanga in the Kalahari Basin region, the Kundelungu Group, continue into Angola, where they are known as the Malombe and Macondo Groups respectively (Instituto Geologica de Angola, *pers. comm.*)

and the lower Katangan sedimentary series can be correlated with similar sequences in the Democratic Republic of Congo (Cailteux *et al.*, 1994). The Lunda and Luane Group conglomerates, sandstones, quartzites and schists were possibly deposited at this time in northern Angola, although they have been placed in the Mesoproterozoic on the Sub-Kalahari Geology map following Hartzer (1998). The rocks of the Zambezi Belt were also deposited in a rift setting, either ensialic in origin, or an intra continental rift basin transecting continental crust (e.g. Porada, 1989), or involving an oceanic basin (e.g. Coward and Daly, 1984).

The age of the Katanga Supergroup rocks had been constrained between ~1200 and 870 Ma (Cahen *et al.*, 1984), but SHRIMP U-Pb zircon age of 877± 11Ma (Armstrong, *et al.*, 1999) from a basal source rock (Binda, 1972), the Nchanga Granite, show the sedimentary rocks to be younger. The Nchanga Granite is unconformably overlain by the Roan Group quartzites, argillites, conglomerates and carbonates (Binda, 1972, 1994), which yielded detrital zircon populations aged ~880 Ma and 1800-2000 Ma (Armstrong *et al.*, 1999). The first age population is interpreted by Armstrong *et al.* (1999) as derived from the Nchanga Granite providing material for the lowermost Roan Group sedimentary rocks, while the latter age population indicates a source area in the older Ubendian basement. Microcline veins cutting the sedimentary rocks were dated at 870 ± 42 Ma (Cahen *et al.*, 1984) and indicate that Roan units were deposited between about 880 and 830 Ma (Armstrong *et al.*, 1999). This is similar to the Zambezi Belt, where crustal extension occurred from about 880 Ma, when the Kafue Rhyolites were deposited (Hanson *et al.*, 1994), until at least ± 804 Ma when the Rushinga Igneous Complex was emplaced (Porada and Berhorst, 2000). An age of ~765 Ma for the overlying Mwaisha Group lavas in the Katangan strata provided a minimum age for the Roan Group in northwest Zambia (Key *et al.*, 2001). A minimum age for the Katanga Supergroup was constrained to 570 Ma based on isotopic studies of the intrusive Hook granite (Hanson *et al.*, 1993), and by U-Pb dating of post-Kundelungu Group uraninite veins which gave an age of 656 Ma (Armstrong *et al.*, 1999),

The Zambezi and Lufilian Belts developed during periods of deformation in the Neoproterozoic to Palaeozoic, during which the Katangan rocks and underlying basement were deformed together (Coward and Daly, 1984). The Zambezi Belt is an east to northwest trending belt of intermediate to high-grade metamorphic rocks extending from a triple-junction formed by a convergence with the north-south trending Mozambique Belt. The belt forms the northern boundary to the Zimbabwe Craton, and in northwestern Zimbabwe, the rocks of the Makuti Group of the Zambezi Belt overlie the rocks of the Magondi Belt (Munyanyiwa and Blenkinsop, 1993). The Lufilian Belt (or Arc) is bordered on its northwestern side by the Kibaran Belt (Unrug, 1983), while its eastern border is formed by the Irumide Belt (Ackermann and Forster, 1960). While the western margin of the Lufilian Belt is obscured under Karoo Supergroup rocks, it has been proposed that it was formed by a transcurrent fault zone that evolved from a transform fault separating the Damara and the Lufilian-Zambezi sectors during the spreading phase of the Pan-African mobile zone (Unrug, 1983). The presence of such a structure could not be determined from the limited geophysical data available for western Zambia however.

The Zambezi Belt is separated from the Lufilian Belt by the ~550 Ma Mwembeshi Dislocation/ Fault/ Suture Zone (De Swardt *et al.*, 1965; Hanson *et al.*, 1994, 1998). The zone separates marked changes in the character of structures, with ENE oriented structures of a lower metamorphic grade occurring to the north of the zone, and high metamorphic grade, deep crustal rocks occurring in SW and SSW facing structures to the south (Coward and Daly, 1984). The zone does not, however, mark a huge change in sedimentology between the rocks of the Lufilian and those of the Zambezi Belts (Coward and Daly, 1984).

For the Lufilian Belt Kampunza and Cailteux (1999) proposed three phases of deformation with the first phase (which they correlated with the main deformation in the Zambezi Belt) occurring between 800 and 710 Ma. The second phase of deformation resulted in a clockwise rotation of the eastern part of the belt, giving the arc its distinctive convex geometry (Fig. 3.4), and the third phase produced the structures younger than 540 Ma that lie transverse to the main structural trend of the Lufilian Belt. Armstrong *et al.* (1999) suggested that the Katangan Supergroup was deformed and metamorphosed during two orogenic events, the Lusakan (~840 Ma) and Lufilian (~ 650 Ma) orogenies. The Lusakan event seems to be the time-equivalent of the metamorphism and deformation in the Zambezi Belt based on various intrusive bodies, of which a syntectonic granite gneiss body in Zambia provided a best constrained U-Pb zircon age of 820 ± 7 Ma (Hanson *et al.*, 1988b). The existence of the Lusakan Orogeny has, however, been disputed by Porada and Berhorst (2000) who believe that the Lufilian Belt was formed in a single orogenic event, bracketed between 645 and 515 Ma. According to Porada and Berhorst (2000), some of the previous evidence for a Lusakan orogeny (e.g. pegmatite veins dated by Cahen *et al.*, 1984) can now be discounted in the light of subsequent age determinations (e.g. Richards *et al.*, 1988) and reevaluation of field relationships.

Cosi *et al.* (1992) interpreted Rb-Sr, K-Ar, and U-Pb ages from the Domes area of Zambia as indicating that the highest grade of metamorphism occurred at ± 700Ma. Porada and Berhorst (2000) cite, however, the ages of ± 550 Ma from the Hook Granite, which is intrusive into Katanga Supergroup strata north of the Mwembeshi Suture Zone, and those on related syntectonic rhyolites in the Mwembeshi Suture Zone (Hanson *et al.*, 1994) as indicating the main stage of deformation in the region. It is believed that tectogenesis in the Lufilian and Zambezi Belts is related to a collision occurring between approximately 560-550 Ma, along a southeast- to northwest-trending suture, between a plate comprising the Kalahari Craton and the southwestern part of the Congo Craton, and a plate consisting of the remaining part of the Congo Craton (Olson, 2000; Porada and Berhorst, 2000). The Mwembeshi Zone is thought to mark this suture (Unrug, 1983; Coward and Daly, 1984), and can possibly be correlated with the Okahandja Lineament Zone of the Damara Belt (see following section) based largely on similarities in alignment, shear sense, and transport direction (Coward and Daly, 1984). The last regional metamorphic event in the Zambezi Belt is thought to have occurred at 550-500 Ma (Hanson *et al.*, 1998; Dirks *et al.*, 1998).

3.5.2 The West Congolian Supergroup

The West Congolian Supergroup comprises sandstones, quartzites, conglomerates, schists, and stromatolitic limestones, and may extend as far north as Gabon (Kogbe and Burollet, 1990). The West Congolian Belt is a possible continuation of the Damara Pan-African Belt, but is separated from the Kaoko Belt by Archaean rocks in Angola, and perhaps is more likely to be related to the Lufilian Belt of Zambia (G.S. de Kock, *pers. comm.*). The lithostratigraphic sequences of the Katanga and West Congolian Supergroups in the Democratic Republic of Congo are also similar (Kanda-Nkula *et al.*, 2004). Deformation of the West Congolian Belt is constrained by an Ar-Ar age of 566 ± 42 Ma (Boven, cited in: Kanda-Nkula *et al.*, 2004).

3.5.3 The Damara Belt

3.5.3.1 Rifting

The oldest Damara Supergroup unit in Namibia that is shown on the sub-Kalahari geological map is the Tsumis Group, consisting of coarse- to fine-grained clastic sedimentary rocks and local intercalated mafic and acid volcanic rocks (Hoffmann, 1989). The Tsumis Group is made up by the Doornpoort, Klein Aub and Eskadron Formations. The Tsumis Group rocks were originally considered as pre-Damaran in age and were included in the Sinclair Supergroup (e.g. SACS, 1980; Miller, 1983), but subsequently were found to be younger than the 1000 Ma regional deformational event affecting rocks of the Sinclair Supergroup and associated plutonic intrusives in the Rehoboth-Nauchas Massif of Namibia (Hoffmann, 1989).

It is believed that rifting began at a triple junction with north, south and northeast trending branches, and the development of the Damara rifts follow the same northeast-southwest and northwest-southeast trends of structural weakness as those that influenced the orientations of the Koras-Sinclair-Ghanzi Rift (Borg, 1988). The northeastern arm extends underneath the Kalahari of Botswana and possibly joins with the Katangan and Zambezi Belts. This extension into Zambia and the Democratic Republic of Congo is, however, tenuous and it is more likely that the Damara tapers out towards the east.

Sedimentation and limited volcanism in northern Namibia are thought to have occurred towards the end of an initial phase of rifting (de Kock *et al.*, 2000). Dates for the Naauwpoort volcanic rocks (Nosib Group), near Khorixas in northern Namibia (shown on Fig. 3.6), of 750 ± 60 Ma and 728 ± 40 Ma (Miller and Burger, 1983), 746 ± 2 Ma and 747 ± 2 Ma (Hoffmann *et al.*, 1986), and 752 ± 7 Ma (de Kock *et al.*, 2000), as well as an age of 756 ± 2 ma for the Oas Syenite which intrudes the Naauwpoort Formation (Hoffmann *et al.*, 1986) suggest rifting was older than 756 Ma but may only have started as late as ~800 Ma (G.S. de Kock, *pers. comm.*), much later than the 900-1000 Ma originally proposed by Miller (1983).

The Nosib Group of northern Namibia has been correlated with the Chela Group of Angola on the basis of lithostratigraphic correlations, as well as similar depositional settings, typical of initial infilling of a basin or basins (Kröner and Correia, 1980). In addition to this, Kröner and Correia (1980) report that work from various researchers indicates similar palaeomagnetic data for the Nosib and Chela Groups. This correlation between the Chela and the Nosib has, however, been questioned by recent SHRIMP age dating which suggests that the Chela Group may be significantly older and is Palaeoproterozoic in age (McCourt *et al.*, 2004b).

Sea-floor spreading in the Southern Zone is thought to have occurred either prior to the deposition of the Chuos Formation (Henry *et al.,* 1988), or later with the onset of the carbonate platforms (de Kock *et al.*, 2000), and resulted in the production of mid-ocean ridge basalts (of the Matchless Amphibolite Belt) in the Khomas Sea (Killick, 2000). Following the period of rifting, subsidence, and drifting and deposition of sediments is thought to have occurred from 730 to 650 Ma (G.S. de Kock, *pers. comm.*). During this time the Otavi and Swakop Groups of northern Namibia were deposited. The Aha Hills Formation of Botswana is correlated with the Otavi Group of Namibia (Carney *et al.*, 1994), and in northern Namibia, rocks of the Leba Formation and Tchamalinda Group (southern Angola) are correlatable with the Tsumeb Subgroup of the Otavi Group, on basis of lithostratigraphy (Kröner and Correia, 1980). The Roibok (or Rooibok) Group of Botswana consists of a highly magnetic group of metamorphic rocks described from a borehole core in the Roibok Laagte valley of north western Botswana (Lüdtke *et al.*, 1986), and is believed to be either an extension of the Matchless Amphibolite Member of the Kuiseb Formation, Damara Supergroup of Namibia (Reeves, 1978c; Lüdtke *et al.*, 1986), or a correlative of pre-Damaran metamorphites (Carney *et al.*, 1994). The Roibok Group comprises amphibolites (Lüdtke *et al.*, 1986), a pink granitic gneiss, and alternating, centimetre scale layers of biotite and muscovite schists (Carney *et al.*, 1994). A U-Pb zircon age for the granite gneiss within the Roibok Group indicates a crystallisation age of 716.8 ± 2.2 Ma. To the north of the Roibok Group rocks, greenschist facies marbles of the Kaonaka (or Koanaka) Group are structurally juxtaposed on the igneous and meta-igneous rocks of the Chihabadum Complex (Key and Ayers, 2000; Singletary *et al.*, 2003).

3.5.3.2 Closure and the assembly of Gondwana

From 650 to around 630 Ma reversal of spreading occurred and a period of subduction caused deformation (Miller, 1983). This was the start of the Neoproterozoic Pan-African event which resulted in the development of the Damara Belt (made up of several tectonic zones)(Fig. 3.9). The intra continental branch of the Damara Belt is 400 km long and 150 km wide, and extends northeastwards from Namibia into Botswana where it is covered by Karoo Supergroup and

Kalahari Group rocks. In Namibia it is divided into a Northern Platform, Northern Zone, Central Zone (Northern and Southern), Okahandja Lineament Zone, Southern Zone, Southern Marginal Zone and Southern Foreland. In Botswana, however, the Ghanzi-Chobe Zone, Okavango Zone, and Shakawe Zone are present(Carney *et al.*, 1994). The Northern and Southern Zones do not appear to extend into Botswana.

From figure 3.9 it is evident that the Okavango Zone is defined by a continuation of Namibia's Southern Margin or Areb Mylonite Zone (Lüdtke *et al.*, 1986; Carney *et al.*, 1994). It also contains part of Namibia's Southern Zone, and its northern boundary appears to be an eastwards continuation of the Okahandja Lineament (Carney *et al.*, 1994). The Okahandja Lineament is a tectonic feature separating the Okahandja Lineament Zone and the Central Zone, and can be easily distinguished on aeromagnetic and satellite images (Miller, 1979). To the west of 17°30' E the magnetics north of the lineament are highly disturbed compared to those of the Okahandja Lineament Zone to the south. To the east, however, the magnetic imprints on both sides of the Okahandja Lineament are disturbed and reflect a change in the structural style of the Southern Zone (Miller, 1979). The enhanced disturbed magnetics in the eastern part of the Southern Zone might be due to the large scale uplift exposing deeper stratigraphic levels (Miller, 1979). It seems that the Okahandja Lineament represents a major synsedimentary as well as syntectonic structural boundary within the Damara orogen, possibly controlled by large-scale crustal faults defining the thinned crustal edge of the Congo Craton (Miller, 1979; de Kock, 1992). The final movements on the Lineament deformed some of the schists of the Okahandja Lineament Zone (de Kock, 1992) and appear to pre-date the Donkerhuk Granite which is intruded across it (Miller, 1979) and which has an age of 476 ±4 Ma (de Kock and Walraven, 1995).

The Shakawe Zone is a possible equivalent of the Northern Platform and Northern Zone of Namibia (Carney *et al.*, 1994), although the orientation of structures in the Shakawe Zone suggests that sediments in this zone may be part of an older cover on the Congo Craton (Key and Ayres, 2000). In Namibia the extent of these rocks is ascertained only from aeromagnetic images as they do not crop out. In the north-western corner of Botswana, they are exposed however, and

these folded carbonates and clastics (Carney *et al.*, 1994) are known as the Xaudum and Tsodilo Hills Groups. Detrital zircons from the sandstones in this area were dated and three main groups of U-Pb SHRIMP ages occurred at about 1020 Ma, 1090 Ma and 2050 Ma (Mapeo *et al.*, 2000b). The youngest detrital zircon age was taken to be the maximum age of deposition which means that these rocks can probably be correlated with the Ghanzi Group.

In the south, the rocks of the Ghanzi Group were deformed at around 650 Ma resulting in the formation of a fold and thrust belt known as the Ghanzi-Chobe Belt (Modie, 2000). The southeastern boundary of the Ghanzi-Chobe Belt was identified as the northeasterly trending portion of the Kalahari Suture Zone, also called the Makgadikgadi Line (Reeves, 1978 b,c)(see 3.4.1.4), although, subsequently, further evidence suggested that the Tsau Fault, a locus of southeast directed thrusting (Pretorius, 1984), may be the southeastern boundary (Carney *et al.*, 1994). The northwestern boundary of the rift is defined by a Damaran thrust zone, with the highly deformed Roibok Group rocks occurring along the edge of the KSG Rift (Key and Mapeo,1999). As these bounding thrust zones are younger than the KSG Rift, its original width is not known. The Roibok Group also separates the supracrustals of the Kgwebe/Goha Hills Formations and younger Ghanzi Group/ Chinamba Hills Formation in the southeast from the magmatic and migmatitic rocks (and locally occurring carbonates) to the northwest (Key and Mapeo,1999).

Deformation of the Neoproterozoic Ghanzi Group increases westwards where the underlying Kgwebe Formation rocks become exposed in antiformal fold cores. There is also a marked decrease in the thicknesses of the Neoproterozoic strata towards the west, varying from over 10 km in the Passarge and Nosop Basins to just 4 km in the more folded western areas. This variation in thickness may be primary, and reflecting slower subsidence to the west but is also possibly attributable to subsequent erosion following Damara orogenesis (Key and Mapeo,1999).

The final closure of the Adamastor Ocean and collision between the Rio de la Plata and São Francisco-Congo Cratons occurred between 650 to 600 Ma ago, with this collisional event followed by the accretion of the Kalahari Craton at around 550 Ma (Prave, 1996; D'Agrella-Filho *et al.*, 1998, Alkmim *et al.*, 2001) resulting in the obduction of Khomas Ocean rocks onto the margin of the Kalahari Craton (Borg, 2000). As collision between the Kalahari and Congo Cratons occurred, related uplift of the Otavi Carbonate Platform foreland terranes resulted in deposition of the Mulden and Nama Group sediments between 560 and 450 Ma in the peripheral foreland basins (Clauer and Kröner, 1979; Borg; 2000). The Gariep orogeny (e.g. Reid *et al.*, 1991) which involved the reworking of the terranes formed during the collision between the South America and Congo Cratons had a peak of metamorphism of around 542 Ma (Frimmel and Frank, 1998).

The so called "Pan-African Event" lasted until about 450 Ma and appears to overprint much of the earlier Kibaran and Irumide orogenies. The peak of metamorphism occurred in the southern Damara between about 528 and 520 Ma, and resulted in the migmatisation of schists and igneous rocks in the southern edge of the southern Central Zone (G.S. de Kock, *pers. comm.*). Diorites and the majority of the granites in Namibia were emplaced between 560-500 Ma (Tack *et al.*, 2002), and later post-orogenic intrusion of granites like the Donkerhuk Granite (~476 Ma) occurred during the continental collapse phase (de Kock and Walraven, 1995). Doming, uplift, folding and thrusting, occurred during the period up to 280 Ma, resulting in the formation of the pre-Karoo surface through the erosional removal of the Damara mountain belt (Miller,1983).

3.5.4 The Passarge Basin

Between the Makgadikgadi Line and the Ghanzi-Chobe Belt is an area of deep magnetic basement (Reeves, 1978 b,c; Fig. 3.3) overlain by a very thick sequence of non-magnetic sediments, possibly belonging to the Ghanzi and the younger Nama Groups. This Passarge Basin occurs in a zone separating the Ghanzi-Chobe Belt from the Kalahari Suture Zone. This zone was originally described by Jones (1979) as a medial rift (graben) and according to Pretorius (1984) the magnetic basement reaches depths of over 15000 m below the surface here (Fig. 3.3). The medial rift, approximately 30 km wide in the northeast and 200 km wide in the southwest, may possibly extend towards the northeast into Zimbabwe, where more recent rift faulting has been observed. According to Jones (1979) the deep magnetic basement of the Nosop and Ncojane Basins are possibly also related to the medial rift.

Pretorius (1979,1984) interpreted the Passarge Basin not as a rifted graben structure, but rather as a down-dragged feature related to the northwestern edge of the Passarge Basin, the Tsau Fault zone. The model Pretorius (1984) proposed involves not extension but rather compression, with the Ghanzi-Chobe Belt material being thrust towards the southeast, onto the older rocks of the craton. This interpretation defines the Passarge Basin as a marginal trough on the foreland side of a thrust-front which is a similar scenario to that seen to the northeast in the Irumide belt. Further evidence to the southwest was described by Coward and Daly (1984) where Damara strike faults were subsequently reactivated as overthrusts verging to the southeast.

The apparent extensional features observed in the area are younger features that are better attributed to the Karoo and post-Karoo rifting (Fig. 3.14) and were not responsible for the Passarge Basin itself. However, the thick sequences of Karoo rocks, as well as the thick Kalahari succession, recorded in the vicinity of the Passarge Basin, indicate the importance of the feature in later crustal developments.

3.5.5 The Nama and Okwa Groups

The Neoproterozoic Nama Basin, situated between the Kalahari Craton in the east and the Damara and Gariep Belts in the north and west, developed as a foreland basin on the southern foreland of the Damara Orogen (Germs, 1983). This foreland basin can be subdivided from north to south into the Zaris, Witputs and Vanrhynsdorp Basins (Gresse and Germs, 1993). The known Nama rocks in Botswana (directly underlying the Kalahari Group) (see sub-Kalahari geological map) are largely restricted to a graben in the Okwa Valley inlier (Aldiss, 1988), but are probably also found beneath the Karoo in the Passarge Basin and the Nosop-Ncojane Basin (Aldiss, 1988). In the latter the Nama Group may exceed 1000 m thickness (Key and Ayers, 2000).

Evidence suggests that deposition in the Nama Basin occurred between 590 and 530 Ma (Gresse and Scheepers, 1993) or 630 and 530 Ma (Gresse and Germs, 1993) in response to the Pan-African I orogenesis, which was synchronous with deformation and granitoid emplacement in the Central and Northern Zones of the Damara Belt (Miller, 1983). The Nama rocks were tectonised towards the end of the Pan-African II orogenesis (540-490 Ma)(Gresse and Germs, 1993, Carney *et al.*, 1994) in response to uplift and thrusting. According to Germs (1983) no further accumulation of Nama Group sediments has occurred after orogenesis ended, which Miller (1983) puts at 420 Ma. Carney *et al.* (1994) suggest that the Passarge and Nosop/Ncojane Basins were inverted following the Pan-African climax at 540 Ma. The strata from these basins were then exposed and eroded until deposition of the Permo-Carboniferous Karoo Supergroup.

In the Okwa valley of Botswana, metasedimentary rocks with associated limestones and felsites of the Okwa Group (Aldiss, 1988; Aldiss and Carney, 1992) outcrop in a small area. The Okwa Group consists of two unconformity-bounded assemblages, the lower and upper Okwa Group. The lower Okwa Group, which can be correlated with the Sinclair Group (Aldiss and Carney, 1992), was possibly deposited in a northeast-trending rift system, possibly corresponding to one of the Koras-Sinclair-Ghanzi rifts (see section 3.4.9). Closure of the Okwa Basin probably occurred at ~150 Ma (Aldiss and Carney, 1992). Ages from detrital zircons extracted from the Takatswane Formation of the upper Okwa Group gave an age of 579 ± 12 Ma, which provided the maximum depositional age for that part of the Okwa Groups (Ramokate *et al.*, 2000). The similarity between this age and that for the Nama Group as well as lithological similarities suggest that the upper part of the Okwa Group can be correlated with the Nama Group (Aldiss and Carney, 1992; Ramokate *et al.*, 2000; T.Majaule, *pers. comm.*).

3.6 Karoo Supergroup deposition and the break-up of Gondwana

Boreholes drilled through the Kalahari Group rocks have shown the extent of Karoo Supergroup rocks underlying the Kalahari Basin to be much greater than previously thought (Figure 3.10), with much of the previous known extent of Karoo in Botswana believed to be limited to an area north of the Zoetfontein Fault (Green, 1966). In the map area, Karoo Supergroup rocks were deposited primarily in the Botswana (Kalahari) and Zambezi Basins (Johnson *et al.,* 1996). The Karoo basins in Botswana consist of the Southwest Botswana Basin, Northeast Botswana Basin, and Central Kalahari Basin (R.A. Smith, 1984). Karoo sedimentary rocks in Namibia occur mainly in smaller basins in the southeast and northwest of the country, and their thicknesses vary between about 1500 m in the south to between 700 and 250 m in the Huab, Owambo and Waterberg Basins in the northwest (Dingle *et al.*, 1983; Hegenberger, 1988; Horsthemke *et al.*, 1990).

The controls over the Karoo Supergroup deposition in the Botswana Basin are not well

documented, but they were believed to have been deposited in an intracratonic thermal sag basin (Johnson *et al.*, 1996). There is strong evidence that faulting had some influence on sedimentation (e.g. R.A. Smith, 1984), with much of the faulting in turn influenced by older structural features. In the mid-Zambezi Valley faulting controlled Karoo Supergroup sedimentation (Nyambe and Nkemba, 2004) and in eastern Botswana, the Karoo sedimentary rocks were deposited in graben-like troughs which also contain the thickest accumulations of Waterberg Supergroup material (Green *et al.*, 1980). Evidence of pre-, intra- and post-Karoo movements on the boundary faults shows that these features have been reactivated at various times in the past (Green *et al.*, 1980). It is thought that syndepositional reactivation along the Kalahari Line and Zoetfontein Fault may have influenced Karoo sedimentation (Hutchins and Reeves, 1980) and R.A. Smith (1984) believed that the same ENE-WSW shears and faults which developed in the Limpopo Belt influenced Karoo sedimentation in a zone between the Zoetfontein and Letlhakane Faults.

3.6.1 Carboniferous

In the Early Carboniferous (360-340 Ma), sedimentation of the upper Cape Supergroup occurred in the Witteberg Basin, covering the tip of southern Africa and the Falkland Islands. The supercontinent of Pangaea formed in the Carboniferous when Gondwana and Laurussia collided (Burke and Dewey, 2002). By the Mid-Carboniferous (330-310 Ma), large-scale uplift occurred over much of Gondwana, resulting in the termination of sedimentation, while ice caps had started forming over elevated areas then situated over the South Pole. During the Late Carboniferous these ice caps had developed into major ice-spreading centres (Visser and Praekelt, 1996).

3.6.2 Carboniferous- Permian

Glacial sedimentation started in the southern regions of Gondwana in the Late Carboniferous-Early Permian (300-280 Ma), with the subsequent Karoo Supergroup deposited in the Karoo-Falklands basin, part of a retro foreland-arc basin caused by the oblique subduction of the palaeo-Pacific plate. This subduction also resulted in the opening of the Botswana -Zambezi Basin (Fig. 3.11)(Visser and Praekelt, 1996). Further to the north, crustal sagging occurred in the Congo Basin while the East Africa-Malagasy and Peninsular India basins resulted from crustal extension.

The Dwyka glaciation started during the Carboniferous and lasted until the Early Permian. Zircons from tuff beds in the Dwyka Group of southern Namibia and South Africa have been dated at 302 ± 3 Ma and 299 ± 3 Ma (Bangert *et al.*, 1999), and palaeomagnetic data from the sedimentary rocks indicated that the Permo-Carboniferous glaciation lasted from about 340 to 290 Ma (Opdyke *et al.*, 2001). Dwyka glaciation had a major smoothing effect on the topography of the area, with glaciers eroding high-lying areas and glacial deposits accumulating in fault-controlled valleys (Visser, 1989). The major ice sheet was centred over Zambia, Zimbabwe and the northern parts of South Africa, with lobes extending outwards to adjacent areas. The ice sheet in the Botswana Basin moved in a generally southwesterly direction, dominated by the Botswana Ice Lobe (Frakes and Crowell, 1970; Visser, 1983). Once the ice sheet started to break up, and the glaciers started moving more freely, deposition of the basal tillites began, with material transported from the surrounding highlands (e.g. the Cargonian Highlands) into lower-lying areas in the Botswana Basin (Visser, 1983) and southwards into the main Karoo Basin (Visser *et al.*, 1997). No deposition of Dwyka Group sedimentary rocks is believed to have occurred on the Cargonian Highlands themselves (Veevers *et al.*, 1994). In South Africa the Dwyka Group consists of a basal tillite followed by a sequence of shale, siltstone and mudstone, sandstone carbonate lenses and a main tillite, and in the Hotazel Valley an upper tillite is also present (Visser, 1983). In the Northern Cape Province of South Africa, the sub-Kalahari geological map shows a trough of Dwyka sedimentary rocks extending southwards from the Botswana border to near the town of Kathu, to the west of Kuruman. It is not clear whether the Dwyka is a remnant of a much larger covering of tillite in the area, preserved in a subsequent down-faulted graben-like structure that may have been reactivated by the Late Jurassic Morokweng meteorite impact (see 3.4.3), or if it is limited to a glacial valley which may have been controlled by existing structures. In south western Botswana, the Dwyka Group is subdivided into the Middleputs, Khuis and Malogong Formations, while in the rest of the country, to the east of the Kalahari Line, only one formation exists, the Dukwi Formation (R.A. Smith, 1984; Johnson *et al.*, 1996). In Angola, the lower units of the Karoo, which are the equivalents of the Dwyka and Ecca Groups of the other countries, are combined under the Lutoe Group (Mouta and Cahen, 1951; M.V. da Silva, *pers. comm.*).

3.6.3 Permian

Sediments of the Ecca sequence were conformably deposited over the Dwyka sedimentary rocks during the Lower to Middle Permian. In Botswana the Ecca Group can be divided into five formations with only three of these present to the east of the Kalahari Line (Rahube, 2003). Tuff beds in the Prince Albert Formation of the southern Karoo Basin have been dated at 288 ± 3 Ma (Bangert *et al.*, 1999). The lower Ecca sequence was deposited under shallow lacustrine conditions, which may have been open to the sea in the west. These muddy sediments grade up to more sandy deposits typical of deltaic settings (Johnson *et al.*, 1996). These deltaic conditions, which favoured the accumulation of peat, resulted in the vast reserves of coal of the Ecca Group (see Chapter 6).

To the north of the Zoetfontein Fault, the Ecca Group appears to change from having three formations on the eastern side of the Kalahari Line, to only two formations on the western side of the Line (Key and Ayers, 2000). There is also some evidence of syn-depositional downthrow on the southern side of the Zoetfontein fault (R.A. Smith, 1984).

During the Early-Late Permian (260-255 Ma), the Congo Basin shrank, and right-lateral movement along the Falklands fracture zone resulted in the formation of the Natal trough. Similar movement along the northern part of the Atlantic fracture zone created a seaway between the Karoo-Botswana and the eastern part of the Paraná Basins (Fig. 3.12)(Visser and Praekelt, 1996).

3.6.4 Late Permian-Early Triassic

A major shear zone is thought to have developed in the Damara Belt, stretching further to the east into Zambia (Coward and Daly, 1984) during the period following glaciation. This Southern Trans-Africa shear system (STASS) (de Wit, M. *et al.*, 1995) may have developed during the Permian (Fig. 3.12; Visser and Praekelt, 1996) but was more probably related to the Late Permian to Triassic collisional event that caused the Cape Fold Belt (G.S. de Kock, *pers. comm.*). The STASS is thought to have controlled basin development in the interior, with inversion tectonics occurring, and basins developing in the uplifted mid-Carboniferous belts (Visser and Praekelt,

1996). It is possible that movement of southern Africa together with Madagascar-India-east Antarctica occurred along the STASS during the Permo-Triassic and then later again in the Jurassic-Cretaceous (de Wit, M. *et al.*, 1995).

The Cape Fold Belt, consisting of a western and southern branch was formed as a result of the crustal shortening from the west and south, causing vertical thickening of the sequence during the Late Permian and Early Triassic (Fig. 3.13)(Visser and Praekelt, 1996). The orogenesis associated with the Cape Fold Belt formation resulted in the destruction of the inland seas and their partial replacement by fluvial (foreland) basins in the southern edge of the continent (Tankard *et al.*, 1982; Visser and Praekelt, 1996). At the same time uplift of parts of the Congo and Kalahari Cratons occurred, and the Late Palaeozoic strata overlying them was eroded (Visser and Praekelt, 1996).

Sediments of the Beaufort Group were deposited in the foreland basins in the south of southern Africa as well as in the Botswana-Zambezi basins further to the north. Deposition occurred in lakes and flood-plains controlled by down-warping and faulting (e.g. Rust, 1975; Yemane and Kelts, 1990). Evidence for syn-depositional regional uplift has been recognised from major unconformities at the top and bottom of the Beaufort Group in Botswana (Key and Ayres, 2000). Sandstones, siltstones and conglomerates of the Lebung Group of Botswana, Lower Stormberg Group of Angola and Upper Karoo Clastics of Zambia and Zimbabwe were deposited in terrestrial fluviatile conditions during the Triassic and Early Jurassic, with the Lebung Group sediments in Botswana being interlayered with basalt flows (R.A. Smith, 1984; Key and Ayres, 2000).

3.6.5 Rifting and the break up of Gondwana

Figure 3.14 shows the distribution of rifts in southern Africa. The rifts shown in blue include both those rifts which may have formed in the Carboniferous or Permian during the final assembly of Pangaea (Burke and Dewey, 2002) as well as those which down-faulted Karoo Supergroup rocks during the break-up of Gondwana. Some of the oldest Karoo rifts in the area probably originated in the Early Permian and formed a small number of basins that were largely limited to the east coast of Africa and southwestern Madagascar (Lambiase,1989).

Down-faulting in the Mid-Zambezi Basin in the Early Jurassic (Cox, 1970; Gough and Gough, 1970; Rust, 1975), followed older fractures and structural weaknesses of Late Proterozoic age (McConnell, 1972), and reactivated some of the Late Triassic basins (Lambiase, 1989) and the Karoo sedimentation in the Botswana-Zambezi Basin was ended by the onset of a period of Early- to Mid-Jurassic volcanic activity during which basalts were extruded across the region (Fig. 3.10). The magmatic episode which resulted in the Karoo volcanism and intrusion of the dykes was believed to be initiated by the emplacement of a mantle plume to the east of the present KwaZulu-Natal coast. This Karoo plume is thought to have produced the Nuanetsi triple junction, as well as a second triple junction further to the south, the Weddell triple junction (Fig. 3.15), which resulted in the formation of the main Karoo Basin dyke swarm at ~180 Ma. This mantle pluming can also be linked to the spreading in the Somali Basin, which eventually displaced Madagascar, India and Antarctic portion of Gondwana southward in relation to Africa (Hartnady, 1990).

The basalts of the Upper Stormberg Group reach thicknesses up to 1000 m (Key and Ayres, 2000) in places and stretch from north of the Zoetfontein Fault in Botswana up into southern Angola and Zambia. In much of Botswana, Zambia and Namibia the uppermost basalts are found overlying older sedimentary deposits. Ages for the Karoo basalts underlying the Kalahari Group range from 181.3 ± 1.5 Ma to 179.2 ± 0.9 Ma near Victoria Falls in Zimbabwe (Jones *et al.*, 2001), with other ages from Botswana giving ages of 180 ± 10 Ma (Coates *et al.*, 1979), 183 ± 1 Ma (Duncan *et al.*, 1997).

Aeromagnetic surveys over Botswana have revealed the distribution of late- or post-Karoo dyke swarms, with the most prominent occurrence of dykes known as the Botswana swarm (Wilson, 1990), or Okavango giant dyke swarm (Le Gall *et al.*, 2002). The Botswana dyke swarm stretches from northern Namibia, where its margins would probably converge to the north of Etosha (Reeves,1978a), to its widest point on the eastern border of Botswana and down to a triple junction near Nuanetsi in southeastern Zimbabwe, where it intersects with the Rooi Rand swarm and a swarm of dykes with a northeasterly orientation extending approximately 380km towards the border of Mozambique (Limpopo Swarm)(Wilson, 1990). Near the intersection with the Limpopo Swarm, some of the southernmost dykes from the Botswana Swarm swing towards the east, closer to the structural trend of the Limpopo Belt (Wilson, 1990). The aeromagnetic and gravity coverages (Figs 3.1 and 3.2) as well as the sub-Kalahari geological map clearly show the

extent and orientation of this dyke swarm and it is also shown on Figure 3.15.

According to Coates *et al.* (1979), the Botswana dyke swarm was controlled by a pre-existing west-northwest fracture pattern in the Karoo strata. Basing his model on an age of ~140 Ma age published for the dolerite dykes near Serowe (R.A. Smith, 1984), Reeves (2000) believed that the clockwise movement of South America and Antarctica away from Africa opened the pre-existing fractures in southern Africa, along which the Botswana dyke swarm intruded, with these fractures possibly propagating centripetally towards a rotation pole situated on a stationary plate, as a second plate moves around an arcuate transform fault (the Agulhas Fault). Supporting evidence for this model came from the coincidental alignment of the Tristan hotspot, the Botswana dyke swarm, and the Nuanetsi igneous province of southeastern Zimbabwe. Reeve's (2000) theory was disproved by recent $^{40}Ar/^{39}Ar$ ages which show that the dykes of the Botswana and Limpopo swarms were emplaced between 179 ± 1.2 Ma and 178 ± 1.1 Ma (Le Gall *et al.*, 2002), some 30 million years before rifting between South America and southern Africa is thought to have commenced (K.C.A. Burke, *pers. comm.*). Evidence does suggest, however, that the dolerite emplacement did follow pre-existing structures as suggested by Coates *et al.* (1979). The dating of older, Proterozoic, dykes in the Jurassic dyke swarm suggests that the Botswana dyke swarm was emplaced along a reactivated Proterozoic or older dyke or fracture zone that had the same NW-SE orientation (Le Gall *et al.*, 2002; Tshoro *et al.*, 2004). This also calls into question the emergence of the plume-induced Karoo triple junction at Nuanetsi, which has been described instead as magma intruding a pre-existing zone of fractures (Jourdan *et al.*, 2004; Le Gall *et al.*, 2004).

The absence of Karoo basalts beneath the Kalahari Group sedimentary rocks in the vicinity of the Makgadikgadi Basin would suggest either post-extrusion uplift along the dyke swarm and subsequent erosion of the basalts, or that the area was a topographical high when the basalts were extruded (Reeves, 1979). The former possibility is supported by work of Stansfield (1973) in which discontinuities in the Karoo stratigraphy were interpreted as suggesting uplift and erosion. Both the Makgadikgadi Basin and Okavango Delta lie along the axis of the dyke swarm, which may suggest a relationship between the northwest-trending structures of the dyke swarm and the northeast-southwest rifting and faulting that controls the formation of these features (see Chapter 5).

As the sub-Kalahari geological map shows, dykes with a similar orientation also occur to the south of the main dyke swarm and numerous dykes and sills have also been interpreted from geophysical data in the southwest of the basin, particularly along the Namibia-Botswana border where extensive dolerite sills lie directly under Kalahari sedimentary rocks.

The continental rifting between Africa and South America is thought to have commenced during the late Jurassic, when the Tristan Plume erupted at about 133 Ma (K.C.A Burke, *pers. comm.*). Soon after eruption, spreading started, and break-up finally occurred around 129-121 Ma (Fouche *et al.*, 1992). The effect of the Karoo Plume on the interior of Africa was to create a system of rifts stretching across the continent. In particular, the separation of Madagascar and the Seychelles from Africa at around 150-112 Ma is thought to have formed transform faults that were responsible for the formation of a series of grabens across southern Africa (G.S. de Kock, *pers. comm.*). The rifting would have resulted in the lowering of Karoo Supergroup rocks into NE-SW trending grabens where they were protected from erosion and preserved, as is the case with the Luangwa graben which is truncated at about 13°-14° S by a dislocation zone before continuing further to the west as the Luano Rift (Basin) and the Kafue Basin. Post-Karoo faulting in western Zimbabwe has the same orientation as foliations in adjacent Precambrian rocks (Vail, 1967) and is represented by the Wankie Fault which has displaced Karoo beds several hundred metres (Rogers, 1936). Down-throw of Karoo rocks on the northern side of the Zoetfontein Fault of up to 300 m has been measured along with several hundred metres throw on other post-Karoo NNW-SSE oriented faults intersecting the Zoetfontein Fault (R.A. Smith, 1984). Older structural trends were once again exploited. The STASS was possibly reactivated during this time (de Wit, M. *et al.*, 1995).

3.6.6 Kimberlites

Kimberlites are for obvious reasons the most studied, explored and economically significant of all the intrusions found beneath the Kalahari Group rocks. Kimberlite pipes are found distributed throughout most of the Kalahari-covered area, with some of the pipes being diamondiferous (see Chapter 7).

Emplacement of kimberlites has occurred in southern Africa since the Archaean, as is evinced by the presence of diamonds in Witwatersrand Supergroup sedimentary beds (Smith and Barton, 1995). Kimberlites of the Kuruman Group are the oldest known kimberlites in southern Africa and were emplaced 1700-1600 Ma (Smith and Barton, 1995). The 2424KD2 pipe at Jwaneng gave a radiometric age of 235 Ma (Kinney *et al.*, 1986), which is the only kimberlite of that age in the region (Smith and Barton, 1995) and the pipe at Finsch Mine has a minimum age of 188 Ma (Smith *et al.*, 1985) and covers 17.9 ha (Lynn, 1998). Kimberlites in the North Cape and Free State Provinces were emplaced at 120-118 Ma, with the range possibly extending from 125-113 Ma (Smith and Barton, 1995). At 106-102 Ma some kimberlites were intruded in the North Cape Province and in Angola. In the Orapa pipes of Botswana a study of plant remains and fossil insects from the crater facies sediments has revealed an age of Middle to Late Cretaceous (McKay and Rayner, 1986; Rayner *et al.*, 1991, 1997), which was confirmed by U-Pb and fission track ages of around 92 Ma (Davis, 1977; Haggerty *et al.*, 1983). The most recent kimberlites in the region were emplaced in the Gibeon area of Namibia and near Prieska at 74-70 Ma (Smith and Barton, 1995).

The kimberlites emplaced during the 120-118 Ma period are thought to be either due to a change in direction in the motion of the African Plate, or as a result of lithospheric stresses induced when the Atlantic Ocean started opening, while the 90-82 Ma kimberlites may have been emplaced during a period of particular rapid plate motion (Smith *et al.*, 1994; Smith and Barton, 1995).

3.7 The Morokweng Impact Structure

Gravity and aeromagnetic images of the Morokweng area in the northwestern Cape Province of South Africa (see Figs 3.16a,b) show the presence of a circular anomaly, roughly 70 km in diameter, beneath the Kalahari sediments. Although the surface topography is fairly flat, Kalahari Group isopachs clearly show a circular feature with anomalous thicknesses of Kalahari Group sediments (see isopach map of the Kalahari Group, Appendix B).

Shallow boreholes drilled by the Department of Water Affairs failed to detect anything unusual and the area was interpreted by Smit (1977) as underlain largely by basement granite as well

as rocks belonging to the Kraaipan Group and small areas of Dwyka Group tillite (as the sub-Kalahari geological map shows). Corner *et al.* (1986) and Stettler (1987) interpreted the anomalous geophysical feature as an intrusive, low-density body of granite or magnetite syenite.

Subsequently, however, a recognition of the similarity of the aeromagnetic signature to known impact sites (Andreoli *et al.*, 1999a), the identification of planar deformation features (PDF's) and other shock deformation features (Andreoli *et al.*, 1995; Corner *et al.*, 1996; Hart *et al.*, 1997) and anomalous Ni and Ir concentrations in possible impact melt rocks (Corner *et al.*, 1996; Hart *et al.*, 1997) has enabled reinterpretation of the feature as a meteorite impact structure.

Corner *et al.* (1996) have described it as follows:

1) It is roughly pear-shaped with an extension to the south, rather than perfectly circular;

2) It has a diameter of about 30 km;

3) It is flanked to the north, east and south by a concentric, 20 km wide, magnetically quiet zone;

4) This quiet zone, also called the inner ring structure, is characterised by a loss of amplitude of a northeast-trending dyke swarm (see sub-Kalahari geological map, Appendix C).

Work conducted by various researchers (see Corner *et al.*, 1996) has identified a charnockitic body overlying granitoid basement. The evidence of shock metamorphism in the Archaean granitoids and a significant iridium enrichment in the charnockite resulted in the interpretation of the latter as an impact melt. A quartz norite buried in the centre of the structure yielded a Late Jurassic age of 145 ± 0.8 Ma (U-Pb age on a zircon) and 144 ± 4 Ma (Ar-Ar age on a biotite) (Hart *et al.*, 1997).

Andreoli *et al.* (1999b) describe the crater as comprising an interior basin of a ± 25-30 km quartz norite granophyre with inclusions of shocked, brecciated basement, surrounded by a magnetically quiet inner ring (diameter of 60 km) of allochthonous impact breccia and minor granophyric impact melt. The outer edge of the inner ring is defined by a series of ring-faults, separating it from the outer ring. The outer ring comprises faulted Proterozoic cover and Archaean basement, as well as undifferentiated megabreccia. The sub-Kalahari geological map shows the extent of the three rings.

Corner *et al.* (1996) have illustrated a much larger circular structure, with a diameter of approximately 340 km, centred on the Morokweng Structure. Andreoli *et al.* (1999b) also record post-impact radial faulting found up to 170 km from the crater centre. They noticed that the geological and geophysical expression of the Morokweng Impact Structure is surrounded by a similar zone of 200 km wide. If the outer structure does exist and represents another ring, the 340 km diameter feature will be one of the biggest impact structures in the world. If this is the case, then the inner anomaly (70 km wide) would represent the remnant of the central uplift area, which would most likely have the appearance of a peak-ring structure.

It is also possible, however, that the outer perimeter of the inner zone (± 70 km) represents the full extent of the eroded structure (Corner *et al.*, 1996). Recent analysis of drill core from 40 km west of the MIC centre suggests that maximum crater diameter is less than 80 km (Reimold *et al.*, 2002) which indicates that the central magnetic anomaly delineates the collapsed central uplift of a 70 km wide impact structure capped by impact melt rock.

In establishing the presence of a meteorite impact crater, the sub-Kalahari geology was adapted accordingly from the data presented by Smit (1977) on his geological map of Bray (1:250000 series Council for Geoscience). The percussion samples analysed by Smit for his map led him to conclude that areas of Kraaipan and Dwyka lie beneath the Kalahari Group sediments. The areas of Dwyka shown on his map could probably be reinterpreted as impact breccia, with the Kraaipan identified by him possibly representing the mafic components of the impact melt (M.A.G. Andreoli, *pers. comm.*).

3.8 Cretaceous Sediments

Cretaceous sedimentary rocks cover a large area of the sub-Kalahari surface, particularly in the northern parts of the Kalahari Basin. Sandstones and siltstones of the Kwango, Calonda and Continental Intercalar Formations extend over much of eastern and central Angola (see sub-Kalahari geological map, Appendix C) and the Democratic Republic of Congo (Giresse, in press). In Namibia, Etjo Formation and other Cretaceous sandstones occur to the south and southeast of Grootfontein, and in Zimbabwe sandstones of the Gokwe Formation are found on the edge of the Kalahari Basin.

The Gokwe Formation sandstones are separated from the overlying Kalahari Group sedimentary rocks by a prominent erosion surface, and are divided into a lower Calcareous Member consisting of well bedded thin units of calcareous sandstones, granule or pebble conglomerates, and pseudopisolitic limestones, and an upper White Sandstone Member composed of clayey sandstones, characterised by red clay pellets (Sutton, 1979). The sediments of the Gokwe Formation are believed to have been deposited in a shallow basin of alkaline water that existed in western Zimbabwe during the Cretaceous (Sutton, 1979).

The Kwango Formation consists of a basal conglomerate containing Late Jurassic or Early Cretaceous vertebrate remains. This is overlain by a fine, argillaceous sandstone or siltstone, which in turn is overlain by a 100 m thick, coarser-grained sandstone, showing some evidence of aeolian accumulation (Giresse, in press).

Cretaceous basalts are recorded in Angola (see sub-Kalahari geological map) where they occur in a northwest-trending zone. Their distribution is possibly related to reactivation of structures controlling the distribution of the Cassanje Group sedimentary rocks in northwest Angola.

3.9 Conclusions

As this chapter has shown, the basement to the Kalahari Group represents a geological history stretching from the Archaean, involving major episodes of tectonism in the Proterozoic, including the Pan-African events, and culminating in the Cretaceous with the intrusion of kimberlites into the Karoo Supergroup rocks and the deposition of sediments in the northern parts of the basin.

Geological and tectonic evolution of the region over the past 3000 Ma resulted in a diverse and sometimes complex geological setting, superimposed onto which was the formation of the Kalahari Basin and the deposition of the Kalahari Group sediments which began near the end of the Cretaceous and beginning of the Cenozoic, some 70-65 million years ago.

CHAPTER 4: DISTRIBUTION AND LITHOSTRATIGRAPHY OF THE KALAHARI GROUP

4.1. Introduction

Although the absence of fossils from the basal Kalahari Group sediments means that the age of initial deposition of Kalahari Group sediments is unknown, a maximum age for the deposition is provided by underlying Late Cretaceous kimberlite pipes, some of which have only been partially eroded (Hawthorne, 1975). Although later uplift events have modified the basin floor and edges of depocentres, an examination of the isopachs of the Kalahari Group sediments nonetheless suggests that initial deposition may have occurred in a number of separate depocentres, with the deposition of the Kalahari Group sedimentary rocks controlled by dominant northwest and northeast-trending structures. Seven main depocentres are recognised in the basin (Fig. 4.1).

The great thickness of sedimentary rocks in an “Angola” depocentre in central Angola is based on data taken from the published 1: 1000 000 geological map of Angola. It is not known how accurate these values are, nor is it known if pre-Kalahari rocks may have been erroneously included in the thicknesses shown on the 1: 1000 000 geological map of Angola. As some great thicknesses of Kalahari Group rocks do occur in this sub-basin, it can, however, not be completely ignored, but further investigation is needed before any interpretations can be undertaken.

The thickest accumulation of Kalahari Group sedimentary rocks in the entire basin covers a large area spanning the border between Namibia and Angola. In this “Etosha” depocentre Kalahari Group sedimentary rocks reach over 400 m in thickness (Appendix B) and have been well described in oil exploration borehole logs. Unfortunately a lack of reliable borehole data from southwestern Angola does means that it is unclear if this basin is part of a larger structure extending to the northeast, although massive thicknesses of Karoo Supergroup sedimentary rocks underlying this depocentre suggests that reactivation of the same structures controlling Karoo deposition may have occurred. Extending towards the southeast from this depocentre is a deep trough of Kalahari Group sedimentary rocks. This “Bushmanland-Hereroland” depocentre is

flanked on its eastern and western sides by exposed areas of sub-Kalahari rocks, and is terminated on its southern side by the uplifted Ghanzi Ridge.

On the eastern side of the basin is a trough of thick Kalahari Group sedimentary rocks extending roughly north-south from Zambia down into Botswana. This trough extends to the southwest towards Okavango Delta, where over 300 m of sediments have accumulated. To the southeast, the Makgadikgadi Basin is filled with over 100 m of Kalahari Group rocks.

Evidence presented in following sections suggests that the southern part of the basin may have initially been separated from the north by the Bakalahari Schwelle. To the south of the Bakalahari Schwelle up to 270 m of Kalahari Group sediment accumulated in the Aranos area of southeastern Namibia. This area also corresponds with great thicknesses of Karoo Supergroup and Nama sediments. A large amount of sediment was deposited along the South African-Botswana border and 180 m of sediment was deposited in the valleys surrounding the Morokweng Impact Structure. The extent of the impact melt shown on the sub-Kalahari geological map (Appendix C) possibly corresponds with a topographical high present during Kalahari Group deposition. Ridges of Olifantshoek and Transvaal Supergroup rocks flanked valleys where more Kalahari Group sediment was deposited.

The topography of the Kalahari-covered areas is, on the whole, very flat, and being a semi-arid to arid area, few rivers have cut down in to the Kalahari Group sedimentary rocks. Consequently, few exposures of the lower Kalahari Group formations exist, and interpretations have to be made from borehole data, mine exposures and the few river valleys where the lower formations are exposed. The problems associated with attempting to define a stratigraphy for the Kalahari Group are complicated by the fact that deposition in a number of small sub-basins, some fault-bounded with syn-depositional movement, has resulted in a lateral inconsistency of lithological units. In spite of this, however, it is possible to recognise a general sequence of lithological units for the Kalahari Group, with basal gravels being followed by clay, sandstone, unconsolidated sand and pan sediments and diatomaceous deposits. Duricrust formation is found throughout the southern parts of the basin with older lithologies commonly having undergone some degree of calcretisation or silcretisation. The main lithological units comprising the Kalahari Group will be described from the base of the succession upwards before the formal stratigraphy is discussed.

4.2. Lithologies

4.2.1 Basal conglomerate and gravel

Conglomerates and gravels at the base of the Kalahari Group have been recorded throughout the Kalahari Basin. Basal gravels have been recognised and described in Botswana by Passarge (1904) and Mallick *et al.* (1981), in Zambia by Money (1972), in Namibia by Mabbutt (1955), Albat (1978) and Miller (1992a), in the Democratic Republic of the Congo by Cahen and Lepersonne (1952) and in South Africa (Smit, 1977). The thickest basal deposits generally seem to be found along the southern edge of the Kalahari Basin in South Africa and in Namibia, where the gravels can reach thicknesses of up to 90m at the Uraninab Escarpment (duToit,1954; Mabbutt, 1955), and up to 120m in South Africa. In Botswana the gravels were considered by Boocock and van Straten (1962) to be of only minor importance. However, boreholes sunk during the Kalatraverse One project to the southwest of Makgadikgadi revealed between 4 and 40 m of basal Kalahari Group sedimentary rocks including units of sandstone conglomerate (Coates *et al.*, 1979). The gravels may have either a clay (Smit, 1977), sandy (Albat, 1978) or calcareous sandstone (Range, 1912), calcrete (eg Rogers, 1936; Mallick *et al.*, 1981) or silcrete (Shaw and de Vries, 1988) matrix.

The composition of the gravels reflects regional source area bedrock characteristics. In the Uraninab Plateau deposits in Namibia the gravels comprise granite, gneiss, quartz and Fish River Sandstone pebbles (Range, 1912; du Toit,1954) and in Bushmanland, Namibia, clasts are derived from nearby Damara carbonates. In the lower Molopo area local Dwyka pebbles are present (Rogers,1936), in Botswana near the Makgadikgadi Basin the clasts are derived from the underlying Karoo Supergroup (du Plessis and Le Roux, 1995), and in the open-pit at Sishen Iron Ore Mine most of the pebbles consist of quartzites from the Olifantshoek Supergroup (see section 4.3.1.1).

The gravels commonly lie directly on pre-Kalahari rocks and were often deposited in ancient river channels or grabens (Jones,1982). The mechanism of deposition of the basal gravels is unclear, largely due to the fact that few good exposures of basal gravels exist. Mabbutt (1955) noted that a coarse basal gravel interbedded with sands displayed evidence of current bedding, while DuToit (1954) regarded the grit-gravel-limestone (calcrete) unit to be of fluviatile origin. The basal gravels observed at Sishen as part of this study suggest fault controlled scree type deposits or debris flows that underwent some fluvial reworking. These are described further in section 4.3.1.1.

4.2.2 Clay

Calcareous clays or marls are present in the southwestern part of the Kalahari Basin (Boocock and Van Straten,1962; Smit, 1977) where they were seen as the most important part of the Kalahari sequence by Boocock and Van Straaten (1962). They are also present in the Grootfontein area of northern Namibia (Thomas,1988b), where they consist of alternating bands of clay and sand. They have not, however, been recorded in Zimbabwe or Zambia.

Calcareous clays may be found either overlying the basal gravels, resting directly on the pre-Kalahari surface (Boocock and Van Straten,1962), or even beneath gravel beds (du Toit; 1907,1954). Although impersistent in their distribution, the pink to red clays described by both Rogers (1936) and Boocock and van Straten (1962) can reach 65m in thickness, and those from the northern Cape reach 100m (Smit,1977). The Kalahari Group clays are generally fine-grained, homogenous and without stratification (du Toit,1954) and while commonly occurring as one bed, have been noted to inter-digitate with one or more silcrete or calcrete layers in some locations. In some localities where the clays are exposed, mottling and leaching as a result of changing groundwater levels may occur, and leaching and bleaching around roots also occurs.

The percentage of sand and silt in the clays can vary with locality (e.g. The Molopo Farms area, Gould *et al.*, 1987), and in some areas of South Africa it is difficult to distinguish between the clays and the overlying Eden Formation sandstones and siltstones.

The environment of deposition of the clays is not clear, although it has been observed in the southwestern Kalahari that the clays of the Budin formation do not extend beyond the confines of the palaeo valleys (Botha *et al.*, 1986). Bootsman (1998) described the blocking of the southward flowing proto-Molopo river in the late Cretaceous, providing the lacustrine setting conducive to the deposition of clays, and suggests that the clay deposits of the southern Kalahari were deposited in a lacustrine environment which may have extended about 400 km in a northeasterly direction. It is most likely that all of the clay deposits were deposited in similar settings, with large, shallow lakes filling up with fine sediments, and occasionally influxes of coarser material causing the interfingering observed in some deposits. In some areas, however, the clays have been produced by in situ weathering of underlying beds (Farr *et al.*, 1981; Bootsman, 1998).

4.2.3 Sandstone

Many of the sandstones described in the literature are either calcretised or silcretised, often resulting in them being mistakenly labelled as quartzites. Where calcretisation or silcretisation is absent, the sandstones may be poorly consolidated, and weather easily under the calcretised zones. The sandstones are commonly yellow, but red, brown and green varieties also occur. Unless silcretised or calcretised, the sandstones are poorly consolidated and may display a gradational contact with the underlying clays where they are present. They may also occur as lenses in the clay (Smit,1977) or in troughs on the pre-Kalahari surface (Levin,1980). The sandstones may often contain lenses of clay, grit or pebbles and gravel layers have also been reported at the base of the calcretes developed at the top of sandstone beds (Rogers,1936). In the case of the Letlhakane Stone-Line (section 4.3.1), these layers may even be assigned a formal stratigraphic position (du Plessis and Le Roux, 1995). The origin of stone lines is uncertain, but studies in central Zimbabwe have concluded that the stone lines found in the unconsolidated sands there are depositional in origin and consist of buried lag gravels deposited during periods of erosion of the finer material (Stocking, 1978). This theory seems consistent with the observations of various gravels layers in the upper part of the Kalahari sequence. Primary sedimentary structures are generally absent or rare in the sandstones, although M.A Thomas (1981) recognised cross bedding in the sandstones and gritstones outcropping in the southwest of the basin, and Money (1972) describes the Kalahari Group sandstones and quartzites in Zambia as having well developed, steeply dipping cross strata, which he interpreted as being indicative of aeolian origin. In the Congo Basin the sandstones of the Grès Polymorphes occasionally exhibit cross-bedding and the good sorting of the grains and grain shapes suggest at least one phase of aeolian transport (Giresse, *in press*). In general, however, the often coarse texture and intercalated pebble layers found in the sandstones throughout the area suggest deposition in fluvial systems, perhaps in a braided stream environment (e.g. M.A. Thomas, 1981; Malherbe, 1984; Botha *et al.*, 1986).

4.2.4 Duricrusts

Duricrusts are one of the most widespread lithologies in the Kalahari, and in much of the Kalahari Basin, other than the unconsolidated sands and pan deposits, duricrusts are the only Kalahari lithologies that can be seen on the surface. Figure 4.2 shows the distribution of pedogenic duricrusts in southern Africa, after Botha (2000). The names calcretes, silcretes ferricretes were first used by Lamplugh in 1902 and 1907 to describe the dominant types of duricrusts. The distribution of ferricretes in the Kalahari may be more extensive than is suggested by this map, but ferricretes are not as common as the calcretes, which may form composite horizons of up to 100 m in the Kalahari. These calcrete horizons are some of the thickest in the world (Shaw and de Vries, 1988). Outcrops of duricrusts occur as cliffs along the river valleys and around major pans, and are particularly prominent in the south of the basin along the Molopo River and its tributaries.

4.2.4.1 Ferricrete

Ferricretes, although reported from eastern Botswana (Kreimeyer *et al.*, 1990; du Plessis, 1993), western Botswana (Litherland, 1982) and Zimbabwe (Munyikwa *et al.*, 2000) are not as common as some of the other duricrust types and very little work has been conducted on them in the Kalahari Basin. They are commonly developed over a wide variety of pre-Kalahari Group rock types, but can occur as layers in unconsolidated Kalahari Group sands as is the case in western Zimbabwe (Sutton, 1979; Munyikwa *et al.*, 2000). In eastern Botswana, ferricrete is found capping pre-Kalahari Group bedrock (Litherland, 1982). Ferricrete generally forms when iron oxide is leached from an iron-bearing source, and is then precipitated in the soil or sediment (e.g. du Toit, 1954). In most recorded cases in the Kalahari, the ferricretes take the form of a limonite-cemented gravel.

4.2.4.2 Calcrete

Calcretes are the most common duricrust found in the Kalahari Basin, with almost all Kalahari Group sediments having undergone some degree of calcretisation in the past. In many areas of the Kalahari where unconsolidated sand covers all other lithologies calcretes are used as a source of road construction materials.

4.2.4.2.1 Definitions

Calcrete is defined by Netterberg (1980, p256) as "including almost any material of almost any consistency and carbonate content formed within the regolith by the *in situ* cementation and/or replacement of a pre-existing material by carbonate precipitated from the soil water or ground water." Wright and Tucker (1991, p1) defined calcrete as "a near surface, terrestrial, accumulation of predominantly calcium carbonate, which occurs in a variety of forms from powdery to nodular to highly indurated. It results from the cementation and displacive and replacive introduction of calcium carbonate into soil profiles, bedrock and sediments, in areas where vadose and shallow phreatic groundwaters become saturated with respect to calcium carbonate."

4.2.4.2.2 Chemistry of calcretes

Goudie (1972) found that the average composition of calcretes is 79.28% $CaCO_3$ (42.62% CaO), 12.30% SiO_2, 3.05% MgO, 2.12% Al_2O_3 and 2.03% Fe_2O_3. These compositions do, however, vary considerably, with the content of CaO in calcretes dropping to around 35% in north America or even as low as 9.68% in Australia (Dixon, 1994; Nash and Shaw, 1998) and there is no clear cut-off point in a calcrete composition that enables it to be defined as a calcrete on the basis of its geochemistry. Nash and Shaw (1998) suggest that an arbitrary boundary of 50% $CaCO_3$ be used, with a duricrust with $CaCO_3$ content above this value being considered a calcrete.

4.2.4.2.3 Ages of calcretes

Netterberg (1969) defined the age of a calcrete as the age of the first onset of calcification of the host material. In addition to this, however, the age of each variety of calcrete can be defined as " the age of the onset of calcification converting it from a less mature to a more mature state" (Netterberg,1978, p381). Hence as the calcrete develops and new phases of calcification occur, it will in fact become "younger".

Although the dating of calcretes may be complicated by the fact that there may be several different ages of authigenic carbonate within any one sample, Netterberg (1978) points out that selective sampling, particularly of a particular part of an authigenic phase, can enable the calcretes to be dated. The calcretes in the Kalahari were dated through their association with fossils and artefacts and divided into 5 age groupings, which can also be used as a classification system (Netterberg,1969,1978,1980).

i) Pre-Pliocene calcretes

Calcrete materials found intercalated in Kalahari beds are thought to be pre-Pliocene in age, and younger than the oldest Kalahari sediments Netterberg (1969).

ii) Pliocene Calcretes

The “Kalahari Limestone” referred to in some literature is probably middle-Pliocene in age, but may even be lowermost Pleistocene (Netterberg,1969).

iii) First Intermediate calcretes

These calcretes are mainly calcified alluvial gravels and sands containing Acheulian tools, along with some which contain Middle Pleistocene mammalian fossils.

iv) Second Intermediate calcretes / Upper Pleistocene

These calcretes contain Middle Stone Age tools. Calcretes from the Kwihabe Valley of northwestern Botswana were dated as having formed at 45 000 to 22 700 and11 000 to 9 800 years ago (Cooke and Verhagen, 1977).

v) Recent calcretes

Occasional hardpan calcretes containing Later Stone Age implements are known.

4.2.4.2.4 Classification of calcretes

Various classifications have been devised for calcretes, with perhaps the most basic classification scheme involving the grouping of calcretes into those containing gastropods, and those that contain none (Rogers, 1936). Different classifications have subsequently been developed with different applications in mind, and civil engineers, geographers or pedologists may all find different classification systems more useful or applicable. Perhaps the most useful classification system for geological field observations is the descriptive classification proposed by Netterberg (1980), based on secondary structure and the following generalised sequence of calcrete development and formation:

nodular calcrete → honeycomb calcrete → hardpan calcrete

Netterberg (1980) expanded on this simplified evolution (Fig. 4.3) and classified calcretes, into 7 stages or forms:

a) Calcareous soil
b) Calcified soil
c) Powder calcrete
d) Glaebular calcrete
e) Calcrete pedotubules, cutans and other structures
f) Honeycomb calcrete
g) Hardpan calcrete

In addition to this, descriptive terms including moisture content, colour, consistency and hardness and weathering would be used in conjunction with the above subdivisions.

Without contradicting this classification, calcretes can also be divided into: pedogenic and non-pedogenic calcretes, and, perhaps a third subdivision covering the *in situ* cementation of eroded, transported and redeposited remnants of older calcretes. However, caution should be taken when attempting to classify calcretes in this way. Pedogenic and non-pedogenic calcretes may occur in the same area, and pedogenic calcretes may develop in sediments that have previously undergone groundwater calcretisation, and may even grade laterally into non-pedogenic groundwater calcretes (Khadikar *et al.*,1998).

4.2.4.3 Silcrete

Silcretes have been recorded and described in numerous localities throughout the Kalahari Basin where they have often been described as quartzites or ortho-quartzites. Silcretes can, however, be distinguished from orthoquartzites on the basis of their porphyroblastic rather than equigranular texture (Hutton *et al.*, 1978, in Nash and Shaw, 1998).

4.2.4.3.1 Definitions

According to Summerfield (1983a) silcrete is the name given to siliceous duricrusts that contain >85% weight SiO_2. Mabbutt (1977) described silcretes as: "Siliceous crusts range from silica-cemented sands and gravels to an amorphous matrix of microcrystalline silica with floating quartz grains." Milnes and Thiry (1992; in Nash and Shaw, 1998, p14) say that "Silcrete is regarded as a product of the cementation or replacement of surficial materials such as rocks, sediments, saprolite or soils by various forms of secondary silica, including opal, cryptocrystalline quartz or well-crystallised quartz."

4.2.4.3.1 Occurrences

The full extent of the distribution of silcretes has possibly been underestimated, and silcretisation of Kalahari sediments, and in particular of older duricrusts, took place over large geographical areas in the Kalahari. The silcretes in the Kalahari commonly either outcrop on valley sides where they may be associated with calcretes, or occur in pans where green silcrete layers are found interbedded with calcrete (e.g. Litherland, 1982) or associated with siliceous rhizoliths (Botha, 2000). Green silcretes have been described at Heuning Vlei Pan in the Northern Cape Province of South Africa, where they consist of a framework of about 25 % subrounded quartz grains, with a few grains of microcrystalline quartz mass (Smale, 1973). Near Makgadikgadi, on the edge of a pan, similar silcretes consist of a quartz sandstone with 70 % subrounded to rounded quartz grains and a matrix of brownish, nearly isotropic clay (Smale, 1973). In northeastern Namibia, weathering of a greenish silcrete has resulted in a brownish colour caused by replacement of silica by Fe minerals (Albat, 1978).

4.2.4.3.3 Chemistry of silcretes

Silcretes may vary substantially, both petrographically and geochemically, and Summerfield (1983b) attributes the morphological and mineralogical characteristics to control by:

a) the geochemical environment during initial silica precipitation

b) host sediment characteristics (also seen as important by Nash *et al.,* 1994c)

c) post silicification diagenesis (not commonly observed)

Silcretes should contain at least 85% silica (Summerfield, 1983a), and silcretes with 99.49% SiO_2 have been reported from New South Wales, Australia (Callender, 1978; in Nash and Shaw, 1998).

Summerfield (1983b,c) distinguished between those silcretes associated with deep-weathering profiles, and those that are not. Silcretes found in the Cape coastal zone belong in the former category, and result from a process of prolonged leaching in a weathering environment. The silcretes of the Kalahari formed through the silicification of the host material, and belong to the non-weathering profile type. The so-called "weathering profile" silcretes of the Cape coastal zone contain glaebular aggregates and colloform structures while these petrographic features are absent in the "non-weathering profile" silcretes of the Kalahari (Summerfield, 1983c). The two types of silcretes are very similar chemically, with the only significant chemical difference between the two being the amount of TiO_2, which has significantly higher concentrations in the silcretes of the Cape coastal zone. Usage of the terms "weathering" and "non-weathering" types should be used with caution however, as Nash *et al.* (1994c) report on the occurrence of silcretes of the "non-weathering type" associated with deeply weathered landforms.

4.2.4.4 Formation of duricrusts

The most important pre-requisites for the formation of silcrete or calcrete are a source of silica or calcium carbonate, and mechanisms for transporting and precipitating them (Nash *et al.*, 1994b). The transport of the SiO_2 or $CaCO_3$ can occur vertically or laterally (Goudie, 1983), but in many cases a combination of both mechanisms may occur, with vertical transfer occurring following the lateral transfer of solutes (Nash *et al.*, 1994b).

Lateral transfer mechanisms occur where there is some kind of hydrological gradient (Nash *et al.*, 1994b). Duricrusts may form due to deposition in valley floors, deposition from sheet floods and lateral movement of groundwater, with deposition of SiO_2 or $CaCO_3$ at or beneath the surface of the soil or sediment (Goudie, 1983; Summerfield, 1983c), and may also form in pans where high potential evaporation rates and fluctuating pH levels cause weathering and leaching of clay minerals, thereby providing a source of SiO_2 or $CaCO_3$ (Summerfield, 1983a,b,c), with the abrasive action of wind resulting in a massive accumulation of silica dust.

Laminar silcretes in Lebatse Pan in southern Botswana are thought to have formed at the junction between underlying saline water and overlying standing water (Holmgren and Shaw, 1997). Modern silcretes are forming today on the saline surface of the Makgadikgadi pans (Thomas and Shaw, 1991a), and are also forming on islands in the lower Okavango due to the evaporitic concentration of solutes (McCarthy and Ellery, 1995). McCarthy and Ellery (1995) propose a model for silica precipitation from river water in a channel in the Okavango Delta. In their model they suggest that transpiration of aquatic grasses has caused the precipitation of fine-grained amorphous silica from groundwater, and that this has happened in conjunction with the deposition of clastic material and phytolithic silica from the river water. This has resulted in silica cement.

The Kalahari calcretes were thought to be dominantly pedogenic (Watts,1980), with their formation dominated by vertical transfer mechanisms. The calcretes are believed to have formed as horizons of lime enrichment develop towards the lower limit of leaching in soils or sediments, with a downward decrease in compactness. Depth of calcification and development of calcrete diminish with rainfall. The early formation of nodules and powdery horizons are influenced by the texture of the matrix with more rapid development in coarse, more permeable sediments. A hardpan may then form at this level, forming an impermeable layer upon which further accumulation of calcrete occurs (Mabbutt, 1977).

The main source of lime for these pedogenic crusts is from solution after extensive shallow flooding or from laterally migrating soil water. An alternative source is wind-blown calcareous dust which lies long enough for rain to leach carbonates downwards into the soil (Mabbutt, 1977; Khadkikar *et al.*, 1998).

The formation of duricrusts can be influenced by the geomorphology of an area, with the type and morphology of duricrusts shown to be linked to valley formation, and more specifically, through groundwater movement into the dry valleys (mekgacha) of the Kalahari (Nash *et al.*, 1994b). Their studies in the Rooibrak Valley of central Botswana and at Letlhakeng in southwestern Botswana, show that the duricrusts in these areas probably owe much of their development to the presence of valleys at that locality, and do not continue laterally away from the valleys in any great thickness.

The formation of duricrusts in the Kalahari therefore seems to be due to a combination of lateral and vertical transfer mechanisms, with lateral transfer mechanisms being of greater importance along the dry river valleys and pans. Substantial thickness of calcretes away from valleys, for example in the Kalahari Gemsbok Park, suggests, however, that duricrust formation is not limited to the sides of valleys, and duricrust horizons form a recognisable layer in some parts of the basin, with the southwestern Kalahari being a notable example.

4.2.4.5 Silcrete-calcrete intergrade duricrusts

Three types of intergrade duricrusts can occur: duricrusts where secondary silica occurs within a primary calcareous matrix; varieties where secondary carbonate occurs within a dominantly siliceous matrix; and crusts where the precipitation of carbonate and silica has been contemporaneous (Nash and Shaw, 1998). The naming of an intergrade duricrust depends on the proportions of $CaCO_3$ and SiO_2 in the rock, with a rock containing, for example, over 65% $CaCO_3$ and 30% SiO_2 being termed a sil-calcrete (Nash and Shaw, 1998), whereas to be a silcrete, a minimum content of 85% SiO_2 should be present (Summerfield, 1982). Where silica and carbonate cements occur in similar quantities, the relationships between the silica and carbonate should be examined in order to determine the evolutionary history of the duricrust (Nash and Shaw, 1998).

In the Kalahari, calcretes are commonly silicified, with the processes of replacement of silicate minerals by calcite resulting in the release of silica (Watts, 1980). Authigenic silica in Kalahari calcretes can occur in void and vein fills and by replacement of the matrix where there are no voids (Nash and Shaw, 1998). Silcretes can also be replaced by carbonates, where it usually occurs as a late stage void or vein fill (Nash *et al.*, 1994c; Nash and Shaw, 1998).

In high pH conditions silica remains in solution, with calcite being precipitated, whereas in low pH conditions the converse happens, with silica being precipitated (Goudie, 1983). In highly alkaline environments, where the pH is greater than 9, it is however possible that calcite and silica could be simultaneously precipitated in localised zones across a pH gradient (Summerfield, 1982;1983b). Complex intergrade duricrusts where SiO_2 and $CaCO_3$ have been precipitated together have been reported from near the Okavango delta, where duricrusts from the Thamalakane River include pisoliths containing concentric laminae produced by alternating phases of SiO_2 and $CaCO_3$ precipitation (Shaw and Nash, 1998) and duricrusts from Sowa Pan consist of a matrix of almost equal proportions of SiO_2 and $CaCO_3$ surrounding detrital quartz grains (Nash and Shaw, 1998).

4.2.4.6 Palaeo-environmental significance of duricrusts

Partridge (1969) believed that because the formation of pedogenic duricrusts is influenced by local soil climate and parent material, particularly in instances where moisture content of the soil may fluctuate, they should not be awarded any climatic significance.

The palaeoenvironmental significance of silcretes is debatable with numerous factors known to affect both the solution and deposition of silica, including the many possible sources of silica, different modes of transportation, and different mechanisms of precipitation. Complicating matters further is the fact that many different ages of silcretes may exist within any one outcrop. Nash *et al.* (1994b) suggest that formation may have taken place under more humid conditions than those in which they are presently found. The same authors refer to several Australian studies which suggest swampy river flood plains and well drained upland soils as possible environments of formation, with the silcretes then being preserved under more arid conditions. This view is confirmed by Southern African researchers like du Toit (1954) who suggests silcrete forms in a poorly drained area, possibly with a fluctuating water table, and Watts (1977) who suggests that the early phase of silicification takes place under low-pH, moist climatic conditions, with later increased aridity resulting in drying out phases, increased subsoil pH, and an associated concentration of silicon and titanium. Summerfield (1983b,c, 1986) differentiates between the high TiO_2 weathering profile silcretes which he believes formed under more humid, low-pH conditions, and the low TiO_2, non-weathering profile types from the Kalahari which are interpreted as having formed in more arid, alkaline environments. The climatic interpretation of the Kalahari silcretes is, however, questioned by Nash *et al.* (1994c) who attribute the low levels of TiO_2 in Kalahari silcretes more to differences in host material than to climatic controls, in line with Partridge's (1969) criticisms.

In the case of calcretes, certain generalisations can be made, and pedogenic calcretes are typical of moderately arid to semiarid climates with between 100 and 400mm rainfall, particularly on the temperate margins of zonal deserts and extending into the mid-latitude deserts (Goudie,1983). Immature or poorly developed calcretes may develop in areas with less than 800 mm rainfall (Netterberg, 1969), but more advanced types of calcretes commonly only develop where rainfall does not exceed 500 mm (Watts, 1980) or 550 mm (Netterberg, 1969).

In spite of these general climatic constraints, however, the climatic significance of calcretes is still unclear. Calcrete formation and preservation is dependant on sometimes complex relationships between precipitation, evaporation, groundwater flow and the composition of solutes.

4.2.5 Unconsolidated Sand

The unconsolidated sands of the Kalahari Group cover an area of over 2.5 million km^2, stretching from the Orange River in the south as far north as the Democratic Republic of Congo, and are thought to form the largest continuous sand body on earth (Baillieul, 1975). The thickness of the unconsolidated sands varies across the basin, from a few centimetres to over 200m in the north of Namibia in an area adjacent to Etosha Pan (Miller, 1983). The dominant landform associated with the sands is the dune fields, the dating and study of which has provided abundant palaeoclimatic and depositional information. The sand dunes are covered by grass in the southern parts of the basin (Fig. 4.4), and in northern parts, may be covered by dense deciduous woodland. In the eastern and northern dune fields the dunes are commonly degraded, and dunes throughout the basin are largely stable, with only some dune surface aeolian activity in the southwestern part of the basin (e.g. Bullard *et al.*, 1996; 1997). The dunes are generally classified as relict- or palaeo-forms as dune construction itself is not currently taking place (Wiggs *et al.*, 1995; Thomas and Shaw, 2002).

4.2.5.1 Sand composition

The unconsolidated sands consist largely of medium- to fine-grained (2.5 -1.5∅)(Lancaster, 1986a), rounded to sub-rounded grains of quartz. Sutton (1979) found a consistent bimodal size distribution in the undisturbed Kalahari sands of western Zimbabwe, with the larger grains generally fairly well-rounded, and the smaller grains subrounded to angular. Sutton removed the iron oxide coating from the grains of sand by using hot hydrochloric acid and revealed that the surfaces of the rounded (larger) grains are generally frosted whereas the angular grains are generally glassy. Lockett (1979), working in the same part of the basin, distinguished two classes of sands on the basis of grain-size distribution and composition. He recognised that in addition to the quartz-rich bimodal sands described by Sutton (1979), pan depressions are characterised by unimodal, partly consolidated clayey sands. A strongly bimodal frequency distribution, thought to be a result of lag accumulation in wind deflated areas, was also recognised in the sands of western Zambia by Binda and Hildred (1973). In the southwestern Kalahari the bimodal nature of the unconsolidated sands has not been recognised. Recent studies along the length of a linear dune in the southwestern Kalahari show no significant grain-size variation (Livingstone *et al.*, 1999), with the only variation in dune-cross profiles from this area being that samples from the crest of the dune are better sorted then those at the base (Lancaster, 1986a; Livingstone *et al.*, 1999). The reason for the bimodal grain-size distributions of the sands in the eastern and northeastern parts of the basin may be due to the higher incidence of fluvially deposited sands in this area, as will be discussed in the following section.

Regional differences in colour may reflect different sand ages and sources. Redness of dunes tends to increase with age, either from the continuing accession of dust rich in pigment, or from the weathering of contained ferromagnesium minerals (Norris,1969). Stokes *et al.* (1998) report that the degree of post-depositional reddening of the sand increases from the south to the north of the basin, possibly reflecting a N-S directed relative dune chronosequence. In the southwestern Kalahari the sands are generally yellowish-red with no distinct colour differences between dune-crests and interdune areas. There are exceptions to this, however, and in the Mafungabusi area of Zimbabwe, leached pale yellow or white sands have been reported at the margins of dunes and at the head of vleis (Sutton, 1979) and the lunette dunes next to the pans of the central and southern Kalahari are commonly grey-white in colour. The occurrence of the pale yellow or grey sands is thought to be due to the removal of a thin layer (± 0.02mm) of iron oxide by saltation in mobile sands (Mabbutt, 1977) and may be limited to the surface in some instances. Sutton (1979) reports that as little as 2 metres below the surface the pale sands show streaks of red and orange.

The sands of Zimbabwe have a high abundance of ilmenite (Lockett, 1979; Sutton, 1979) with much of the other opaques consisting of iron and titanium (Sutton, 1979). Zircon, staurolite, rutile and tourmaline have also been panned from nine samples from western Zimbabwe (Lockett, 1979), while Moore and Dingle (1998) found that tourmaline, staurolite and kyanite were the most common heavy minerals found in the sands sampled in a large area of central Botswana.

4.2.5.2 Origin of the unconsolidated sand

Unconsolidated sand deposits are underlain by a large number of lithologies which could weather into sand-sized particles and Kalahari Group and Karoo Supergroup sandstones cover a large percentage of the sub-Kalahari surface (see sub-Kalahari geological map, Appendix B). The theory that in situ subsurface bedrock weathering and bioturbation has made significant contributions to the sand deposits has been proposed by Wellington (1955) in the Democratic Republic of Congo, Boocock and Van Straten (1962) in the southern Kalahari and Baillieul (1975) in Botswana. More recent work done in the sandplains of western Australia showed that a strong relationship between sand distribution and bedrock type occurs, and the sandplains themselves are believed to be a product of in situ weathering of underlying sandstones (Newsome, 2000).

Compositional variations in the unconsolidated sands suggest that some transportation of the material has taken place. Poldervaart (1957) suggested that the decreasing proportions of kyanite and staurolite content in the sand, as well as an inverse relationship for zircon, from the western to the eastern side of the basin, implied a westerly source for kyanite and staurolite. Moore and Dingle (1998) arrived at similar conclusions, and interpreted staurolite and kyanite distributions as pointing towards a source to the west and northwest, in the headwaters of the Okwa and Deception rivers. Tourmaline distribution patterns in the sands of central Botswana were believed by Moore and Dingle to indicate a provenance corresponding with the shallow relief of the watershed between the Okwa and Molopo drainages, called the Kalahari or BaKalahari Schwelle

(Passarge, 1904). Moore and Dingle (1998) also suggest a source of sand on the northeastern rim of the southern Kalahari depocentre, and Bond (1948) also believed the sands in western Zimbabwe originated from the north.

Baillieul (1975) recognised that the unconsolidated sands of the Kalahari could be divided into different types according to their origin and mechanism of deposition, and divided the surface sands of the basin into four types (Fig. 4.5). According to Baillieul (1975), the sands of eastern Botswana had a fluvial and reworked aeolian origin with local bedrock components providing regional variations. The northwest of the country is characterised by aeolian sand which in places is becoming mixed with sand derived from the Ghanzi Ridge towards the south. The southwestern Kalahari consists largely of well-sorted sands including a component from underlying Karoo sandstones.

Bond and Fernandes (1974) examined quartz grains from a number of localities in the Kalahari and found that the earliest texture to be seen on all of the grains is upturned plating of aeolian origin, with the grains having subsequently undergone surface reworking by wind and water. The aeolian characteristics of the sand grains may, however, have been inherited from an original source rock of aeolian origin (Lancaster, 2000), with aeolian processes only acting on the sand subsequent to deposition.

De Ploey *et al.* (1968) suggested that the sands of the Sables ochres of the Democratic Republic of the Congo were deposited by fluvial processes as did Verboom (1974) for the unconsolidated sands in western Zambia. Moore and Dingle (1998) presented evidence from central Botswana suggesting that the textural and heavy mineral distribution patterns are at variance with aeolian deposition controlled by the dominant easterly to northeasterly prevailing winds. The same authors believe that the patterns are more readily explained by transport into the central basin by sheet wash and ephemeral streams rising off a provenance area of low relief, and that aeolian processes may have only resulted in a local secondary imprint on an essentially fluvial sedimentary sequence during recent arid episodes.

Erosion of fluvial or lake deposits by wind and their deposition in a adjacent area is a common source of dune material, with the sand and sediment eroded from pans being deposited as lunette

dunes on the downwind side of pans or playas (e.g. Young and Evans, 1986), and fine-grained sediment from pans forming an important component of fine-grained aeolian sediment that may be transported substantial distances (e.g. Middleton, 1997). The relationship between fluvial deposits and dunes is well illustrated in northern Namibia where linear dunes occur on the downwind side of channel systems containing medium- to fine-grained sands (Thomas *et al.*, 2000), and in the southwestern Kalahari, where mixed sands sampled from interdune areas close to rivers and pans were found to be predominantly fluvial in origin, whereas the sands from dune crests showed signs of aeolian reworking and were commonly more fine-grained, better sorted and more finely skewed (Schlegal *et al*., 1989). The availability of fluvial sediment is of great importance to initial dune construction, with dune formation only occurring in areas where abundant sediment was available during dry periods (Thomas *et al.*, 2000) and dunes becoming vegetated and stabilised when a supply of sediment is cut off (Lancaster, 1990). According to Thomas *et al.* (2000) this may partly explain why in the southwestern Kalahari fewer periods of dune formation have been recorded than in northern Namibia and western Zambia where an abundance of deflatable material was available for new dune construction. In Zambia the Zambezi River itself was a major source of sand (e.g. O'Connor and Thomas, 1999). Discriminant analysis of grain-size, shape and mineralogy of the Kalahari sands by Thomas (1987) did show, however, that the sands are largely homogenous, and an important conclusion of this is that following deposition, the dominant influence on the unconsolidated sands has been one of aeolian activity.

4.2.5.3 Sand Dune morphology

Over 85 % of the dunes in the Kalahari are of the linear (alab or sief) variety (Fryberger and Goudie, 1981) and they occur in extensive dune fields throughout the basin. The dune fields of the Kalahari can be divided into Northern, Southern and Eastern dunefields (Fig. 4.6), with each dunefield thought to have formed at a different time based largely on a distinct orientation (Lancaster, 1980,1981; Thomas, 1984b; Thomas and Shaw, 1991a). The ridges of linear dunes are usually elongated in the general direction of the sand-transporting wind and their orientation is therefore thought to reflect the dominant wind direction during the time of their formation (Mabbutt,1977; Lancaster, 1980, 1984; Mallick *et al.*, 1981; Fig 4.7). According to Lancaster (1990), these patterns were probably caused by the increasing development and strength of anticyclonic circulations which also prevented moisture reaching the interior of southern Africa.

The northern dune field is characterised by broad linear dunes of approximately 25 m high and 1.5 to 2 km apart and extending as much as 200 km (Grove, 1969; Lancaster, 1980). Degradation of the dunes in this field increases towards the north (Flint and Bond, 1968; O'Connor and Thomas, 1999). The eastern field is characterised by degraded linear dunes with up to 50 km of unbroken length (Flint and Bond, 1968) and by barchan dunes and transverse dunes to the west of Makgadikgadi (Grove, 1969; Mallick *et al.*, 1981). The transverse dunes are N-S oriented, sinuous and bifurcating and between 10 and 20 m high (Mallick *et al.*, 1981). The dune fields in South Africa are especially prominent near the borders with Namibia and Botswana and fall within the southern dune field. The dunes in this area are characterised by partly vegetated linear dunes of 2-15 m in height, dune widths of 150-250 m (Lancaster, 1988, 2000) and are characterised by broad, interdune areas which are commonly grassed. In the southern field many of the dunes can be described as compound forms (Lancaster, 1987; Thomas and Martin, 1987; Bullard *et al.*, 1995) and in the Kalahari Gemsbok Park Goudie (1969,1970) recognised dendritic, reticulate and clustered dendritic forms, with small areas of hummocky or parabolic dunes possibly representing more recent reactivation (Eriksson *et al.*, 1989; Thomas *et al.*, 1997). A more detailed study by Bullard *et al.* (1995) divided the dunes of the southwestern Kalahari into two classes of simple dunes and three compound dune classes (Fig. 4.8). The entire dunefield was then divided by Bullard *et al.* (1995) into areas showing the dominant class of dune (Fig. 4.9), and this reveals a general trend from the northwest to the southeast of increasing dune pattern complexity. Dune pattern statistics applied to the dune field has suggested that dynamic equilibrium between environmental factors and dune pattern geometry was established prior to stabilisation of the dune field (Thomas, 1986). A later study on the relationship between the dry river valleys of the southwestern Kalahari and the dune fields revealed that in some areas the valleys may have had an effect on the wind regime and the sediment supply and that both of these factors appear to have influenced the morphology of the dune field (Bullard and Nash, 1998).

Lunette dunes on the downwind fringes of some of the pan depressions also occur within the main linear dune system (Thomas *et al.*, 1997) and these are formed as a result of deflation of the pan floors during arid periods (Lancaster, 1978 a,b). The dunes consist of either the pan sediment itself (Goudie and Wells, 1995), or material derived from sediment blown or washed into the pan from surrounding areas (Thomas *et al.*, 1993). The sediment making up the lunette dunes may have up to 12-14 % $CaCO_3$ (e.g. Lancaster, 1978b; Buch and Zöller, 1992). Many of the dunes

are composed of an outer sandy lunette thought to have formed during initial pan formation and an inner, more clayey lunette consisting of materials deposited on the pan floor (Lancaster (1978b). This is not always the case, however, with the outer lunette sometimes consisting of a greater proportion of fine material than the inner lunette (Goudie and Thomas, 1986).

4.2.5.4 Age of the sand deposits

Prior to the application of luminescence dating, very few absolute ages for the sands or the dune systems themselves existed and these were derived mainly using ^{14}C dating. Luminescence dates have now given us an insight into the timing of dune development in the Kalahari Basin. These luminescence dates generally give indications only of the more recent phases of dune formation and reworking, however, and do not necessarily give us a maximum age for the unconsolidated sands themselves. As already discussed earlier in this section, the initial deposition of the unconsolidated sand deposits probably occurred via fluvial action, therefore pre-dating the earliest dune formation. The ages of dune construction do, however, give a chronology of redeposition and stabilisation of the sand dunes and are an important source of palaeoclimatic data (see Chapter 6). Figure 4.10 summarises dune and aeolian sediment luminescence data (Thomas and Shaw, 2002).

The oldest sands were thought to be late Pliocene (Mabbutt, 1955; Maufe, 1930, 1939) to Pleistocene in age (Bond, 1948; Wayland, 1953; Flint and Bond, 1968). Helgren and Brooks (1983) suggested that the large linear dunes were formed during the Early Pleistocene or late Tertiary and Cooke (1980) suggested deposition of the sands during intervals of Lower-Mid Pleistocene aridity. Although several other dates for conditions favouring dune formation have been suggested by various other workers (e.g. Lancaster, 1981, Van Zinderen Bakker, 1982, Deacon *et al*., 1984; Heine, 1981, 1982, 1990) there were not many absolute dates for the formation of dunes other than a dune-base peat date of 19 680 ± 100 years BP in the Makgadikgadi (Thomas and Shaw, 1991a) and a powder calcrete within a dune near Etosha dated at 3 510 BP (Rust,1984). In the southwest of the basin, Lancaster (1988) obtained a radiocarbon date of 32 500 BP from calcrete pedotubules in unconsolidated sands in the lower Molopo valley, and Hövermann (1988) obtained a radiocarbon date of 22 240 BP for alternating layers of calcrete and aeolian sand in the Nossob River. In addition to this, Lancaster (1989) suggested periods of

dune formation occurred in the northern dune field prior to 32 ka, in the eastern dune field between 19 and 17 ka, and in the southern dune field between 10 to 6, and 4 to 3 ka. These radiocarbon dates from calcretised sands have, however, been questioned by Blümel *et al.* (1998) due to the probability of carbon exchange in the open systems of the sandy sediments.

The development of luminescence dating techniques not only allows the date of deposition of non-carbon-bearing sediments to be ascertained, but also allows the dating of deposition of detrital sediments older than 40 000 years old, and can provide depositional chronologies up to 800 ka (Stokes, 1999).

To the northwest of Kuruman in South Africa, luminescence dating on a sand sheet at Mamatwan Mine gave ages of 60.2 ± 2.5 and 58.6 ± 2.3 ka. These are the oldest recorded ages for the sands in the southern Kalahari Basin (Thomas and Shaw, 2002). Further to the west and northwest, luminescence dating has revealed that in much of the southern dune field two significant phases of linear dune development occurred, between about 30 and 23 ka and 17 and 8 ka ago (Stokes *et al.*, 1997a,b; Thomas *et al.*, 1997; Eitel and Blümel, 1998). Dating of lunette dunes in the southwest of the basin at Luitenantspan, Soutpan, and Koopan Suid gave predominantly Holocene ages, although four of 18 ages gave dates of between 17 and 11 ka (Lawson, 1998; Lawson and Thomas, 2002). Optically stimulated luminescence (OSL) dating by Thomas *et al.* (1997) on minor dune forms within the linear dune field reveals that Holocene dune building activity possibly occurred at 6 and 2 to 1 ka, but most of the linear dune activity had probably ceased by 9-8 ka (Blümel *et al.*, 1998). Following this, dune building activity may have continued around pans, and dates of lunette dunes at Witpan gave ages of 1.1-1.5 ka (Thomas *et al.*, 1998).

In the northern and eastern dune fields, south of the Zambezi River, OSL ages reveal that periods of dune building in the late Pleistocene occurred between 115-95, 46-41, 26-20, and 16-9 ka, with the relatively short arid periods favouring dune building separated by longer periods of greater humidity (Stokes *et al.*, 1997b, 1998). The oldest of these ages for a dune building episode comes from northwestern Zimbabwe, and coincides with a dune date from near Tsodilo Hills of 98 ± 9 Ma (Thomas *et al.*, 2003), as well as dates from two basal dune samples from the Victoria Falls area of Zimbabwe which yielded Thermoluminescence (TL) ages of 82 ± 8 ka and sands beneath a ferricrete layer which gave TL ages of 120 ± 15 ka and 96 ± 8 ka (Munyikwa *et al.*, 2000). In western Zambia dune formation occurred from 32-27, 16-13, 10-8 and 5 to 4 ka (O'Connor and Thomas, 1999), with the middle two periods coinciding with the last reworking of dunes in western Zimbabwe between about 16 and 10 ka (O'Connor and Thomas, 1999; Munyikwa *et al.*, 2000). Optically dated dunes in the Caprivi strip of northern Namibia suggest periods of dune formation at 48-41, 36-29 and 23-21 ka (Thomas *et al.*, 2000; Thomas and Shaw, 2002).

4.2.6 Pan sediments and Diatomaceous Deposits (Kieselguhr)

Most pans are filled by a layer of clayey sand or alkaline calcareous clays (Rogers, 1934; Boocock and van Straten, 1962) and are flanked by lunette dunes. The clays predominantly consist of smectite, illite and sepiolite, but glauconite has also been described at Heuningvlei Pan in the southern Kalahari Basin (Bühmann *et al.*, 1999). Lancaster (1978a) divided pan deposits in the southern Kalahari into an upper sandy phase composed of non-saline, slightly to moderately calcareous sands, clayey sands and sandy clays, and a lower clayey phase which itself is divided into an upper, fine-grained, moderately calcareous, saline member, and a lower sandy, highly calcareous, moderately saline member. According to Lancaster (1978a) the lower clayey phase is present at all pans, whereas the upper sandy phase in largely absent from the surface of bare, clay pans, and is more commonly found in grassed pans. The sequence of deposits identified by Lancaster (1978a) is interpreted as being formed as desiccation of the lakes filling the pans occurred. Lancaster (1978a) believed that the deposition of sandy clays occurred in the permanent lakes during periods of moderate runoff, with saline clays being deposited during periods of lower runoff and the sandy phase being deposited on the margins of the shrinking lakes as a wash deposit. As more of the pan floor was exposed, so deflation occurred, resulting in the formation of lunette dunes downwind of the depression as has been described in the preceding section.

In some localities in the southwestern Kalahari spring-fed tufas have formed on the margins of pans during periods where groundwater discharge was high (Lancaster, 1986b). These tufas may contain evidence of algal mats and stromatolites and may also be associated with calcified reed and root tubes (Lancaster, 1986b).

Many pans in the Kalahari are characterised by diatomaceous deposits. Diatomaceous earth, diatomite or Kieselguhr is a white or grey, porous, light-weight, fine-grained sediment that is chalky in appearance, and may be either stratified or massive. It is composed mainly of the fossilised skeletons of diatoms, which are microscopic, unicellular plants forming a class of algae (Kent and Rogers,1947; Pettifer, 1982), as well as spicules of sponges and grass skeletons (Rogers,1936). Each diatom comprises a tiny speck of protoplasm enclosed by a shell (or test) of opaline silica, and when the diatom dies, the insoluble siliceous shell settles on the bottom of the lake (Pettifer, 1982).

The conditions favouring diatom growth are outlined by Strydom (1998):

- low water temperatures of 3°-6°C which inhibit bacterial activity thereby preventing acidification due to decomposition, and contain abundant O_2 and CO_2 necessary for diatoms to survive
- slightly alkaline water conditions
- an adequate supply of soluble silica
- small amounts of lime and magnesia which are essential to diatom growth
- low phosphate and nitrate contents which encourage the growth of diatoms over other micro algae

Diatomaceous deposits can occur in many different forms and diatomaceous limestones, diatomaceous shales and mudstones have all been described from the Kalahari Basin. Diatomaceous earths were first described by Passarge (1904) along the Boteti River and in the bed of Lake Ngami in Botswana. They have subsequently been described as a frequent occurrence around the Makgadikgadi Basin (e.g. Rogers, 1936; Grove, 1969; Shaw and Thomas, 1988; Shaw *et al.*, 1997) and diatoms have been described in the terraces of the Molopo River (Rogers,1936), at a shallow depth related to old drainage alignments (Smit,1977), in the Kalahari sands themselves (Rogers,1936) and in and around pans (Kent and Rogers,1947)(Fig. 4.11).

Most diatomaceous deposits found in the Kalahari Basin have fairly high quantities of calcium carbonate, and can be described as calcareous diatomaceous earth or diatomaceous limestones (Kent and Rogers, 1947). The diatomaceous earths are often penetrated by vertical, root-like, sandy, calcareous concretions, thought to have formed by roots which later decayed and either

left cavities which remained empty or were filled by sand (Fig 4.12).

Worldwide, the most significant diatomite deposits were formed in the late Tertiary and Quaternary, and according to Thomas and Shaw (1991a) the proximity of diatomaceous earths to the surface suggest a relatively recent origin, probably during the last glacial. Kent and Rogers (1947) suggest that apart from a few deposits, which are of Recent and sub-Recent age, the diatomaceous earths date back to the so-called "Third Wet Phase" of the Pleistocene period in South Africa.

Diatomaceous limestones from the Bayip and Sewe Panne in the Kalahari Gemsbok Park (Malherbe, 1984) and from the farm Lonely 174 (M.A. Thomas, 1981) have previously been described and the species shown in Table 4.1 were recognised.

Table 4.1 Macrofossils, diatoms and ostracods found in the diatomaceous deposits at Lonely Farm on the Kuruman River and at Sewe and Bayip Panne in the Kalahari Gemsbok Park.

Lonely Macrofossils **M.A. Thomas (1981)**	**Lonely Ostracods** **M.A. Thomas (1981)**	**Lonely Diatoms** **M.A. Thomas (1981)**	**Kalahari Gemsbok Park Diatom genera Malherbe (1984)**
Corbula Africana	*Cypridopsis aculeata*	*Epithemia gibba*	*Epithemia*
Xeroceratus damarensis	*Cypridopsis elizabethae*	*Epithemia argus*	*Diplonais*
	Gomphocythere expansa	*Cymbella gastroides*	*Mastogloia*
	Gomphocythere obtusata	*Cymbella lanceolata*	*Pinnularia*
		Melosira cf roesana	*Surirella*
		Melosira distans	
		Diplonais sp.	
		Fragillaria spp.	
		Mastogloia sp.	
		Stauroneus sp.	
		Synedna sp.	

In addition to this when the samples were treated with hydrogen peroxide, charophyte stems and ooginia were liberated (M.A. Thomas, 1981).

4.3 The Stratigraphy of the Kalahari Group and regional stratigraphic variations

Figure 4.13 shows representative composite borehole logs from different localities in the Kalahari Basin. As can be seen, the stratigraphy does vary somewhat throughout the basin, and local variations in the space of even a few kilometres can be extreme. Various studies throughout the Kalahari Basin have been conducted over the years, with stratigraphies having been proposed for certain areas. In some cases these stratigraphies have been formally adopted for the countries in which they occur. In this section, the proposed Kalahari Group stratigraphies for each country will be discussed, with a correlation between them being attempted at the end of the chapter.

4.3.1 Stratigraphy of the Kalahari Group in South Africa

In South Africa the Kalahari Group subdivisions, proposed by Smit (1977) and approved by the South African Committee for Stratigraphy (SACS ,1980), consists of four main formations: The Wessels Formation, named after the basal gravel unit exposed in the Wessels Manganese Mine, Hotazel; the Budin Formation consisting of calcareous clays, from the farm Budin 495 on the Moshaweng River; the Eden Formation consisting of sandstones and thin gravel layers from the farm Eden 703/46 also on the Moshaweng River; and the Gordonia Formation which consists of the widespread unconsolidated sands. M.A. Thomas (1981) suggested four additional formations, the Goeboe Goeboe Formation, Obobogorop Formation, the Lonely Formation, and the Mokalanen Formation, based on observations made along the Molopo River and in the Kalahari Gemsbok Park. Figure 4.14 is a schematic stratigraphic column for the Kalahari Group in South Africa

The calcretes in the Kalahari Gemsbok Park and towards the east along the Molopo River form prominent cliffs between the aeolian sands of the Gordonia formation and the underlying Eden Formation sandstones. As these calcretes appear to occupy a fixed stratigraphic position in the area they were proposed as the Mokalanen Formation by M.A. Thomas (1981). The formation name has not yet been accepted by SACS, and the problems associated with assigning a fixed

stratigraphic position to calcretes is discussed at the end of this chapter. The Lonely Formation consists of diatomaceous limestone from the Kuruman River as well as from pans, and the Goeboe Goeboe Formation refers to the pan sediments themselves. The gravels of the Obobogorop Formation are thought to originate from the weathering of Dwyka tillite and clasts may vary in size from 2-25cm (M.A. Thomas, 1981). They are sub-angular and consist of a wide variety of rock types. The gravels may be found either resting directly on Karoo Supergroup rocks or even resting on older calcrete horizons. M.A Thomas (1981) suggests that these gravels may still be being produced today. The Formation status of the Obobogorop, Goeboe Goeboe, and Lonely Formations has not yet been approved by SACS.

Borehole data from across the basin has shown that the formations of the Kalahari Group are often not laterally continuous for any great distance and the stratigraphy of the Kalahari Group may vary substantially over fairly small areas. The basal gravels may be either absent completely, or may occur as several thin layers separated by clay-rich horizons. They may occur over several kilometres or are limited to the bottom of palaeo-valleys and their distribution and thicknesses are discussed further in Chapter 8. The clays and siltstones of the Budin Formation may grade laterally and vertically into coarser-grained siltstones and sandstones of the Eden Formation and calcretisation or silcretisation of any of the older lithologies may occur. As their deposition is thought to have been limited to lakes forming along the floor of the Kalahari Basin, their distribution is not as widespread as the overlying sandstones. A schematic cross-section through the Kalahari Group in South Africa attempts to illustrate the variability in the stratigraphy over a fairly small area (Fig. 4.15).

Kalahari Group lithologies were looked at in more detail in three areas of the South African portion of the Kalahari Basin as part of this study (Fig. 4.16). The field descriptions exclude the overlying unconsolidated sands, pan sediments and diatomaceous deposits, all of which have been well described in previous studies and with the results of previous work having been have been summarised in preceding sections.

4.3.1.1 Sishen Mine (Locality 1)

Sishen Iron Ore Mine is an open cast mine situated in the Northern Cape Province of South Africa. At Sishen, some 80 m of Kalahari Group sedimentary rocks have been excavated, with the sequence well exposed on the pit walls (Fig. 4.17). The Kalahari Group rocks lie unconformably on underlying lavas, shales and quartzites belonging to the Olifantshoek Supergroup. Two main lithologies can be distinguished in the pit, 1) basal gravels, which could be assigned to the Wessels Formation, and 2) overlying calcareous clays which probably belong to the Budin Formation. Within each of these, however, a number of units can be recognised, and for convenience the units exposed are described as Units 1-8 from the base to the top of the succession. Figure 4.18 is a simplified measured profile through the succession at Sishen.

Units 1 and 2

The base of the succession at Sishen consists of a thick gravel unit which can be subdivided into two sub units (Fig. 4.19). The lowermost portion of the bottom unit (Unit 1) is very poorly sorted with large boulders (up to 1,4 m) found alongside small (<1cm) rounded, spherical pebbles. The thickness of this unit varies between 0 and 5 m and is matrix supported (Fig. 4.20). The top of Unit 1 may consist of a 20-30 cm layer of smaller, rounder pebbles floating in a clay matrix. The base of the overlying Unit 2 consists of a layer approximately 50 cm thick, made up of large, angular clasts of up to 36 cm in size, and averaging 16 cm for their longest axis. Above this lower clast layer, the unit consists of small, sub-rounded pebbles and shows some upward fining. It is moderately sorted, has a polymodal size distribution and is clast supported. There is no sign of bedding or imbrication and it is therefore an unordered fabric. Unit 2 can reach up to 3 m in thickness.

The basal gravels of units 1 and 2 show no signs of calcretisation with no calcrete nodules occurring in the clayey matrix. Calcrete nodules were, however, observed in chips from a percussion borehole drilled into this unit to the west of the open pit, where calcretisation seems to have taken place at the interface between the bed rock and the overlying gravels. Clasts in both units consist largely of quartzite pebbles.

Borehole evidence suggests that the gravels may be very localised and in boreholes drilled 100 m

apart to the west of the pit at Sishen, evidence for a gravel layer only appeared in one out of four boreholes examined. Towards the south, Unit 1 becomes less distinct, with the density of clasts decreasing markedly. Continuing to the south Unit 2 is found lying directly on bedrock and then itself pinches out completely. At this locality the bedrock shows no evidence of great variations in topography and in the bedrock and within the gravels themselves evidence for channels is absent.

The poorly sorted gravels of Unit 1 exposed at Sishen may have been deposited in an alluvial fan, but the lack of evidence of significant fluvial reworking suggests that deposition by a debris flow appears most likely. Large scale faulting is known from the vicinity of the deposit, and it is likely that debris flows along fault scarps resulted in the deposition of these beds. The more rounded and better sorted gravels of Unit 1 possibly represent small channels on the surface of the debris flow, and some reworking of the basal gravel unit may have occurred in a similar fashion to that observed in Namibia by Albat (1978). The disappearance of many of the larger boulders found in Unit 1 towards the south and the pinching out of Units 1 and 2 in the same direction, suggests that the source of the clasts was towards the north where quartzites outcrop today.

Unit 3

At Sishen Mine red brown clays generally form a sharp contact with the underlying basal gravels, although within the lower 2 m of the clay deposits scattered pebbles are in the clay. Pebbles are also sometimes found occurring in small lenses, commonly 40 cm long, 5 cm thick and reaching several metres in length in places (Fig. 4.21). These lenses are matrix supported and the pebbles are 1-2 cm in size, and comprised of the same rocks as in the underlying gravel beds. The matrix consists of the same clay as above and below the lenses. Generally, however, there are few pebbles found in the clays.

Higher up in Unit 3, the clays are characterised by numerous thin, vertical streaks approximately 20 cm long and 2 cm wide, which appear to have no difference in composition to the surrounding red-brown clays and may be caused by bleaching or extraction of the hematite from the zones around roots with some calcite precipitation (Fig. 4.22). Calcified and silicified rhizocretions as well as calcrete nodules (Fig. 4.23) are also present and they weather out as positive features in the otherwise easily weathered and unconsolidated clays (Fig. 4.24). At the top of this streaky

zone a mottled zone 2-3 m thick occurs (Fig. 4.25). The zone is undulating, and although in some places is absent, it can be traced for a considerable distance around the pit face and is probably laterally very extensive. This zone appears to have been caused by a fluctuating water table at this level. Within the mottled zone numerous horizontal and vertical joint planes filled with calcareous material may occur, and there appear to have been several phases of infilling and replacement. There is evidence of slumping and movement along some of the joints, with up to 15 cm displacement of secondary infillings. Soft, irregular carbonate nodules occur in this horizon.

A sample from this unit was analysed and found to consist largely of smectite, with quartz making up a significant percentage of the sample (Table 4.4). The high percentage of quartz is confirmed in hand specimen and thin section, with the clays found to contain a fairly high amount of fine- to medium-grained quartz particles, possibly blown or washed into the depositional environment from surrounding areas. The sand grains are generally fairly well rounded and well sorted. The absence of calcite is surprising but may be due to the fact that calcareous nodules were not included in the sample.

Table 4.2 - XRD analysis of clay sample from Unit 3, Sishen Iron Ore Mine, South Africa

Mineral	**weight %**
Smectite/ Montmorillonite	33 %
Quartz	56 %
Hematite	7 %
Microcline	4 %

Bootsman (1998) collected two clay samples from the Molopo river valley about 90km to the west of Mafikeng, and a third sample from further south at the Mamatwan mine some 45 km to the north of Sishen. Although it is not certain if these samples were from the same clay layer as Unit 3, it is interesting that Bootsman also found that smectite/ montmorillonite was the dominant clay mineral in his samples.

Unit 4

In some areas of the pit, the mottled zone is overlain by 3-4 m of a reddish clay, which is capped by a dark red/brown clay that appears to have been deposited in channels or pans as Figure 4.26 suggests. In some parts of the pit it is either very thin or absent. The white material at the base of the channel is rich in calcium carbonate, and carbonate nodules occur along the base of the channel. This is due to groundwater preferentially moving along the base of the channel with calcium carbonate precipitated as a result of either evaporation or by carbon dioxide degassing (Khadkikar *et al.*, 1998).

Unit 5 and 6

The dark red/brown clay is overlain by thick layers of white clay, which in places has small clay pellets at its base, and appears to have been deposited in channels or pans (Fig. 4.27), suggesting an erosional contact with the underlying clays. The white clay may reach a total of 8 m in thickness.

The white clays can be separated into two units, with the upper unit (Unit 6) forming a sharp contact with the underlying Unit 5. Each unit varies in thickness but on average are about 4 m thick. In some localities the two units are separated by a layer of clay pellets occurring at the base of Unit 6. The clay pellets are fairly round and about 6 cm in size, and appear to be the clasts of Unit 5 which were ripped up and incorporated into Unit 6 during a period of erosion. Unit 6 is blocky, with powder calcrete forming between the blocks. Higher up in the sequence, the white clays are punctuated occasionally by thin pebble layers consisting largely of quartzite pebbles as well as some clay pellets. Both of these clay layers are cut by a series of vertical and horizontal cracks filled with a silty, light pinky-brown matrix, that is calcretised and contains numerous rock fragments of under 1 cm in size (Fig. 4.28). The vertical fissures possibly formed during drier periods, when the clay horizon started cracking, and were filled in by younger detrital sediment, or overlying soil matter.

Unit 7

Overlying the white clays is a silty, light pinky brown layer of siltstone of about 5-6 m thick. The sediment from this unit may have provided the source material observed in the cracks in the lower units. The bottom 30-40 cm of this unit contains pebbles of clay and calcrete, and it appears to

be an erosional contact with the underlying clays. Cracks and joints in this unit are lined with calcite crystals and calcrete has formed in the wider cracks. The calcrete is enlarging the cracks as the calcium carbonate precipitates, and in the process is brecciating the surrounding rock. Towards the north of the open pit this layer is almost completely replaced by calcrete. The upper 3m of Unit 7 is characterised by clasts of quartzite, iron formation, and clay supported by the siltstone matrix.

Unit 8

The upper 15 m of the succession at Sishen consists of very hard calcretised siltstones, pebble horizons and clays (Fig. 4.29). Approximately 10 m from the surface, solution of the calcrete appears to have occurred. This is particularly prominent along joint planes, with the holes or pipes having subsequently been filled with a light-brown calcareous silt. These pipes are commonly about 6-7 cm wide, and can be up to 1 m long (Fig. 4.30). About 5 m from the surface calcretised pebble lenses 1.5 m long and 0.5 wide occur, with the clasts commonly angular with no visible evidence of sorting (Fig. 4.31). Clasts consist predominantly of quartzite pebbles. At a similar level the remnants of an older silcrete horizon has been partially eroded, and subjected to several subsequent phases of calcretisation which have resulted in its break-up and deformation. In some places the original rock has been broken up into small angular fragments of 1-2 cm (Fig. 4.32). Joint planes at this level are commonly covered by calcite crystals (Fig. 4.33).

An increasing degree of calcretisation from the bottom of the entire clay sequence (Unit 3) to the top of the sequence at Sishen Mine is evident. Calcareous nodules occur in some of the lower layers, honeycomb and powder calcrete have formed in the middle of the sequence, and the top 15 m of the sequence is almost completely calcretised, culminating in hardpan calcrete near the surface. The increasing maturity of the calcrete towards the surface suggests that the source of the calcium carbonate is from the surface, with transfer down to lower levels through movement of water. In some places a gradational contact occurs beneath different types of calcretes, whereas in other instances a sharp, and sometimes undulating surface separates a massive, blocky, highly indurated calcrete from an underlying softer, less indurated and probably less mature calcrete. These contacts are possibly due to old water table levels.

Percussion chips taken from a borehole directly to the west of the open pit at Sishen Mine were sampled and submitted for chemical analysis. The percentage of CaO reflects the amount of calcification of the sediments and shows a decrease from the surface to the base of the Kalahari Group sequence in the borehole. The SiO_2 values show an opposite relationship and increase with depth (Fig. 4.34).

The upper 4 m of the sequence has a very high value for CaO of over 40 % (> 71 % $CaCO_3$) and throughout the upper 34 m of the succession CaO remains very high, with values of 30 % or higher (~53.5 % $CaCO_3$). According to Nash and Shaw (1998) values of over 50 % for $CaCO_3$ for a duricrust allow it to be classified as a calcrete (see section 4.2.4.2). The high values for $CaCO_3$ encountered in the lithological sequence near the open pit at Sishen Mine means that over 40 % of the sequence can be classified as a calcrete.

An average percentage of 38.25 % for $CaCO_3$ in the percussion chips sampled at Sishen suggests that calcretisation may have added significantly to the thickness of the clay beds. The widespread occurrence of both calcretes and silcretes throughout the Kalahari Basin means that precipitated $CaCO_3$ and SiO_2 should be taken into account when interpreting the thickness of the Kalahari Group sedimentary pile. In some parts of the Kalahari Basin precipitated $CaCO_3$ and SiO_2 may account for as much as 50 % of the sedimentary thickness. The full analytical results from this borehole are contained in Appendix A (Tables A2, A3).

Bootsman (1998) refers to work done by Droste (1961,a,b) who found in his analyses of desert saline sediments in southern California that clay-mineral composition is controlled by the composition of surrounding rocks. This seems to be confirmed from the Kalahari and a geochemical comparison of Ongeluk lavas and clay from the Mamatwan Mine suggest that the clays were derived from the weathered Ongeluk lava (Anthony, 1996). Bootsman (1998) found that the clay sample from Loprung in particular had high amounts of Na and Cl, possibly indicative of a saline environment of deposition, and suggests that the clays were deposited in saline playa environments.

4.3.1.2 Moshaweng River (Locality 2)

Cream and brown sandstones and siltstones of the Eden Formation are exposed along the Moshaweng River to the west of Severn and upstream of the intersection with the Kuruman River at approximately S 26°34.8' E 22°42.85' (Fig. 4.35).

Fig. 4.35 - Exposure of Eden Formation sandstones along the Moshaweng River, South Africa.

A simplified measured section from this locality is shown in figure 4.36. The sediments appear to have been deposited by rivers with periodic aerial exposure enabling plants to be established and weathering of older sediments to occur. Silcretes in the sequence may have been precipitated in shallow lakes or pans, with gritty layers representing periodic in-wash of sediments from surrounding areas, possibly following heavy precipitation.

The lower 3.5 m of the exposure is covered by scree from the cliff face, with the following 3 m consisting of alternating beds of finely layered and more massive sandstone. The sandstone is fine- to medium-grained with the grains generally fairly well rounded. Each of these beds is approximately 40-70 cm thick, but the more massive sandstone beds may be as thick as 1.2 m.

There does not appear to be any difference in grain-size between the layers, but as the more finely layered beds provide a preferential conduit for groundwater to pass along, they weather more

readily, and may be partly calcretised, in particular beneath vertical joints visible in the massive sandstone layers. The more finely layered beds vary in thickness more than the massive beds and may become more massive laterally. In the more massive sandstones the rocks are characterised by a number of tubes which are either hollow, or filled with a sandy matrix of similar grain size to the surrounding sandstone (Fig. 4.37). These structures are commonly vertical or sub-vertical, but some are inclined at an angle, and some are even horizontal. They are generally 8-10 cm long, but some can be traced over 30 cm. Their diameter is generally between 0.7 and 1 cm. It is thought that these structures were formed when the holes previously occupied by roots were filled with sand, and in some cases the hole left by the root is visible, with a bleached zone around it (Fig. 4.38).

The sandstone layers in the next 3-4 m weather more readily, with the slope angle of the exposure becoming markedly more gradual. Partial collapse of a soft, white band of calcretised material of about 1 m thick in the middle of this layer is probably the main contributing factor to the decrease in slope angle. This layer of powder calcrete contains harder lenses and blocks of fine-medium-grained sandstone within it, and in places is cut by a horizontal layer of sandstone of about 20 cm thick. The soft powder calcrete has precipitated as water moved along bedding and joint planes depositing calcareous material. An increase in the amount of weathering and calcification beneath vertical joint planes appears to confirm this. The layer of calcified material is overlain again by a 1 m layer of alternating massive and layered sandstone.

At the top of the succession, a vertical 3 m cliff face has a thin 30 cm fine-grained sandstone layer at its base which is characterised by small ferruginous clasts (< 1cm) and jasper and quartz pebbles. This layer weathers easily, with the ferruginous clasts weathering out, leaving small hollows and pits. Authigenic silica has been precipitated in some of the voids and joints. Overlying this layer is a 30 cm thick silcrete layer which can be traced over a distance of about 12 m before it pinches out. This layer also contains some pebbles floating in the silcrete matrix, as well as evidence of some burrows or root casts filled with a fine-grained sandy matrix. This thin silcrete layer may have been precipitated in a pan environment. The top 1.5 m of the exposure is characterised by a massive sandstone that has root-fill structures similar to those described lower in the sequence. Some of the grains of quartz found in this sandstone exhibit characteristics suggesting some aeolian transportation of the grains occurred. Thin lenses of grit of 2-3 cm in

height and generally only laterally continuous over 1.5-2 m are found at irregular intervals throughout this uppermost sandstone bed. The top 75 cm of the massive sandstone is brecciated by calcretisation into angular fragments which are cemented by a very hard, pale red, fine-grained sand, calcretised matrix.

The top of the sandstone surface appears to have been deeply weathered and then calcretised, with a 1 m zone of cemented sandstone fragments and boulders indicating that the top of the sandstone was exposed for a substantial period of time before being calcretised and covered by the aeolian sand of the Gordonia Formation. The matrix of this calcretised regolith layer is comprised by fine-grained greenish-white and pinkish coloured sands, and these possibly have an aeolian origin.

Jointing and faulting is rare in the Kalahari sandstones exposed in South Africa, but as Figure 4.39 shows, it does occur in the area along the Moshaweng River. The apparent brecciation along some of the joint planes suggests low angle faulting, but further work needs to be done in the area to determine its origin as neotectonic activity in the area is not well documented. The joint planes may provide a conduit for groundwater and water from above, and are sometimes a source of preferential weathering (Fig. 4.40).

4.3.1.3 **Kalahari Gemsbok Park** (Locality 3)

Yellow, cream, reddish, brown and even greenish sandstones and siltstones of the Eden Formation outcrop along the Auob and Nossob Rivers in the Kalahari Gemsbok Park. In general the sandstones in this area of the southern Kalahari are not as well consolidated as those found further to the east along the Moshaweng River, and are also generally finer grained.

Figure 4.41 shows the exposure of sandstones along the Auob River. At this exposure, the bottom 2-3 m consists of thin (~5 cm) beds of pebbles interspersed with layers of sandstone and siltstone with clasts randomly scattered in the matrix. The conglomeratic or gritty layers found in the sandstones (Fig. 4.42) consist of red and grey sandstone and mudstone clasts which may represent the reworked remnants of older Karoo Supergroup sandstone, or even reworked Kalahari Group

sandstones. At some localities along the Auob River as well as directly to the south of the Kalahari Gemsbok Park, the basal pebbly layers are found in large channels, with some cross bedding recognised at Koopan. The basal pebbly layer may vary in thickness, and is commonly overlain by a fine- to medium-grained siltstone. The sandstones in the Kalahari Gemsbok Park often exhibit signs of bioturbation, with an interlocking network of tubes, generally filled with either a calcium carbonate or siliceous matrix, or preserved as hollow tubes. Such bioturbation is especially common in the Kalahari Gemsbok Park and has also been reported from Bushmanland, Namibia (Albat, 1978), Zambia, and Victoria Falls, Zimbabwe where it is known as the Pipe sandstone (Maufe, 1939; Dixey, 1941,1945). There appears to be a relationship between the weathering characteristics of these tubes and their fill, with the tubes filled with a siliceous matrix weathering in positive relief (Figs 4.43, 4.44), and those with a calcareous fill generally being left with a hollow tube as the fill dissolves (Figs 4.45, 4.46). The tubes are possibly the result of the burrowing activities of organisms as well as having been formed by plant roots, and can be up to 30 cm in length.

Irregularly shaped calcareous nodules and powder calcretes generally develop in some of the softer, less consolidated sandstones and siltstones (Fig. 4.47a,b), with thin section work revealing that in some cases up to 50 % of the rock may consist of these calcareous precipitates. Within the sandstones and siltstones calcretised layers of 20-30 cm thick may be developed where they weather as positive relief features (Fig. 4.48). Well developed nodular calcretes overlying this horizon can reach up to 3-4 m in thickness (Fig. 4.49). Overlying the nodular calcretes, hardpan up to 1.5 m thick may occur and may either be exposed, or covered only by the aeolian sand of the Gordonia Formation. When exposed at the surface, or when groundwater moves through the rocks, a layer of calcrete may erode and dissolve, leaving solution cavities (Fig. 4.50), which may be either filled by rock fragments and re-cemented, or covered and recapped by a hardpan calcrete (Fig. 4.51). Older rocks and calcretised layers may be eroded, redeposited as layers of rock fragments, and re-calcretised, or the process of calcretisation or silcretisation may disrupt and distort an older calcretised sequence. Within the calcretised sediments, silcrete veins and lenses and silcrete fillings of cavities or voids are also observed. The silcrete veins are commonly pale brown to grey in colour, and characterised by a conchoidal fracture. The silcrete itself is very hard, and may weather out in positive relief. In thin section the silcrete is seen to be composed of equigranular, well-rounded, quartz grains cemented by a silica matrix.

4.3.2 Stratigraphy of the Kalahari Group in Botswana

The first attempt to define the stratigraphy of the Kalahari was attempted by Siegfried Passarge in the beginning of the 20th century. Passarge (1904) divided the sediments of the Kalahari into five groups based on his work in northern Botswana. Passarge's basal unit was called the Botletle Beds (Botletleschichten) and included the cemented weathering products of underlying strata, overlain by a chalcedonic sandstone and in turn followed by a sandstone which may have been calcareous. The Botletle Beds may have been absent in some areas, and seldom reached any great thicknesses. Overlying the Botletle Beds is the widely distributed Kalahari Kalk, consisting of a lower layer of calc-sinter or poorly consolidated calcareous sandstone. The upper portion of the Kalahari Kalk consisted of calcareous clays and pan tufa (which includes the diatomaceous deposits) or limestone. The Kalahari Kalk is then overlain in most areas by unconsolidated sand (the Kalahari Sand), which Passarge subdivided into four varieties mainly on the basis of colour differences. The uppermost unit recognised by Passarge consists of a reworked surface (the Decksand) and alluvial deposits (Alluvialle Bildungen).

Alluvialle Bildungen (Alluvium) *1. In swamp zone* *2. In sandveld* **Decksand**
Kalahari Sand (divided into 4 subgroups) *Basal River gravels*
Kalahari Kalk (Limestone) *2. Young marls and pan deposits* *1. Calc-sinter and calcareous sandstone*
Botletleschicten (Botletle Beds) *3. Pan sandstone and sandy limestone* *2. Chalcedonic sandstone* *1. Cemented weathering debris of underlying geology*

Fig.4.52 - Passarge's (1904) stratigraphy of the Kalahari sediments (after Thomas and Shaw (1991a).

The major difference between the stratigraphic succession described by Passarge (1904) and that found elsewhere in the basin is the stratigraphic position of his Kalahari Limestone, which included gastropod shells and was possibly formed in shallow water settings similar to those in which the diatomaceous deposits of the pans in the south of the basin accumulated. As has been discussed in preceding sections, the oldest unconsolidated sands predate the oldest known diatomaceous deposits.

While Passarge's (1904) stratigraphic succession provides a useful record for the Okavango-Zambezi area, as with everywhere across the basin, stratigraphic variations across Botswana do occur. Du Plessis (1993) and du Plessis and Le Roux (1995) described the succession further to the south at Sua Pan and divided the Kalahari Group into four formations (Fig. 4.53). At the base of the Kalahari is the Orapa Subgroup, which consists of the Mmashoro Sandstone Formation, probably a rough equivalent of the Botletleschicten of Passarge (1904). The Mmashoro Sandstone Formation consists of a matrix supported, poorly sorted basal conglomerate consisting mainly of basalt clasts. This is then overlain by a bioturbated, trough cross-bedded, coarse-grained sandstone. The fining upward sequence continues into a more fine-grained sandstone with lenses of siltstone and grit, and then back into a mix of fine- and coarse-grained sandstones. Unconformably overlying the Mmashoro Sandstone Formation is the Makoba Subgroup, the base of which consists of a gravelly bed called the Letlhakane Stoneline Formation (du Plessis, 1993). According to du Plessis and Le Roux (1995) this bed varies in thickness between 8 and 118 cm, with an average thickness over 17 sample points of 68 cm. Clasts described are mainly from the Stormberg basalts and the Lebung Group, with some clasts derived from the underlying Mmashoro Sandstone Formation. Calcretisation of the Letlhakane Stoneline Formation has occurred, and it is overlain by calcretes called the Debe Formation (du Plessis, 1993). The unconsolidated sands of the Gordonia Formation form the youngest unit (du Plessis, 1993).

Du Plessis (1993) suggests that the Letlhakane Stoneline Formation could still be forming today. If this is indeed the case, then it should not be assigned a set stratigraphic position below the Gordonia Formation sands.

4.3.3 Stratigraphy of the Kalahari Group in Namibia

In southern Namibia, the Kalahari Group sedimentary rocks outcrop on the Weissrand Plateau and form the northern part of the Urinaib Plateau (SACS, 1980). Only one formation has been identified in the Kalahari succession in the area, the Weissrand Formation which is made up of a basal conglomerate up to 90 m thick, grits and a sandy limestone (du Toit, 1954; SACS, 1980). SACS (1980) reports that the Weissrand Formation stretches to the southeast where it outcrops again in the river valleys of the Kalahari Gemsbok Park. In this area the Kalahari Group sedimentary rocks have been assigned to the Eden Formation and the overlying unconsolidated sands of the Gordonia Formation, but it seems likely that Weissrand is the equivalent of a combination of the basal Wessels Formation and the overlying Eden Formation.

The Kalahari Group sedimentary rocks in northern Namibia were initially divided into three formations, the Beiseb, the Olukonda and the Andoni Formations (SACS, 1980), with Miller (1992a) later adding the basal Ombalantu Formation. The Ombalantu Group consists of very fine-grained semi-consolidated mudstones sometimes containing silt and sand-sized grains. In some areas the base of the Ombalantu Formation is pebbly and thin layers of sandstone may occur. The Beiseb Formation has a widespread occurrence in the Owambo Basin, and consists of a conglomerate made up by well-rounded clasts of brown and grey sandstone and mudstone as well as grey and black chert (Miller, 1992a). The clasts are up to 12 cm in diameter and are set in a matrix of fine- to medium-grained sandstone. According to SACS (1980) the Beiseb Formation could be Eocene in age. The Olukonda Formation comprises a friable, poorly consolidated, reddish brown, poorly sorted, massive sand and sandstone up to 120 m thick. In some areas a thick layer of red clay may occur as well as thin pebble and grit layers (Miller, 1992a). The uppermost unit in the Kalahari sediments of northern Namibia is the widespread Andoni Formation, which consists of interbedded white medium-grained unconsolidated sand, light greenish clayey sand and green clay (Miller, 1992a). The sand can occur in zones of up to 200 m thick and may contain nodules of dolocrete and calcrete up to 30 cm across towards the top of the section, with silcrete nodules becoming more abundant towards the northeast. The clay layers may reach thicknesses of between a few centimetres and 150 m, and may constitute more than half of the Andoni Formation in some areas of the Owambo Basin.

Miller (1992a) mentions a possible correlation between the Cretaceous Kwango Formation of Angola (originally suggested by Furon, 1963) and the Ombalantu Formation. If the Ombalantu and Kwango can indeed be correlated, it suggests that either the Ombalantu may either be a pre-Kalahari lithology, or that the basal Kalahari Group rocks are older than thought.

Formation	Lithology
Andoni	*clayey sand or sandy clay*
Olukonda	*calcareous sandstone*
Beiseb	*gritty to conglomeratic sandstone*
Ombalantu	*siltstone, mudstone*

Fig. 4.54 - The stratigraphy of the Kalahari Group in Northern Namibia (after SACS, 1980; Miller, 1992a).

In northeastern Namibia, in the Bushmanland and Hereroland area, Albat (1978) and McGhee (1979) proposed a slightly different stratigraphy, which was adopted by SACS (1980)(Fig. 4.55). At the base of the Kalahari the Tsumkwe Formation consists of a cemented basal conglomerate. The conglomerate is made up of angular clasts and is poorly sorted, but towards the top of the unit, the clasts are almost absent, and the sandy sediment is loosely cemented. The overlying Eiseb Formation (McGhee,1979; SACS, 1980)(or Kalahari Formation, Albat, 1978) consists of silcretised and calcretised sandy sediment interbedded in places with limestone and chalcedony. The sandy layers are characterised by numerous tubes ranging in diameter from 4 to 8mm (Albat, 1978). The uppermost Kalahari unit observed in this area is the Omatako Formation (Albat, 1978)(or Kalahari Formation, McGhee, 1979) comprising ferricretes and ferruginous sandstones with occasional inclusions of ferruginous sandstone fragments overlain by unconsolidated aeolian sand.

Formation	Lithology
Omatako	*Ferricrete and ferruginous sandstone*
Eiseb	*Hard to porous silcrete-cemented quartz sand overlain successively by partially silicified cream-coloured limestone, silcrete cemented quartz sand with a network of porous channels, limestone and pebbly, slightly lime cemented sandstone, silcrete cemented sand and calcrete cemented sand*
Tsumkwe	*Lime-cemented sandy conglomerate and grit which interfinger with and pass upwards into sandy layers of variable lime content that are overlain by thinly bedded, poorly cemented sand.*

Fig. 4.55 - The stratigraphy of the Kalahari Group in Bushmanland, Namibia (after SACS, 1980).

In spite of being adopted as the type section for Bushmanland, Namibia by the South African Committee for Stratigraphy (SACS, 1980), the stratigraphic successions proposed by Albat (1978) and McGhee (1979) received criticism from Balfour (1981) as they ignored local variations in stratigraphy and failed to recognise duricrusts, attempting to give them a stratigraphic position. Balfour (1981) suggests that the stratigraphy for Bushmanland should be based on the following lithological succession:

Unconsolidated sand
Calcretes, silcretes, minor ferricretes
Red brown, fine-grained sandstone with occasional intercalated mudstone
Red mudstone
Quartzitic grit/conglomerate

Fig. 4.56 - The Kalahari Group sequence in Bushmanland, Namibia (Balfour, 1981).

The similarities between the succession identified by Balfour (1981) and that occurring South Africa are clear, with even the position of the duricrust layer mirroring that of the calcretes of the Mokalanen Formation. The differences between Balfour's succession and that of SACS (1980) for the same Bushmanland area is, however, quite substantial, with the major differences being the absence of mudstones and the interpretation of limestones instead of calcretes in the latter stratigraphy. One of the major problems incurred with stratigraphic correlation across the Kalahari Basin is the lateral inconsistency in the sedimentary succession, as Figures 4.57 and 4.58 show. In addition to this, the sediments have been variably described as sandstones (eg borehole Strat Test 1; SACS, 1980), unconsolidated sands (e.g. borehole 9563 at Ombalantu; Miller, 1983) in the Owambo region, in Bushmanland as sandstones overlying unconsolidated sands (SACS, 1980) or as unconsolidated sands overlying sandstones and mudstones which may or may not be capped by calcretes or are calcretised themselves (Balfour, 1981). The description of the grain size of the sediments in various logged boreholes also seems to be inconsistent and sediments previously described in borehole logs as siltstones and sandstones were described by Miller (1992a) as mudstones that contained silt and sand-sized grains.

The paucity of detailed boreholes in which the Kalahari Group sedimentary rocks have been accurately described in northern Namibia has meant that the Kalahari Group stratigraphy in northern Namibia is not clearly defined. The inconsistency between which has been called sandstone and which has been termed partly cemented sand hampers the correlation of units even further, and it is tempting to say that they are the same thing. This interpretation is dangerous, however, because if the sandstone is the equivalent of (for example) the Eden Formation of South Africa, it places it at a possible Eocene age. If on the other hand, it represents the deposition of more recent sediments, possibly by the Cunene River in the last 3 million years, an interpretation of the sediments as older sandstones is flawed.

4.3.4 Stratigraphy of the Kalahari Group in Zambia

Money (1972) divided the Kalahari Group in Zambia into two formations, the Barotse Formation and the overlying Zambezi Formation (Fig. 4.59). The Barotse Formation was in turn divided into Upper, Middle and Lower sections. At the base of the Kalahari Group the Lower Barotse Formation is made up largely of basal conglomerates. Overlying this, the Middle Barotse comprises units of bedded ferruginous sandstone and quartzites, and the Upper Barotse comprises

massive quartzites, sandstones and conglomerates. At the top of his succession Money divided the Zambezi Formation into the Mongu sand member (consisting of the unconsolidated sands), overlying and interspersed duricrusts, and pan sediments and limestones.

Boreholes drilled into the Kalahari Group sedimentary rocks near the town of Mongu in southwestern Zambia show that, as with elsewhere in the basin, the lateral variations in the Kalahari Group stratigraphy in Zambia can be substantial, even over a relatively small area (Thomas and Shaw, 1991a; Fig. 5.60).

4.3.5 Stratigraphy of the Kalahari Group in Zimbabwe

Much of the early information on the Kalahari Group from Zimbabwe comes from Maufe (1939) from the Victoria Falls area. Maufe attempted to fit his scheme into Passarge's (1904) stratigraphy. At the base of the Kalahari he correlated a greyish, mottled and translucent to opaque rock called the Kalahari Chalcedony with the Botletle Beds. This lower unit was interpreted as silicified limestone by Lamplugh (1907). Overlying this unit Maufe identified a silica-cemented white, pink or red sandstone which he named the Pipe Sandstone. The Pipe sandstone, as its name implies, is characterised by numerous tubes in the rock, which Maufe attributed to deposition in a reed bed. Overlying the Pipe Sandstone a layer of weathered Pipe sandstone containing nodules and a carstone rubble bed, interpreted as a pisolithic ferricrete (Thomas and Shaw, 1991a) was described, with a matrix consisting of a pebbly sand made up of either the weathering products of the Pipe Sandstone (Dixey, 1941, 1945), or of the unconsolidated Kalahari sand at the top of the sequence (Maufe, 1939).

Kalahari Group	**Kalahari Sand**	*Unconsolidated and semi-consolidated sands*
	Pipe Sandstone	*Sandstones, numerous pipe-like structures*
	Kalahari Chalcedony	*Silicified limestone*

Figure 4.61 - The stratigraphy of the Kalahari Group in Zimbabwe (after Maufe, 1939).

Stratigraphic variations in the Kalahari sequence in Zimbabwe (Fig. 4.62) are well illustrated by a series of boreholes drilled in the Kennedy area of Zimbabwe. The orthoquartzites referred to in Figure 4.62 (Thomas and Shaw, 1991a) are possibly sandstones that have undergone some degree of silcretisation.

4.3.6 Stratigraphy of the Kalahari Group in Angola

Unfortunately due to the ongoing civil war in Angola very little is known about the Kalahari Group stratigraphy, particularly in the southwest of the country. It is known from mapping and borehole evidence that over large parts of the country the Kalahari Group sediments are underlain by either Karoo Supergroup sediments or basalts, or by Cretaceous rocks, belonging mainly to the Kwango Formation. The similarity between the Cretaceous sediments and the Kalahari Group sediments makes borehole logging difficult and the contacts between them are not always clear.

A general stratigraphy for the Kalahari Group in Angola consists of two main formations, the Série Superior and the Série Inferior (Fig. 4.63). These two formations correspond to the Sables Ochres and Série des Grès Polymorphes of the Democratic Republic of Congo.

	Formation	Lithology
Sistema do Kalahari	**Série Superior (Kalahari Superior)**	*Unconsolidated and semi-consolidated sands*
	Série Inferior (Kalahari Inferior)	*Sandstones, conglomerates, clays*

Fig. 4.63 - The stratigraphy of the Kalahari Group in Angola (Martins, 1966; Pachero, 1976).

The Kalahari Superior consists largely of unconsolidated sand and more recent duricrusts and pan sediments. The underlying Kalahari Inferior consists largely of pale yellow, white or pink sandstones and conglomerates. The sandstones may have variable amounts of clay in them and in some areas contain calcrete nodules or have been silcretised (Pachero, 1976). In northeastern

Angola there was a clear period of weathering following the silicification of the sandstones and preceding the deposition of the unconsolidated sands and in much of northeastern Angola the silicified sandstones were deeply eroded prior to the deposition of the overlying sands (Janmart, 1953).

4.3.7 Stratigraphy of the Kalahari Group in the Democratic Republic of Congo

Lepersonne (1945) and Cahen and Lepersonne (1952, 1954) divided the Kalahari sedimentary rocks of the Democratic Republic of Congo into three series, largely based on the position of erosion surfaces. At the base of the Kalahari Group, the lower Kamina Series comprises mainly sand, gravels and sandstones, and was considered to be older than any of the Kalahari Group sedimentary rocks identified further south as it was eroded by a Cretaceous land surface. It is considered to be late Cretaceous in age (Giresse, in press). Overlying this, the Grès Polymorphes (Cornet, 1894), consist of a basal conglomerate, silicified sandstones and chalcedonic limestones and overlying unstratified sands (Giresse *et al.*, in press). This is overlain by the unconsolidated sands, termed the Sables Ochres. In the Congo Basin, the thickness of the Série des Grès Polymorphes is fairly consistently around 80 to100 m while the overlying Sables Ocres reaches thicknesses of up to 120 m (Giresse, in press).

	Formation	Lithology
Kalahari Series	**Sables Ochres (Etage Superior)**	*fine-grained, unconsolidated sands*
	Grès Polymorphes (Etage Inferior)	*sands* *silicified sandstones and chalcedonic limestones* *basal conglomerate*
	Kamina Series	*sands, gravels and sandstones*

Fig. 4.64 - The stratigraphy of the Kalahari Group in the Democratic Republic of Congo (after Claeys, 1947; Cahen, 1954; Cahen and Lepersonne,1952, 1954; Lepersonne, 1945; Giresse, in press).

As with the sediments found in Namibia and Angola, there seems to be some confusion between what forms the base of the Kalahari Group, and what can be classified separately as Cretaceous sediments. The Kamina Series may be the equivalent of the Wessels Formation of South Africa.

4.4 Discussion

The Kalahari Group consists of a fairly complex succession of sedimentary rocks and attempts to define and formalise the stratigraphy have been hindered by a lack of good exposures, the lateral inconsistency of sedimentary units and the tendency by previous researchers and mining operations to ignore the Kalahari Group sediments and concentrate on what might lie beneath them.

Tectonic activity which is still ongoing in parts of the basin has resulted in local stratigraphic variations in the stratigraphy, through faulting, possible rifting, and the formation of several sub-basins, or depositional centres throughout the basin. Borehole information from the Kalahari Basin has often been poorly described. In many instances percussion drilling was used through the Kalahari Group formations, and the chips retrieved from these logs were either discarded or only described in the broadest terms. In the areas where Cretaceous sedimentary rocks underlie the Kalahari Group rocks, for example in northern Namibia, Angola, and the Democratic Republic of Congo, it may be difficult from borehole percussion chips to distinguish between the Cretaceous and the overlying Kalahari Group sedimentary rocks. It is more than likely that Cretaceous sedimentary rocks have been assigned to the Kalahari in borehole logs in the past, and conversely, many of the Cretaceous sedimentary rocks recorded in boreholes may belong to either the lower Kalahari Group or even the Upper Karoo. Many of the Cretaceous rocks recorded in the Congo Basin may indeed be the equivalent of the basal Kalahari Group rocks found further south. This may result in thicknesses of Kalahari or Cretaceous rocks being misrepresented in some areas.

The extensive occurrence of duricrusts has also led to stratigraphic problems, particularly in cases where they may be assigned a fixed stratigraphic position and in some cases even a formation name. In the Congo Basin a silicified sandstone was used as a marker bed until it was discovered that silicification occurred under similar climatic conditions at different levels of the unit

(Sekirsky, 1956; Giresse, in press). In a similar way, the calcretes in the Kalahari are not all the same age, nor do they occur at only one stratigraphic level within the Kalahari succession. Calcretes deposited by ground water processes would, in all probability, have formed at the level of the water table and may cut across older lithologies. In addition to this, younger calcretes may commonly form below older boulder calcretes (Netterberg,1978) and it is theoretically possible for calcification to occur at any depth (Netterberg,1969). Calcretes do not form in a discrete event and any one calcrete profile will probably contain carbonate of different ages. The recommendation by Netterberg (1978) that calcrete should be regarded as a lithostratigraphic unit but not as a regional chrono-stratigraphic horizon is concurred with.

A proposed generalised lithostratigraphic sequence for the Kalahari Group is shown in Fig. 4.65. It is important to note that the different stratigraphic successions proposed for different parts of the Kalahari Basin may not contain all of the lithologies shown and, as has been discussed, local variations in the litho-stratigraphic sequence do occur and may be significant. Many of the formal stratigraphic successions do, however have basic similarities and correlations can be attempted.

Pan sediments and diatomaceous deposits
unconsolidated sand
sandstones (calcretised, silcretised or ferruginous) with some grit or coarse to fine gravel layers
clays with some sandy, grit, or coarse to fine gravel layers
coarse to fine gravels

Fig. 4.65 - Proposed litho-stratigraphic sequence for the Kalahari Group.

Correlations between the various Kalahari Group formations identified from various parts of the basin are largely based on lithological similarities, as well as on geomorphological relationships. The absence of absolute dates for the Kalahari Group makes some of the correlations tenuous, but it is possible to correlate the main units throughout the region (Fig. 4.66).

South Africa (Smit, 1977) (M.A.Thomas, 1981)*	**Southern Namibia** (SACS, 1980)	**Northern Namibia** (SACS, 1980) (Miller, 1992)*	**Northeastern Namibia** (SACS,1980)	**Zimbabwe** (Maufe, 1939)	**DRC** (Cahen & Lepersonne, 1952,1954) (Claeys, 1947)*	**Angola** (Pachero, 1978)	**Zambia** (Money, 1972)	**Botswana** (Passarge, 1904)	**Botswana** (Du Plessis, 1993)
Lonely Formation* ***Goeboe Goeboe Formation**** ***Obobogorop Formation****							***Zambezi Formation*** *Limestone and clays on pan floors*	***Alluvialle Bildungen***	
Gordonia Formation		***Andoni Formation***		***Kalahari Sand***	***Sable Ochres (Etage Superieur)****	***Série Superior***	***Zambezi Formation*** *Mongu sand member*	***Kalahari Sand*** *unconsolidated sand*	***Gordonia Formation***
Mokalanen Formation* ***Eden Formation***	***Weissrand Formation***		***Omatako Formation***	***Pipe Sandstone*** ***Kalahari Chalcedony***	***Grès Polymorphes (Etage Moyen)****	***Série Inferior***	***Upper Barotse Formation*** ***Middle Barotse Formation***	***Kalahari Kalk***	***Debe Formation***
								Botletleschicten	***LSL Formation***
		Olukonda Formation	***Eiseb Formation***						***Mmashoro Formation***
		Beiseb Formation							
Budin Formation							***Lower Barotse Formation***		
Wessels Formation			***Tsumkwe Formation***		Kamina Series		***Basal conglomerate***		
		Ombalantu Formation*							

Fig. 4.66 - Attempted correlations of the Kalahari Group formations across the region

CHAPTER 5: REGIONAL GEOMORPHOLOGY AND KALAHARI BASIN EVOLUTION

5.1 Introduction

Southern Africa is characterised by anomalously high topography, with southern Africa having an average elevation of around 1200 metres above sea-level (Gilchrist and Summerfield, 1990) compared to less than 200 m in north eastern Argentina and southern Brazil (Brown *et al.*, 2000). The coastal topography of southern Africa is dominated by an escarpment, which continues from the southern Cape up both the east and the west coasts of Africa. From the southern Cape, where the escarpment reaches some 1500 m in elevation, it becomes progressively lower and less well developed up the west coast, becoming very low in northern Namibia, before rising up again to over 2500 m in Angola (Brown *et al.*, 2000). The combined area of the eastern and southern African plateaus is some 1 x 10^7 km^2, which makes this the largest topographical anomaly on any continent. To the southwest of Africa lies an area of anomalously shallow bathymetry, with an amplitude of about 500 m. This area of shallow bathymetry combined with the Southern and East African plateaus is termed the African Superswell (Fig. 5.1; Nyblade and Robinson, 1994). The present drainage of South Africa is strongly influenced by the escarpment which forms an important watershed separating short rivers draining towards the coast, from the generally larger drainage networks of the interior of the country. The escarpment is breached in several places by the large rivers, with the Orange, Limpopo, Cunene and Zambezi being the most notable (Fig. 5.2).

Recent analysis of the topography of Africa by Doucouré and de Wit (2003) suggests that by the Mesozoic, Africa was already characterised by a “bi-modal topography”, with uplift having occurred in the Palaeozoic and Partridge *et al.* (in prep.) suggest that at the time of Gondwana break-up Africa had an average elevation ranging from 2000 to 2500 m. Nyblade and Sleep (2004) acknowledge the possibility of Mesozoic uplift of the Southern African Plateau, Quennel (1960), believed that phases of uplift in the Cretaceous and Cenozoic collectively resulted in the formation of the East African Swell and contributed to the formation of the African Superswell and Burke

(1996) believes that the African Superswell has only developed over the past 30 million years, with

the subswells within it episodically active throughout this time.

It is likely that the topography of southern Africa is a combination of episodic uplift over the last 400 million years. This chapter concentrates initially on uplift in the Cretaceous and the Cenozoic on a subcontinental scale, but as the focus of this study is the Kalahari Basin and its development, the influence of tectonic activity on drainage patterns and basin evolution are then examined more closely by examining certain parts of the Kalahari Basin in more detail.

5.2 Cretaceous and Cenozoic uplift

Cretaceous marine sediments on the west coast of South Africa were deposited in a series of offshore basins since about 127 Ma ago, with sedimentation controlled by the tectonic uplift or subsidence of the continental margin (McMillan, 2003). According to McMillan (2003) during the Early Barremian to the Early Cenomanian uplift and subsidence in all of the basins happened in unison, but from around 95 Ma to about 65 Ma southern Africa underwent repeated east-west rolling motions resulting in different stratigraphic sequences on the Atlantic and Southern Margin, compared with the Indian Margin.

Apatite fission-track data across the western escarpment yielding ages of 100-130 Ma (Brown *et al.*, 1990), and between 166 ± 6 Ma and 70 ± 5 Ma (Brown *et al.*, 2000), was interpreted as indicating a phase of cooling of the samples along the south western margin of southern Africa during the early, middle, and late Cretaceous (Brown *et al.*, 2000). Brown *et al.* (2000) prepared maps showing estimates of the amount of syn-rift (158-118 Ma) and post-rift (118-0 Ma) denudation in southwestern Africa and Brazil, and show that while the syn-rift denudation was moderate to low (~ 1km), the amount of denudation following break-up was substantial (~3-5km). While the depth of denudation shows a general decreasing trend from the coast towards the interior, the spatial distribution of the denudation can be highly variable, with anomalously high denudation values being indicated for a region some 500 km inland, adjacent to the western border of the Kaapvaal craton (Brown *et al.*, 2000). Further evidence for high topography and possible uplift along the western margin of southern Africa is suggested by the major sediment deposition in the Walvis and Orange offshore basins during the late Cretaceous- early Tertiary (Brown *et al.*, 1990; Dingle and Hendy, 1984; Rust and Summerfield, 1990). The volume of sediment deposited offshore in the Orange and Walvis Basins is estimated to be ~2.8 x 10^6 km^3 (Rust and Summerfield, 1990), which equates to a depth of denudation of some 1.8 km averaged over the present Orange River Basin and the other drainage catchments on the western side of southern Africa (Brown *et al.*, 2000). Mid-Cretaceous uplift of the southern escarpment is suggested by increased sediment deposition in the Outeniqua Basin during this time as well as from fission track data gathered from the escarpment to the north. (Tinker and de Wit, 2004). The presence of massive amounts of Cretaceous sediment on the east coast of Africa near the current mouths of the Limpopo and Save Rivers provides additional evidence of either Cretaceous uplift, or of subsidence in that area (Moore and Larkin, 2001).

While the increase in denudation during the Cretaceous may have been influenced by factors such as the differential weathering potential of different lithologies in the drainage basins (Brown *et al.*, 2000) and changes in climate during the late Cretaceous (see Chapter 6), the offshore data gathered by various researchers suggests that high topography along the western margin of southern Africa probably existed both prior and subsequent to Gondwana break-up. The greater depths of Cretaceous denudation recorded closer to the continental margin as compared with the continental interior were interpreted as possibly being due to a flexural isostatic response which would have been able to maintain an elevated topography parallel to the margin right into the Cenozoic (e.g. Brown *et al.*, 2000; de Swardt and Bennet, 1974; Gilchrist and Summerfield, 1990). Partridge (1998) and Partridge and Maud (2000) believe, however, that there is little evidence for the inland migration of this flexure and believe instead that a marginal escarpment existed from the time of Gondwana break-up and is a rift-generated feature. According to Partridge and Maud (2000) this escarpment was rapidly eroded during the Cretaceous before reaching a position some 20 km from the current position of the Great Escarpment by the end-Cretaceous. An alternative suggestion put forward by Burke (1996; *pers. comm.*) that the escarpment may be related to the development of the African Superswell and is only be about 30 million years old can not be discounted but further evidence is needed before this can be fully accepted.

The erosional surfaces that developed during the Cretaceous were placed by King (1967, 1976, 1983) into erosional cycles separated by tectonic episodes (Fig. 5.3). His first cycle, termed the Gondwana planation, was terminated by the break-up of Gondwana at the end of the Jurassic, and

the subsequent post-Gondwana surface was terminated by uplift in the mid- to late Cretaceous. This was followed by the formation of the African Surface, both above and below the Great Escarpment (King, 1963, 1967), which lasted until the Oligocene before it to was terminated by an uplift event and the formation of the Post African 1 surface by the Miocene (Partridge and Maud, 2000). The potential for longevity of sub-aerial continental landforms or erosional surfaces has been questioned by the findings of Brown *et al.* (2000) in South Africa and recently by Belton *et al.* (2004) in Australia. As significant denudation has undoubtedly occurred in southern Africa since Gondwana break-up, some of the earliest erosion surfaces described by King (1976, 1983) and others (e.g. Partridge and Maud, 1987) may not have been able to survive for any meaningful period of time.

During the period from the end of the Cretaceous to the Pleistocene, four hiatuses in sedimentation have been identified from seismic and drilling surveys conducted in the sedimentary fills of offshore basins (e.g. Siesser and Dingle, 1981; Aizawa *et al.*, 2000). Although Partridge and Maud (2000) do not believe that significant regional uplift of southern Africa occurred in the Palaeocene, uplift and/or subsidence did occur along some parts of the continental margin. Recent seismic surveying and drilling off the Namibian coast suggests that significant uplift of that part of the continent occurred in the beginning of the Tertiary (Aizawa *et al.*, 2000) and Palaeocene uplift of the Agulhas Arch is also believed to have occurred (McMillan, 2003). Periods of uplift of the continent in the Oligocene, Miocene and Late Pliocene to early Pleistocene have been identified (Burke, 1996; Summerfield, 1985; Partridge, 1998).

Several axes of epeirogenic flexuring have been identified, namely the Griqualand-Transvaal, Kalahari-Zimbabwe, Khomas, Otavi and Soutpansberg (du Toit , 1933), Escarpment (King, 1963), Harts (Mayer, 1973) and Ciskei-Swaziland (Partridge and Maud, 1987; Partridge, 1998) Axes, where uplift along the axes and/or adjacent subsidence occurred during the Cenozoic (Fig. 5.4). The uplift initiating the formation of the Post African I Surface (King and King, 1959; Partridge and Maud, 1987) was asymmetrical and also involved uplift along the Griqualand-Transvaal and Kalahari-Zimbabwe axes (Partridge and Maud, 2000). An $^{40}Ar/^{39}Ar$ date of 12-15 Ma from a sub-surface pedogenic manganese crust (van Niekerk *et al.*,1999) provides a minimum age for the formation of the surface, confirmed by the fossil evidence from deposits resting on the Post African I Surface (e.g. Partridge and Maud, 2000). Evidence for Miocene uplift of some 150 m of the western escarpment is clearly seen in the incised river valleys of the Orange and Koa and the eastern

escarpment was raised by some 300 m (Fig. 5.5)(Partridge, 1998). Massive uplift of up to 900 m on the eastern side of southern Africa occurred in the Pliocene (between 5 and 3 Ma), as is evinced by the convex upward profiles of rivers across remnants of the African and Post-African surfaces (Partridge, 1998; Partridge and Maud, 2000; Fig. 5.5). According to Partridge and Maud (2000) the axis of maximum movement (the Ciskei-Swaziland Axis) lay about 80 km inland and stretched from inland of Port Elizabeth to Swaziland. In the interior rejuvenation of the Griqualand-Transvaal and Kalahari-Zimbabwe axes took place, along with simultaneous subsidence in the Bushveld Basin (Partridge and Maud, 1987; Fig. 5.5).

5.3 Drainage evolution and basin formation

Following the epeirogenic uplift of the edge of the continent in the Cretaceous, drainage patterns in southern Africa were characterised by a dual drainage pattern comprising rivers draining directly from the uplifted margin towards the sea and rivers which flowed in the opposite direction, into the interior of the continent. From the Cunene River southwards to the modern Orange River there were no major rivers draining westwards from the interior of the continent, a phenomenon possibly attributable to the Parana plume and its associated doming (Cox, 1989; Moore and Blenkinsop, 2002). On the eastern side of the Kalahari Basin the only path to the Indian Ocean was via the Limpopo River in the south. It is believed that the rivers originating in the highlands of Angola therefore flowed southwards until they reached the Kalahari, Karoo or Limpopo Rivers and were diverted towards the Atlantic or Indian Oceans. Although the topography of southern Africa has been modified during the Cenozoic by faulting and tectonic uplift, the sub-Kalahari topographical surface prepared as part of this research (Fig. 5.6) provides some indication of how the drainage patterns looked prior to Kalahari Basin formation. Evidence from the sub-Kalahari geological map and isopachs of the entire Kalahari Group and the basal gravels help to further constrain the courses of the Cretaceous rivers.

Subsidence of the interior of southern Africa, probably in the Late Cretaceous, back-tilted some of the interior rivers into the newly formed Kalahari Basin, and deposition of Kalahari Group sediments began. The evidence for drainage and basin evolution in various important parts of the Kalahari Basin are discussed in more detail in the following sections.

5.3.1 Drainage evolution of the southwestern Kalahari Basin

Several theories have been put forward for the nature of the drainage system in the southwestern part of the continent during the late Cretaceous. The most widely accepted and cited mid to late Cretaceous drainage for southwestern Africa proposes two major westward flowing river systems, the Karoo and the Kalahari Rivers (de Wit, 1993; 1999; Partridge, 1998; Fig. 5.7). In this model the Karoo river drained much of the North-West and Free State Provinces and Lesotho, while the Kalahari River drained much of southern Botswana and Namibia. The sediment deposited in the Atlantic Ocean adjacent to the mouth of the Karoo River is older (117-103 Ma) than at the mouth of the Kalahari River (100-75 Ma) and this may be due to the Karoo River having joined the Kalahari River some time in the mid- to late-Cretaceous (J.D. Ward, *pers. comm.*).

The Karoo River was believed to have had several right bank tributaries, including the proto-Morokweng, proto-Molopo and proto-Harts which flowed southwards from Botswana in the mid to late Cretaceous (Bootsman, 1998; Partridge, 1998; Figs 5.7). A major southward flowing river, the Trans-Tswana River was believed to have joined the Karoo River (the middle proto-Orange) some 30 km downstream from its confluence with the Vaal River, between Douglas and Prieska (McCarthy; 1983) and this may have been the same as the proto-Molopo. The Morokweng Impact Structure had a significant effect on the drainage of the area, with the deeply weathered crater being exploited by rivers. During the mid-Cretaceous, the proto-Morokweng River was believed to have flowed around the impact structure towards the south where it formed one of the tributaries of the Karoo River or upper proto-Orange (Bootsman, 1998, Bootsman *et al.*, 1999). By the end of the Cretaceous this river is believed by Bootsman (1998) to have joined with the proto-Molopo River.

Recent evidence from the southern rim of the Kalahari Basin suggests that the proto-Morokweng and proto-Molopo flowed northwards, away from the Karoo River and towards the Kalahari Basin. To the southeast of Kuruman, at Mahura-Muthla, gravel beds believed to have belonged to the proto-Morokweng River are preserved as a series of sinuous channel segments (Bootsman, 1998; de Wit *et al.*, 2000). Fossil wood constrains the age of the gravels to the Upper Cretaceous (Partridge, 1998, Bamford, 2000; Ward *et al.*, 2004). Although clasts from the channels were originally interpreted as indicating a flow direction towards the southeast (Partridge, 1998), Moore

and Moore (2004) believe that flow might have been northwards. Recent work by Ward *et al.* (2004) appears to confirm a northward flow direction, with clast lithologies, fining direction, imbrication of clasts and ilmenite chemistry being cited as evidence.

In a recent paper Moore and Moore (2004) also concluded that rivers flowed northwards into the Kalahari River, but have highlighted the lack of related Cretaceous palaeochannels or gravel terraces associated with a Karoo River and believe much of the mineral-chemical evidence for the river (de Wit, 1993, 1999) can be explained by erosion of Dwyka sedimentary rocks. Moore and Moore (2004) believe that the Karoo River did not exist and that prior to the exposure of the pre-Karoo Cargonian basement, most of the rivers in central South Africa flowed northwestwards into the Kalahari River system (Fig.5.8).

Although the northward flow direction for the Mahura Muthla system fits the model of Moore and Moore (2004), the evidence for a Karoo River cannot be ignored (de Wit, 1993, 1999, *pers. comm.*) and the northward flow of the Mahura Muthla system can be more easily explained by the presence of a watershed coinciding with the position of the Cargonian Highlands (Visser, 1987) and separating the drainages of the Karoo and Kalahari river systems (see Chapter 8). The sub-Kalahari topographical surface generated as part of this research gives some indication of the courses followed by the Kalahari River and some of its tributaries prior to the Late Cretaceous formation of the Kalahari Basin. The possible courses followed by the Kalahari River and its tributaries are superimposed on a portion of the sub-Kalahari topographical surface (Fig. 5.9). When the isopachs of the Kalahari Group, isopachs of the basal gravels and the sub-Kalahari geological map are used along with the sub-Kalahari topographical surface additional support for the configuration of the Cretaceous drainage for the southern Kalahari is forthcoming (Figs 5.10, 5.11, 5.12).

Gravels at the base of the Kalahari Group are believed to have been deposited at the base of the Kalahari Group by Cretaceous rivers. The isopach map of the basal gravels (Fig. 5.10, Appendix D) covers the portion of the Kalahari Basin within the borders of South Africa and suggests that a river flowed ENE-WSW near the Botswana-South Africa border. This large river is possibly the Kalahari River of de Wit (1993). It is not clear if this river followed the north-south path of the lower reaches of the Molopo River or if it continued towards the southwest through what is now

southeastern Namibia. To the east of the area, the unavailability of reliable data for the basal gravels in the Morokweng Impact Crater vicinity has prevented any thicknesses of gravels being shown for this area, but the sub-Kalahari topographical surface and the isopach map of the Kalahari Group clearly show the presence of valleys flanking the crater (Figs 5.9, 5.11). The force of the meteorite impact at Morokweng in northern South Africa at around 144 Ma may have reactivated some of the older northeast-trending faults in the border area between South Africa and Botswana, and the ring faults formed by the impact were later exploited by the rivers in the area which removed the weathered material surrounding the central part of the structure and carved out large valleys. The Mahura-Muthla River, flowed northwards off the Cargonian highlands (Ward *et al.*, 2004), around the crater and westwards into either the Kalahari river or another river whose course is suggested by a zone of thick gravels extending along the Hotazel valley. This river possibly also flowed northwards off the Cargonian highlands as did a parallel river to the west of the Korannaberg and Langeberg hills flowing northwards along a longitude of approximately 22° 15' E and into the Kalahari River. On the other side of the watershed to this river, a river possibly drained southwards into either a tributary of the Lower Kalahari River (proto-Orange) or the Karoo River before downwarp of the Kalahari Basin reversed its drainage towards the north.

An examination of the sub-Kalahari geological map (Appendix C, Fig. 5.12) suggests that the basement geology had an important influence on the courses of the Cretaceous rivers. In the southern, South African, part of the basin, rocks of the Archaean Kraaipan Group and Proterozoic Transvaal and Olifantshoek Supergroups and Brulpan and Wilgenhoutsdrif Groups were exposed by erosion and formed north-trending ridges which controlled drainage patterns and Kalahari Group deposition in that area during the late Cretaceous and Cenozoic. To the north of the Kalahari River, the northern extent of its right bank tributaries is not clear, but the sub-Kalahari geological map does suggest that these tributaries only extended as far northeast as the present position of the watershed separating the Okwa from the Molopo drainage basins (the Kalahari Schwelle). The southern portion of the sub-Kalahari geological map clearly shows that, while Karoo Group sediments of the Dwyka and Ecca Group are exposed to the south and southwest of the Kalahari Schwelle, along the schwelle itself upper Karoo rocks are preserved. This would appear to indicate a pre-Kalahari drainage basin existed to the southwest of the schwelle which acted as a watershed at this time and the sub-Kalahari topographical surface gives some idea of the shape of the basin. There appears to be a relationship between the distribution of the pre-Kalahari rivers and the distribution of Dwyka rocks. As Fig 5.12 shows, the upper part of the Kalahari River, one of its right bank tributaries and the river flowing along the Hotazel valley all followed valleys underlain by Dwyka. Neotectonic movement along the Hebron Fault, one of the major faults forming the Nama Lineament (Fig. 3.1), is clearly seen on satellite imagery as well as aerial photographs (Corner, 2000; B.Corner, *pers. comm.)* and it is possible that Cretaceous movement along these and other NW-trending faults may have controlled the NNW-SSE orientation of certain rivers flowing southwards down the southeastern edge of the Kalahari Basin and into the Kalahari River. A large river flowing from the northwest, the "Aranos" which was later responsible for the deposition of large amounts of sediment in the Aranos sub-basin, was possibly controlled by reactivation of faults lying parallel to the Nama Lineament. This river was possibly fed by the upper part of the Fish River as suggested by Wellington (1955). Wellington (1955) suggested that the Fish River in Namibia originally extended down to the Molopo-Auob along a course through an area characterised by a broad swathe of pans between the Molopo-Auob and Orange Rivers. The distribution of Karoo dolerite dykes and sills also had a major influence on the course taken by this Aranos River which possibly flowed around the western side of the sills. To the east of the Aranos River a parallel southeasterly flowing river possibly flowed through an area now occupied by the Sewe Panne in the Kalahari Gemsbok Park (see section 5.36).

During the late Cretaceous headwaters of the proto-Harts River beheaded the Mahura Muthla system and extended into south-eastern Botswana and the channels of this river may be preserved today as gravels at Lichtenburg and Mafikeng (de Wit *et al.*, 2000). At about this time, uplift of the Griqualand-Transvaal Axis (Moore, 1999) or downwarp to the north of the Griqualand-Transvaal Axis began, back-tilting the southwards flowing rivers into the newly formed basin and resulting in deposition of sediment in river channels and in the lakes that formed (Fig. 5.13a).

By the Miocene, the capture of the middle Orange by the lower Orange River had already occurred (de Wit, 1999) and de Wit *et al.* (2000) believe that the Orange River may have occupied its present course since at least the late Oligocene. Arid conditions at the end of the Miocene are thought to have resulted in the choking of one of the main left bank tributaries of the lower Orange, the Koa (de Wit, 1999) and this was possibly accentuated by a continuation of the relatively minor uplift in the order of 150 m and 100 m along the Griqualand-Transvaal Axis that is thought to have

occurred in the early Miocene and again in the Pliocene respectively (Fig. 5.5; Partridge and Maud, 1987; Marshall, 1990; Partridge, 1998). The Pliocene uplift along the Griqualand-Transvaal Axis (du Toit, 1933; Partridge, 1998) resulted in a westward shift of the Molopo drainage, with the upper Molopo, Kuruman and their tributaries following a westward course (Bootsman, 1997) and the drainage patterns looking similar to those existing today (Fig. 5.13b).

Moore (1999) suggests that uplift in the southern Kalahari was not limited to the Griqualand-Transvaal Axis. According to Moore (1999), the lower 13 km of the Molopo River is characterised by a steep gradient and distinct concave-up profile, as is the course of the Orange above this confluence. In addition to this, tributaries joining the Molopo in the lower parts of its course form hanging valleys where they join the Molopo. Further up the course of the Molopo, above Riemvasmaak, the profile changes to convex up, and the gradient decreases rapidly. According to du Toit (1933) and Partridge (1998) the convex-up profile is characteristic of drainages which cross a crustal upwarp, and Moore (1999) therefore suggests that a crustal flexure may have crossed the course of the Molopo, cutting it off from the Orange. The inability of the Molopo to easily incise through this line of flexure is attributed to the simultaneous capture of aquifers in the Transvaal dolomites by drainages flowing into the Bushveld Basin, as well as climatic deterioration (du Toit, 1933; Bootsman, 1997; Moore, 1999). Further evidence for the crustal flexure comes from further to the east, to the north of Postmasburg, where it appears to separate the northward-draining Ga-Mogara tributary of the Kuruman River from a number of dismembered southwesterly oriented ephemeral drainage lines which terminate in a pan-field.

Wellington (1955) believes that crustal flexuring across the original line of the Upper Fish River caused low and reversed gradients and the resultant formation of the pans. After uplift along the axis the lower Fish was rejuvenated and had vigorous headward erosion. The lower Fish River captured the headwaters of the upper Fish River, with the elbow of capture being marked by a deflection of the modern Fish River to the north of Keetmanshoop to the west of Tses (Wellington, 1955).

Moore (1999) proposes that the northeastern portion of the Griqualand-Transvaal axis of du Toit (1933), can be extended through an area to the north of Postmasburg and up to the northwest where it disrupted the Molopo and Fish River drainages. He named this axis the Etosha-

Griqualand-Transvaal (E-G-T) Axis (Fig. 5.14).

Away from the main Griqualand-Transvaal axis, the differing character and gradients of the tributaries on either side of the watershed between the Vaal and Harts rivers led Mayer (1973) to propose an axis separating the two rivers. According to Mayer (1973) the presence of such an axis would help explain the development of a major pan field to the southeast of the Vaal River (see section 5.2.7) and river capture which modified the former Vaal and Harts drainage lines.

On the basis of disrupted drainage lines to the south of Postmasburg, Moore (1999) suggests that the southwestern portion of du Toit's Griqualand-Transvaal axis be extended eastwards to join up with the Harts Axis which forms the watershed between the Harts and the Vaal rivers (Mayer, 1973). Moore (1999) calls this the Bushmanland-Harts Axis. According to Moore there is no clear evidence for the continuation of this flexure across the Etosha-Griqualand-Transvaal axis, and it may therefore have either terminated close to the latter axis, or merged with it, as proposed by Mayer (1973). Evidence of changes in drainage character of one of the tributaries of the Sak River, suggests that uplift along the Bushmanland-Harts axis occurred in Plio-Pleistocene times (Moore, 1999).

The Kalahari-Zimbabwe axis forms the central watershed between the Zambezi and Limpopo drainage basins in Zimbabwe, before continuing to the southwest in Botswana, where it separates the Limpopo Basin from the Okwa-Mmone fossil endoreic drainage system which formerly emptied into the Makgadikgadi Basin (Moore, 1999). Du Toit (1954) notes that in eastern Botswana, the Kalahari terminates in an abrupt erosional scarp, and infers that the sedimentary rocks of the Kalahari Group must therefore have originally extended further to the east, a conclusion also reached by Mayer (1986) in the North West Province of South Africa. The eastern extent of the Kalahari Group sediments in Botswana coincides with the Kalahari-Zimbabwe Axis and uplift along this axis would have rejuvenated drainages such as the Limpopo and its tributaries, initiating erosion of the coastal side of the flexure, which would thus form the local boundary to the Kalahari Group (Moore, 1999). This provides a lower date for the age of this axis, which must post-date the deposition of the Kalahari Group sediments in this region.

Between the Molopo River and the mekgacha of the central Kalahari (see section 5.3.5) the low,

east-southeast to west-northwest-oriented Kalahari Schwelle is characterised by a broad band of pans (see section 5.3.6). Moore (1999) argues that the Schwelle represents a line of crustal flexure, which is an extension of the Kalahari-Zimbabwe Axis and proposes that the extension of the Kalahari-Zimbabwe axis along this line of uplift be called the Ovamboland-Kalahari-Zimbabwe (O-K-Z) axis (Fig. 5.14). The Otavi axis is associated with a major embayment in the western margin of the Kalahari Basin, and a less pronounced embayment is associated with the Khomas axis. Both embayments extend across the line of the Ovamboland-Kalahari-Zimbabwe axis, and this was interpreted by Moore (1999) as an indication that uplift along the Otavi and Khomas axes and ensuing erosion of the Kalahari sedimentary rocks post-date the former flexure.

Further evidence of neotectonic activity in the southwestern part of the subcontinent is found in the Vaalputs area of the Northern Cape Province of South Africa, where structures of tectonic origin, namely fractures and faults with slickensides, are extensively preserved in Late Cretaceous residual silcretes, in early Cenozoic alluvial deposits of the Dasdap Formation, and in the more recent siltstones of the Vaalputs Formation (Andreoli *et al.*, 1996). The faults bounding the basin in which the sediments of the Vaalputs Formation were deposited, appear to have been reactivated in the period postdating the deposition of the Vaalputs sediments themselves. Overlying sand dunes, possibly belonging to the Gordonia Formation of the Kalahari Group, appear in satellite images to have been truncated by these faults (Andreoli *et al.*, 1996), and south of Vaalputs fractures of Cenozoic age were recorded in the Vanrhynsdorp area (Pike,1959 in Andreoli *et al.*, 1996).

To the north of Vaalputs, in Namibia, the NW-SE trending Kuiseb-Hebron fault downfaults Cenozoic to Quaternary deposits by up to 65 m (Andreoli *et al.*, 1996).

5.3.2 Drainage evolution of the Etosha sub-basin, western Kalahari Basin

The present day Cunene river flows from Angola southeastwards towards the Namibia border where it swings to the east and flows towards the Atlantic Ocean. Some of its flow is diverted towards the Etosha Pan in Namibia through small overflow channels which join other tributaries some 70 km north of the pan and enter it as the Ekuma River. It has long been recognised that prior to the capture of the upper part of the Cunene River in the Pliocene by a river draining towards the coast, the upper Cunene drained into the Etosha Basin (e.g. Beetz, 1933; Wellington, 1939, 1955). Uplift of some 2 km associated with the reactivation of the Omaruru Lineament (Fig. 3.1) about 70 Ma ago (Raab *et al.*, 2002), may have been responsible for accentuating the highlands which blocked the southwards flow of the upper Cunene. Deposition may have been occurring in this area prior to deposition in the rest of the Kalahari Basin, accounting for some of the huge thicknesses of sediment that have accumulated there (Fig. 5.15). The isopachs of the Kalahari Group sediments constructed as part of this study show that up to 450 m of sediment were deposited by the Cunene and other rivers in this part of the Kalahari Basin. These represent the thickest sediments to be found anywhere in the Kalahari Basin.

Stuart-Williams (1992) believes that as recently as 7 Ma the drainage pattern of the Etosha sub-basin continued to be dominated by southward flowing rivers, evinced by all the palaeo-drainages which flow south and west, including the Cuvelei, Oshigambo, Nipele and Omuramba Owambo Rivers, as well as several drainages present under the modern dune fields that originally flowed from the Angolan highlands through into the Etosha Basin. The upper Cunene flowed south and then southwest at this time, joining the Mui-Mui and other small tributaries and flowing into the Etosha Basin where it joined the Hoanib which drained the basin to the west (Stuart-Williams,1992)(Fig. 5.16a).

At around 3 million years ago, uplift on the western side of the basin closed the outlet of the Hoanib River, generating a layer of saline water at least 45 m deep, and giving the lake a surface area in excess of 82 000 km^2. Evaporation progressively salinised the water and all of the sediments underlying it (Stuart-Williams,1992)(Fig. 5.16b). Stuart-Williams (1992) found basal Kalahari Group sediments at 1300 m elevation in the west compared with an average elevation of 1100 m for upper Kalahari sediments in the main part of the basin and suggests based on this evidence that uplift in the west was at least 400 m at the modern basins margin and probably much more. The basin originally extended much further west but has been reduced by erosion. Uplift was almost certainly episodic and the presence of faulting at shallow depth throughout the western part of the basin and evidence of folding suggests that uplift occurred across large parts of the basin.

Further uplift in the late Pleistocene had the effect of rejuvenating all westward flowing drainages, including the Lower Cunene. To the east of the axis of uplift, the opposite effect was true, with the

gradients of the rivers decreasing and in many cases, the rivers became largely dormant. As the Cunene drainage assumed its modern form, so the Lower Cunene captured many of the headwaters of the Etosha Lake System including the Upper Cunene. According to Stuart-Williams (1992) this had the double effect of substantially reducing the water flow into the system and almost certainly allowed water to escape to the sea via the 1100 m outlet of the Mui River and the Cunene River. This outlet is still preserved in the Etaka and Mui Rivers which are one and the same (the Mui River drains north, the Etaka drains south). This caused a geologically recent stabilisation of the system at the 1100 m level, generating the 1100 m shorelines seen at the eastern side of the Andoni Flats and in the west at Okondeka. As the water level dropped so the lower shorelines were generated (Stuart-Williams,1992)(Fig. 5.17a).

Evidence in the sedimentary record for the presence of a large body of water in the region, possibly extending as far back as the Cretaceous, comes from borehole data. The presence of clay layers, sometimes reaching over 100 m in thickness, found throughout the Kalahari Group succession in the area (Miller, 1992a) may be the remnants of the lacustrine deposits. Sedimentary evidence for a lake existing as recently as 35 ka is not forthcoming, however, with an absence of allochthonous material over all but the southern margin of the pans surface (Buch and Rose, 1996), and it is possible that the 35 ka stage of the lake envisaged by Stuart-Williams (1992) was shallow and evaporating, with little or no fluvial input.

Figure 5.17b shows the current condition of the Cunene River drainage system. The proto Lower Cunene and Upper Cunene now form one river system and contribute no water to the Etosha Basin unless high precipitation allows overflow. The Omaramba Owambo, Akazulu, Nipele and Oshigambo rivers are all dry, with current climatic conditions only generating small flows in the Ekuma River.

The current Etosha Pan is thought to be an erosional feature unrelated to the large palaeo lake, and was formed by the joining of smaller pans following scarp retreat (Rust, 1984,1985; Buch and Zöller, 1992; also see section 5.2.7).

5.3.3 Drainage evolution of the central and eastern parts of the Kalahari Basin

5.3.3.1 Current drainage patterns

Figure 5.18 is a satellite image of the central Kalahari, clearly showing the dominant geomorphological features of the area today. The Kwando (or Cuando) River flows southeastwards into the Linyanti swamps before being diverted along faults to the northeast, becoming known as the Chobe before it enters the Zambezi River. To the southwest, the Okavango River flows into the Okavango Delta with overflow from the delta via the Boteti River to the Makgadikgadi Basin. The Makgadikgadi Basin itself is covered by two pans, the Sua and Ntwetwe. The modern course of the Zambezi River (Fig. 5.2) flows in a southerly direction across the Barotse Plain. Near the border between Zambia and Namibia it steepens and changes course to the east. At the Mambova Falls the course of the Zambezi again steepens and follows the down-faulted Karoo sediments of the Mid-Zambezi Basin before breaching the escarpment on its way to the Indian Ocean.

5.3.3.2 Late-Cretaceous and early Tertiary drainage patterns and evolution

The profiles of the Zambezi River above and below the Victoria Falls exhibit separate concave upwards profiles, a phenomenon that is thought to be either due to the up-stream progression of a knick-point (Nugent, 1990), or due to the Middle and Upper Zambezi Rivers having evolved as separate systems (e.g. du Toit, 1927; Thomas and Shaw, 1988; Moore and Larkin, 2001) with the section between Katima Falls and the Mambova Falls providing the link between the Upper Zambezi and the Middle Zambezi (Thomas and Shaw, 1991a). The development of a similarity index (SI), calculated by dividing the number of fish species common to two rivers divided by their total number of species allows for rivers to be compared, with a high SI suggesting the two rivers may have been linked in the past (Skelton, 1994). The low SI for the Upper and Middle Zambezi (table 5.1) suggests that they developed as separate rivers, although Skelton (1994) cautions that a low SI may in part be due to differences in river ecology.

Table 5.1 - Fish species Similarity Indices (SI) between selected southern African rivers (after Skelton, 1994).

Rivers compared	**Combined number of species**	**Species shared**	**Similarity Index (SI)**
Upper Zambezi/Okavango	96	77	0.8
Upper Zambezi/Kafue	97	62	0.64
Kunene/Okavango	93	51	0.55
Save/Limpopo	59	28	0.48
Limpopo/Mid-Lower Zambezi	88	36	0.41
Upper Zambezi/Middle-lower Zambezi	133	36	0.27
Upper Zambezi/Limpopo-Inkomaas-Pongola	122	23	0.19
Limpopo/Orange	48	6	0.13

The abrupt changes in direction of some of the rivers are thought to be indicators of river capture. The Luangwa River changes its southwest flow direction to one towards the south-southeast before joining the Zambezi and the Kafue River also changes course sharply from southerly to easterly (Thomas and Shaw, 1991a; Moore and Larkin, 2001; Fig. 5.2). The area where the Zambezi changes course from southerly to easterly was interpreted as an area of river capture of the upper Zambezi by the Middle Zambezi (Wellington; 1955).

Figure 5.19 shows the possible position of the Cretaceous rivers of the central and eastern Kalahari Basin superimposed on the sub-Kalahari topographical surface generated as part of this study. The massive delta that the Limpopo crosses in Mozambique on its way to the Indian Ocean is believed to indicate a much larger Limpopo Basin than that in existence today and Hartnady (1985) believes the Limpopo may have had its headwaters as far north as Angola. If the upper Zambezi evolved as a separate river system as is commonly believed, it probably flowed southwards through the centre of what is now the Kalahari Basin, before joining either the proto-Orange River (e.g. Lister, 1979; McCarthy, 1983), or the Limpopo River (Wellington, 1955; Thomas and Shaw, 1988; Moore and Larkin, 2001). Moore (1998) cites the occurrence of paired, isolated populations of riverine vegetation on both the Limpopo and Upper Zambezi rivers to argue a link between them.

The

Luangwa and Kafue Rivers may have continued their southwestwards courses to intersect the Zambezi as left bank tributaries as it flowed southeastwards and towards the Limpopo (Thomas and Shaw, 1991a; Moore and Larkin; 2001).

To the west of the Zambezi River, the Okavango and Kwando rivers possibly also originally flowed across the Kalahari Basin and into the Limpopo (du Toit, 1927, 1933; Fig. 5.19). Kalahari Group isopachs constructed from borehole data in southeastern Botswana have revealed a buried river channel (Davidson, 1988). Although the isopachs suggest the river flowed northwestwards into the Kalahari Basin, ilmenites recovered from the vicinity suggest that they originated from the kimberlites at Orapa which lies to the northwest (Moore and Larkin, 2001). According to Moore and Larkin (2001) this channel may represent the original course of the Okavango River. A high similarity index (SI) between the Upper Zambezi and Okavango Rivers (table 5.1) suggests they might have been linked in the past (Skelton, 1994) and while this link may have occurred in the vicinity of the modern Okavango Delta it is also a possible result of both rivers feeding into the Limpopo (Fig. 5.19). The Okavango River may have reached the Limpopo through the headwaters of one of the tributaries of the Limpopo, possibly initially through the Bonwapitse, then the Mahalpswe and finally the Lotsane (Moore and Larkin, 2001). Spits projecting in a northwest direction from the eastern shoreline of Sua Pan are believed to represent the edges of relict fluvial channels, formed in a similar way to those currently forming in the Okavango Delta (Moore and Larkin, 2001). Moore and Larkin (2001) suggest that a drainage system encompassing these spits can be traced southeastwards where it includes a windgap at Mea Pan and a belt of small pans. As section 5.2.7 discusses, the pans may be indicative of old drainage lines. Moore and Larkin (2001) suggest that the Kwando River originally flowed southeastwards through the Sua Pan area and ultimately into the Limpopo via the Motloutse River.

The Cubango and Cuito rivers would have possibly carried on their courses southwards, instead of diverting sharply southeastwards near the current Namibia-Angola border into the Okavango River as is the case today. This current deflection into the Okavango River is possibly an elbow of capture. After joining together, the combined Cubango-Cuito rivers (hereafter referred to as the Bushmanland River) probably flowed towards the SSE before either terminating against highlands coinciding with the western portion of the Ghanzi Ridge, or flowing eastwards towards the Okavango River as suggested by the sub-Kalahari topographical surface (Fig. 5.19). The eastwards deflection of the Bushmanland River along the northern border of the Ghanzi ridge may have occurred along NE-SW and NW-SE trending faults associated with the Damara and Ghanzi Belts. These faults have been reactivated at several times, with reactivation during Karoo deposition and again during Gondwana break-up probably having a significant influence on drainage patterns in the region and can be extended to the northeast where they are currently controlling the deposition in the Okavango Delta. The course followed by the Bushmanland River may follow an older NW-SE-trending structure named the Khoisan Lineament by Corner (2003). Similar NW- trending faults control the pan-handle of the Okavango Delta, the orientation of small rivers draining into the Thamalakane and Kunyere rivers, like the Nxotega and Shashe rivers (Modisi, 2000) and probably controlled the southeastwards course of the Kwando River. Reactivation of NW-trending structures may also have been responsible for uplift of the NW-trending high-lying area separating the Bushmanland River from the Okavango River.

Down warping of the Kalahari Basin (du Toit, 1927, 1933), possibly together with late Cretaceous block faulting in northern Botswana (du Plessis and le Roux, 1995) may have caused back tilting of the drainage away from the Limpopo and into the basin resulting in huge lakes forming in the Kalahari Basin. At this stage the upper Zambezi may have existed as an endoreic river system, flowing into the Kalahari Basin in much the same way as the Okavango does today (e.g. Thomas, 1984b). This back tilting of the drainage would explain why the buried channels in southeastern Botswana (Davidson, 1988; Moore and Larkin, 2001) suggest northwestwards flow, although Moore (1999) and Moore and Larkin (2001) suggest that it was uplift along the Kalahari-Zimbabwe Axis that severed the link between the Limpopo and the rivers to the northwest. This will be discussed further in Chapter 8.

5.3.3.3 The East African Rift System

During the early Tertiary the Middle Zambezi was aggressively capturing some of the tributaries of the Upper Zambezi, (including the Luangwa) and rifting along the Gwembe and Chicoa troughs aided its headward erosion (Thomas and Shaw, 1991a). Moore and Larkin (2001) suggest that the Luangwa was captured by the Middle Zambezi around 38 Ma, but note that this capture cannot be accurately dated. The Kafue was probably not captured until the Upper Pleistocene. Derricourt (1976), Lister (1979) and Thomas and Shaw (1990) believe that the capture of the Upper Zambezi by the Middle Zambezi occurred in the Pliocene or early Pleistocene. McCarthy *et al.* (2002) suggests that during the same time movement on faults blocked the southwards path of the Kwando and Okavango Rivers, damming them up in the Okavango Delta. This faulting was possibly related to the southwest extension of the East African Rift System (e.g. du Toit, 1926,1927; Fairhead and Girdler,1969; McConnel, 1972; Scholz *et al.,* 1976).

The East African Rift System (EARS) extends along a broad intracratonic swell down much of the length of Africa, from Ethiopia to Mozambique in the south (Figs 3.14, 5.20). The EARS consists of two main branches, the Western and Eastern (Kenya or Gregory) rift valleys each of which have certain individual characteristics. The southern junction of the western and eastern branches of the EARS is obscured by volcanics, but south of this, the rift continues through the Rukwa Rift, the Nyasa/Malawi Rift and south into Mozambique through the Shire and Urema Rifts to the Mozambique coast. Off the main line of the EARS there are no Tertiary rifts (with the possible exception of the Okavango and Makgadikgadi areas), but the older late and post-Karoo rifts shown in blue on Figure 3.14 form an important link between the EARS and the Kalahari Basin. The northeast-southwest trend of the Luangwa and Mid-Zambezi Basins can be continued into Botswana, where older rifted features like the Koras-Sinclair-Ghanzi rift may have been reactivated several times in the past to form deep Karoo- and Kalahari-filled features.

Although there is debate as to the extent in which the EARS is exploiting the positions of older Karoo and Post-Karoo rifts, the Cenozoic reactivation of faults and ongoing seismicity in some of the older rifts suggests that in some cases there is indeed an ongoing relationship. According to Dixey (1956) the Tertiary rifting represents a continuation of rifting initiated in Mesozoic and even Palaeozoic times. Fairhead and Stuart (1982) attempted to show that the orientation of much of the EARS is controlled by the orientation of the basement rocks, in particular of the Pan-African belts and Lambiase (1989) gives an example of the southern part of Lake Tanganyika where a Tertiary lake basin overlies the southern continuation of the Permo-Triassic Luama rift, which extends to the northwest. There are, however, many cases where the EARS does not follow older structures or rifts and Tertiary rifts may cut completely across the older rifts. An example of this is the truncation of the Ruhuhu Graben and the Maniamba Rift by the younger Malawi Rift without any apparent reactivation of the older rifts (Lambiase, 1989). The link to older structures or zones

of structural weakness, although important in some cases, should therefore not be seen as the main factor controlling the nature and orientation of the EARS.

Evidence from the volcanic-stratigraphic sequence of the eastern rift suggests that the sequence of events was one of initial volcanism, followed by doming and then rift faulting (Sahagian, 1988). According to Sahagian (1988) the western rift is cooler than the eastern rift, an observation seemingly confirmed by the lower heat flow and geothermal gradient and the lower incidence of volcanic activity associated with the western rift. The sequence of events that occurred in the western branch was also different and in contrast to the eastern branch. The western rift started with initial Miocene downwarp or subsidence, followed by faulting and formation of rift grabens and finally by Pliocene volcanism (Logatchev *et al.*, 1972; Sahagian, 1988). In spite of these differences, however, the two branches are similar in that they both consist of a zone of thinned lithosphere where normal faulting has taken place along an elongated zone of a few hundred kilometres.

The causes of the uplift and formation of the EARS are debatable. The African Superswell has been linked in the past the formation of the East African Rift System (EARS), with Quennel (1960) suggesting that a genetic relationship exists between the formation of the EARS and uplift of east Africa. According to Quennel (1960) the formation of the horsts and grabens associated with the EARS is linked to isostatic readjustment following the failure of the uplifted and therefore extended crust. According to Sahagian (1988) a combination of hot spot heating and rift shoulder stretching caused the uplift, but the ongoing development of the rift after hot spot migration suggests that more than one hot spot may have existed, resulting in the tensional stress present in the lithosphere. The thinning of the continental lithosphere and rifting resulting from these tensional stresses may have resulted in further uplift of the shoulders of the rift, resulting in some of the anomalous topography found in east Africa.

The development of the EARS started between 35 and 25 Ma, and has continued its development, episodically in some instances, for the last 25 Ma. Between 35 and 25 Ma, the development of the EARS was dominated by the eruption of the Afar plume and the initiation of the Red Sea Rift, which was well developed by 25 Ma. At about 30 Ma the rifting began to intensify, and the first rift structures in Kenya, at Turkana, began to develop, with some elevation further to the south on the future site of the Kenya Rift (Burke, 1996).

At around 23 Ma, the Samburu flood basalts erupted in the central plateau area of Kenya, possibly caused by a plume, and rifting propagated from Afar into southern Ethiopia. At the same time, in the Early Miocene uplift of the eastern part of the plateau began (Ebinger, 1989; Burke, 1996). Between 23 and 16 Ma subsidence of the Turkana depression continued with some volcanism (Logatchev, 1972).

Between 15 and 5 Ma half-graben development at Turkana continued, accompanied by more volcanism and sediment deposition. At about 13 Ma the area of volcanic activity shifted to the Kenya domal uplift and massive volcanic activity along the strike of the future Eastern Rift occurred, preceding the eruption between 10 and 5 Ma of the Kenya and Kilimanjaro volcanoes (Logatchev, 1972). It was during this period that the formation of the Western Rift began, with its initiation possibly related to stress changes caused by the Zagros collision at about 15 Ma (Burke, 1996). Doming began at about 20 Ma and volcanism started at about 12 Ma in the north of the rift, and about 7 Ma in the south (Ebinger, 1989). The development of the basins of the western rift began either just prior to the volcanism, or at the same time, with propagating faults linking individual basins over time, and the main phase of tectonism occurred in the last 5 Ma (Braile *et al.*, 1995). The current Western Rift stretches northward into Sudan, and according to Girdler and McConnell (1994), seismicity and faulting in the region suggests that the rift extends as far as 5.5°N.

The period from 5 Ma to the present was characterised by the southward propagation of the rift at Turkana to form the Gregory Rift. This extension of rifting may have been in response to the thinning of the lithosphere above the Samburu plume (Burke, 1996). At around 1 Ma volcanism and major faulting occurred in present day Tanzania in the Eyasi and Pangani grabens.

Tectonic activity continues to the present along much of the length of the EARS, and possible extensions of the EARS, towards the east in particular, have been hypothesised. Trends of seismicity make up a compelling contribution to these theories and will be discussed in the following section.

The extension of the East African Rift System to the southwest

The southward extension of the East African Rift System (EARS) has been the subject of debate for many years, with possible branches of the rift being recognised around the edges of the Zimbabwe Craton (e.g. Vail, 1967), as well as extensions of the eastern branch of the rift to the Indian ocean on the Tanzanian coast (Mougenot *et al.*, 1986) and the western branch down to the southern Mozambique coast (Mougenot *et al.*, 1986). There have also been proposals (described in McConnell, 1967) of an extension of the eastern branch of the EARS through the Zimbabwe craton, following the Great Dyke of Zimbabwe, and continuing in a trend that includes the Bushveld Igneous Complex, Vredefort Dome and Trompsberg anomaly of South Africa. The latter proposal did not have the benefit of the identification of Vredefort as a meteorite impact site (e.g. Reimold and Gibson, 1996) but more recently Corner (2003)

As Figure 5.21 shows, the western branch of the EARS is characterised by a large number of seismic events running down through Lake Malawi and along the eastern escarpment, before it branches southeastwards towards the Indian Ocean and westwards along the Limpopo valley. Also visible are a number of southwesterly trending branches of seismicity, which have long been recognised by various researchers (e.g. Fairhead and Girdler, 1969; Reeves, 1972a,b). The southernmost concentration of these seismic events, in a northeast-southwest belt in South Africa has been proposed as possibly being due to either an extension of the EARS, a sublithospheric hotspot, or even a combination of the two (Hartnady, 1990) and may be related to the development of the African Superswell (Hartnady and Partridge, 1995). Figure 5.22 highlights four zones of southwest trending seismicity which are felt to be particularly relevant to the development of the Kalahari Basin.

Zone 1 roughly follows the trend of the Kibaran Belt, and extends through the Upemba Graben in the Democratic Republic of Congo, through Zambia and into Angola (Bram, 1972).

Zone 2 strikes southwest from the southern end of Lake Tanganyika, through Lake Mweru and Lake Tshangalele, before terminating at about 15°S (Fairhead and Girdler,1969; Bram, 1972; Fairhead and Henderson, 1977). According to Fairhead and Girdler (1969) this zone may then extend due south through eastern Botswana and into South Africa as far south as 24°S, although

this author finds little seismic or geological evidence to substantiate this view.

According to Fairhead and Girdler (1971) both the Upemba fault and the Lake Mweru faults are still active, and both the area to the west of Lake Tanganyika at 3°S and the Upemba-Mweru area have seismic as well as geothermal activity.

Zone 3 defines the zone of more diffuse earthquake epicentres in western Zambia centred around 13.7°S and 26°E (Bram, 1972) and also recognised by Fairhead and Girdler (1971) who described it as a zone extending from Lake Tanganyika to west of Lake Kariba at about 18°S. According to Fairhead and Henderson (1977) the zone of diffuse seismicity is due to normal faulting along a series of reactivated NE-SW faults that are related to the Pan-African belts.

Zone 4 extends from Lake Malawi down through the Luangwa Valley of Zambia, through Lake Kariba and into Botswana where it may branch to the west, under the Okavango Delta in another distinct cluster of epicentres possibly related to movement on the faults controlling the delta itself (Fairhead and Girdler, 1971). Some of the epicentres under Lake Kariba and the Okavango Delta may be related to sediment and water loading. Possibly related to this zone of seismicity and more closely aligned to the extension of Zone 4 is a southwesterly trending zone of epicentres to the south of the Ghanzi Ridge to which Reeves (1972b) fitted a linear regression, which he termed the "Kalahari Seismicity Axis" (Fig. 5.22). This line is thought to coincide roughly with a buried southern boundary of the Ghanzi-Chobe Belt (Reeves and Hutchins, 1975) and also coincides approximately with the Makgadikgadi Line, possibly marking the northern boundary of the Kaapvaal Craton (see Fig. 3.4).

Various studies (eg. Gregory, 1921; du Toit, 1926,1954; Shaw and Thomas, 1992; McCarthy *et al.,* 1993) have proposed that the EARS continues through the Luangwa-Middle Zambezi rifts and into Botswana. It has been hypothesised that the observed zones of seismicity are related in some manner to the EARS, and that NW-SE tensional stress field associated with these zones is related to the crustal extension observed in the EARS (e.g. Bram, 1972; Fairhead and Henderson, 1977). A study of micro-earthquake activity in northern Botswana in a region stretching along the southeastern border of the delta towards the Zambian border in the northeast was conducted in 1974 and it was found that the rate of seismicity was found to be comparable to that in the EARS

(Scholz *et al.*, 1976). Scholz *et al.* (1976) postulated that the tectonic setting of the Okavango Delta appears to be that of a developing graben some 150 km wide, with the delta at the tip of a proposed incipient zone of rifting that follows an old post-Karoo, pre-Cretaceous rift zone. Gumbricht *et al.* (2001) suggest, however, that the Okavango Delta does not appear to occupy a rift, but rather a depression between two basement arches at the tips of southwesterly propagating rifts. According to Gumbricht *et al*. (2001), upthrow to the southeast of the northeasterly striking Thamalakane and Kunyere faults resulted in the formation of the Ghanzi Ridge, which now separates the Kalahari Basin into two sub-basins. This ridge is believed to have risen some 100 m since the Okavango started developing, and extends to the northeast where it is cut by the axial rift extending from the EARS, down through Luangwa and Kariba. The northwestern boundary of the Okavango depression is characterised by thinning of Kalahari Group sediments (see Kalahari isopach map), and pre-Kalahari rocks outcrop along the Okavango River, above the pan handle, at Popa Falls. Gumbricht *et al*. (2001) believe uplift of the area to the northwest of the Okavango has taken place along the Gumare Fault, which they interpreted as a horst bounding fault, and that the uplifted area is related to a southwestward extension of the Mweru-Tshangalele-Kabompo Rift.

To the northeast of the Okavango and Linyanti swamps, in Zambia, the Bangweulu and Lukanga Swamps, may have formed between the uplifted flanks of the Mweru-Tshangalele-Kabompo and Luangwa-Kariba rifts, or alternatively are themselves in a southwestwards propagating rift the tip of which is manifested as Linyanti, Chobe and Gomare Faults.

According to Scholz *et al.* (1976), the African rifts represent an incomplete, growing plate margin which is propagating to the south along at least two arms that bifurcate south of Lake Tanganyika. Coblentz and Sandiford (1994) used an elastic finite-element analysis of the African intraplate stress field and found that there are large extensional stresses present in the Ethiopian highlands (15 Ma), the East African Rift (9 Ma), and southern Africa. They therefore predict that the ambient state of stress in the continental lithosphere is extensional and work by Banghar and Sykes (1969) and Maasha and Molnar (1972) suggests tensile deviatoric stress in an east-south-east to west-north-west direction. The Kalahari Seismicity Axis of Reeves (1972b)(Fig. 5.21) does appear to continue the trend of the Luangwa Rift and Mid-Zambezi Basin into the southwestern parts of the Kalahari Basin. Evidence for neotectonic movement along faults in the Makgadikgadi Basin comes from the elevations of shorelines and ridges (e.g. The Magagikwe and Gidikwe ridges) which clearly show that neo-tectonic activity in the vicinity of the Okavango Delta and Makgadikgadi Pan has led to tilting of the palaeo-shorelines, largely through continued movement along bounding faults (Gumbricht *et al.*, 2001). The effect of neo-tectonic activity on water dispersal in the Okavango Delta itself has also been documented (e.g. Shaw, 1984a,b, 1985; McCarthy *et al.*, 1993), with movement on faults diverting water flow between interconnected grabens. Further evidence of neotectonic activity is recorded at the nearby Gcwihabe Hills, where Cooke (1975a, 1980) described faulting of Late Pleistocene calcretes and cave deposits and Mallick *et al.* (1981) described the truncation of sand dunes by faults. Neotectonic subsidence in the Okavango-Linyanti depressions has been estimated at between 300 m (Greenwood and Caruthers, 1973) and 1000 m (Hutchins *et al.*, 1976). The crustal sagging underneath the delta is thought to be occurring partly as a result of the accumulation of sediment, in turn resulting in the seismicity associated with the delta and causing localised depression and faulting of the southeasterly arch (Gumbricht *et al.* (2001; McCarthy *et al.*, 2002).

5.3.3.4 The Okavango Delta and Makgadikgadi Basin

The topography of the Okavango Delta is fairly flat, and elevation contours superimposed on a Landsat TM image of the Okavango region (Fig. 5.23) shows that the drop in elevation from Shakawe at the top of the panhandle to Maun, which is situated on the Thamalakane Fault, is only about 60 m (McCarthy *et al.*, 1997; Gumbricht *et al.*, 2001). The lowest point in the delta is 920 m in the Mababe Depression, which has an area of ~3000 km^2, and to the west is the Ngami Basin (1800 km^2) which is occupied by Lake Ngami (Shaw, 1986). Further to the northeast, Lake Caprivi is a marshland formed upstream of the Mambova Falls.

The distal end of the delta is confined by the southwesterly trending Thamalakane and Kunyere faults (Figs 5.24), with two more major faults to the east of the delta having been interpreted from aeromagnetic and elevation data (Modisi, 2000). These major northeast-southwest trending faults appear to form an en echelon pattern resulting from remote regional extension in an east-west direction (Modisi, 2000)(Fig. 5.25). A study by Scholz *et al.* (1976) showed that most of the seismicity in the area originates from movement on the Thamalakane and Kunyere faults, and Hutchins *et al.* (1976) show that the origin of all the hypocentres for the area are found along their fault planes. A composite focal mechanism for events in the southeast of the delta suggests the

faults dip at 60° to the northwest (Scholz *et al.*, 1976). The main orientation of the faults is northeasterly, with two main trends - the faults associated with the Ghanzi ridge at 58° and the Thamalakane and Kunyere faults trend at 38°. The northwest boundary of the Delta is marked by the Gomare ridge (or Gumare Fault in Gumbricht *et al.*, 2001)(Fig. 5.24) which has a trend midway between these values. There is a possibility that the 38° trend on the Kunyere fault crosses the 58° trend in the neighbourhood of Toteng (Hutchins *et al.*, 1976). Down throw on the Kunyere, Thamalakane, Linyanti and Chobe faults is to the northwest and downthrow on the Gumare fault to the southeast (Hutchins *et al.,* 1976; McCarthy *et al.*, 2002). Greenwood and Carruthers (1973) found using a seismic refraction survey that the throw on the Thamalakane fault is about 117 m.

The Okavango Delta and Linyanti Swamp were thought to have developed in a graben or series of grabens (e.g. Shaw and Thomas, 1992) and the increase in the thickness of sediments towards the Thamalakane and Kunyere faults in the southeast (Reeves, 1979) has been interpreted as being due to the presence of a half-graben structure containing the delta (McCarthy *et al.*, 1993). The Thamalakane Fault follows the strike of sub-surface Precambrian rocks (Reeves,1972a, Mallick *et al.*, 1981) and Modisi (2000) and Modisi *et al.* (2000) believe that the northeast-southwest trending faults are a reactivation of older basement structural trends.

The orientation of the Okavango River as it enters the delta is thought to be controlled by parallel faults at the "panhandle", which may form grabens or half-grabens controlling drainage (e.g. McCarthy *et al.*, 1993;1998), although the offset of contours across the flood plain of the pan handle (Fig. 5.23) could imply that the pan handle has developed along a single northwest-trending fault (Gumbricht *et al.*, 2001). Similar northwest-southeast trending faults also control the orientation of several small rivers draining into the Thamalakane and Kunyere rivers, with the Nxotega and Shashe rivers being the most notable of these (Modisi, 2000).

To the southeast of the Okavango Delta the now largely dry Makgadikgadi Basin (Figs 5.17, 5.24) was also possibly controlled by similar northeasterly trending faulting (Baillieul, 1979), and it is not coincidental that it lies astride the Kalahari Seismicity Axis identified by Reeves (1972b). The Nata drainage network that enters the Makgadikgadi Pans from Zimbabwe was probably influenced by faulting along the Nata fault which enabled the Nata river to capture the Tegwani and Maitengwe streams by altering its course to truncate their lower reaches (Mallick *et al.*, 1981).

More recent movement on the Nata fault has disrupted dunes and resulted in the partial burial of the linear dunes on the downthrown eastern side of the fault by alluvium (Mallick *et al.*, 1981). Inflow from the northern side of the Makgadikgadi Basin was also influenced by tectonic movement, and it is thought that the Chobe (Kwando) originally flowed south-eastwards, parallel to the Okavango, and possibly into the Makgadikgadi Basin. It is now, however, deflected by a number of northeasterly trending faults (the Linyanti and Chobe) towards the east where it joins up with the Zambezi at Kazungula (Mallick *et al.*, 1981). The Mababe depression formed between the Kunyere and the Chobe faults, and in historical times has received water from the Linyanti swamps via the Savuti channel (Gumbricht *et al.*, 2001). Closer to the Makgadikgadi Basin itself, Wright (1978) and Mallick *et al.* (1981) identified a series of northeast-trending faults that have partially disrupted the linear dune system and controlled the drainage entering the basin, and the development of the sub-basins within the Makgadikgadi Basin. Between the Thamalakane fault and the Makgadikgadi depression, the Boteti River flows through a depressed area of topography called the Makalamabedi depression (Fig. 5.23). The depression is bordered on its southeastern side by a possible fault (the Makalamabedi fault) along the southern side of which the Moremaoto ridge extends (Gumbricht *et al.*, 2001). The Gidikwe Ridge, which occurs to the immediate east of the Moremaoto Ridge (Fig. 5.23), was thought either to be part of the shoreline of palaeo-lake Makgadikgadi (Grove, 1969; Cooke and Verstappen, 1984), or formed as an offshore sand bar in an even greater lake (Grey and Cooke, 1977). A small lake is also thought to have existed in the Makalamabedi depression (Shaw *et al.*, 1988). The Boteti River cuts through both the Moremaoto and Gidikwe Ridges, and Gumbricht *et al.* (2001) believe that the present course of the Boteti has been superimposed with incision by the river keeping place with fault displacement on the Thamalakane and probably the Makalamabedi faults.

Ancient shorelines show that at various stages in the past, large lakes covered extensive parts of northern Botswana. The presence of beach ridges and shorelines in Ngami, Mababe and Makgadikgadi at 945 m suggests a common lake level and has led some researchers to believe that at some stage the Ngami-Mababe-Makgadikgadi area was unified into a single system, with water from Lake Thamalakane overflowing via the Boteti River into the Makgadikgadi Basin (e.g. Shaw, 1988). This system was given the name Lake Palaeo-Makgadikgadi (Grey and Cooke, 1977) and is thought to have covered over 60 000 km^2 (Shaw, 1988; Shaw *et al.*, 1988). According to Grove (1969) a lake at 945 m would have required some 50 km^3 of water a year to sustain it. While Ebert and Hitchcock (1978) have calculated that rainfall increases of up to three times the present levels would have been needed to fill the lake, Grove (1969) and Cooke (1980) suggest that it would have taken inflow from the Zambezi to provide that amount of water. A lower level at 920 m has been recognised in the Makgadikgadi Basin (Cooke, 1984) with ^{14}C ages of between 40 200 and 10 070 BP (Thomas and Shaw, 1991a). An OSL date taken from diatomaceous earths in the lower Boteti River has given an age of 32 to 37 ka (Shaw *et al.*, 1997), and this is believed to represent the age of the lake maximum. Further to the northeast, Lake Ngami occupied a shoreline at 936 metres above sea level, some 20 m above its present level, and similarly, the Mababe Basin and Lake Caprivi both display evidence of palaeo-shorelines well above the present levels, and are characterised by shorelines of 936 m and bounding ridges of approximately 940 m (Shaw, 1986; Shaw *et al.*, 1988; Partridge and Scott, 2000). It is thought that at the 936 m level, water covered much of the Okavango-Ngami-Mababe-Caprivi areas about 17-12 ka (Shaw, 1986) forming a lake named Lake Thamalakane (Shaw,1988), and would have overflowed at this level, into the Savuti and then the Chobe rivers (Shaw, 1986; 1988), with water possibly ponded up behind the Mambova Falls in a lake (Lake Caprivi)(Shaw, 1988; Shaw and Thomas, 1988, Shaw *et al.*, 1988) before reaching the Zambezi. Moore and Larkin (2001) believe that the high lake levels were sustained by water diverted from the Upper Zambezi into northern Botswana along the Linyanti and Chobe faults. The diversion of water from the upper Zambezi may have reoccurred periodically with the reunification of the Upper and Lower Zambezi occurring after each of these events (Moore and Larkin, 2001). This may go some way to explaining the flood evidence presented by Nugent (1990) who believed that unification of the Middle and Upper Zambezi drainage systems occurred about 125 ka, an interpretation that was criticised by Thomas and Shaw (1992).

5.3.4 The Congo Basin

The Kalahari Group sediments are thought to continue up through Angola where the basin narrows before opening up again in the Democratic Republic of Congo where similar sediments were deposited to the north and west of the African Superswell in a basin that is thought to have originally developed as a result of Late Proterozoic failed rifting and subsequent thermal sag (Daly *et al.*, 1992). This Congo Basin (or Cuvette Centrale) is not covered in detail by this study largely because of the limited amount of information, and also because the prohibitive task of gathering what data does exist is beyond the time frame allowed for this investigation. It is, however, important to note that the Kalahari sediments may very well extend into the Democratic Republic of Congo and enormous scope for future work exists there.

Borehole and geophysical evidence from the Congo Basin suggests that sedimentary deposition in the Precambrian to Proterozoic occurred in a basin which could be roughly divided into northern and southern sub-basins (Giresse, in press). These rocks were in turn overlain by a Cambrian clastic sequence, and a Late Ordovician to Devonian shale and arkosic sandstone sequence (Daly *et al.*, 1992; Giresse., in press). Carboniferous-Permian glacial deposits found over these rocks (the Lukuga Formation) are believed to be the equivalent of the Dwyka and Ecca Groups, but an equivalent of the Beaufort Group is missing, with a major unconformity existing between the Lukuga and rocks thought to be the equivalent of the Upper Karoo (Stormberg equivalent) (Giresse, in press).

Deposition of sediments in the Cretaceous followed relative minor subsidence, not believed to be linked to the break-up of Gondwana in any way (Giresse, in press). Cretaceous deposits belonging to the Kwango Formation were overlain by sediments of the Kalahari Group (see section 4.3.7) and the Congo Basin at this time could be divided into the Kwango Basin and the Cuvette Centrale (Giresse, in press).

Uplift in the mid-Miocene (Bostrom, 1985) or mid-Oligocene (K.C.A. Burke, *pers. comm.*), firstly to the east of Lake Victoria, associated with the Eastern branch of the EARS, and subsequently to the west of Lake Victoria where the western branch began to form around its edges (K.C.A. Burke, *pers. comm.*), formed an eastern watershed and established the north-south course of the upper Congo River in this region (Veatch, 1935). Prior to this Tertiary uplift the Congo River may even have flowed eastwards into the Indian Ocean (Stankiewicz and de Wit, 2004). The northward tilting of the southern edge of the current Congo Basin resulted in parallel northward flowing rivers, originating in the northern border areas of present day Angola and Zambia (Bostrom, 1985) and the northern side of the basin is defined by the uplift associated with the Ngaoundere-Abu Ghabra rift system (Bostrom, 1985). Bostrom (1985) summarises some of the findings of previous investigators, which suggest that until the end of the Tertiary, the Congo River emptied into a large lake which had a small outlet on its northern side. The uplift associated with the Ngaoundere-Abu

Ghabra rift system is thought to have resulted in the capture of the Congo by a small coastal river, which now forms the lower course of the present Congo River. This hypothesis of Late Tertiary river capture is debatable, however. Burke and Dewey (2002) believe that the Congo River developed in the last 30 million years and P. Giresse (*pers. comm.*) believes there is sufficient evidence of an early Cenozoic fan having already been formed by the Congo River.

5.3.5 The mekgacha of the central southern Kalahari Basin

The mekgacha (or mokgacha) are the dry valleys found throughout the Kalahari (Fig. 5.26) and include the Molopo River system which has already been discussed in some detail. The Okwa-Mmone system is the largest of the Mekgacha systems, with a potential catchment of 90 000 km^2 (Thomas and Shaw,1991a) stretching from the Namibian border in the west right across central Botswana and almost as far as the Makgadikgadi Basin. The mekgacha very seldom have any surface flow, but after heavy precipitation water may flow for short distances along their length.

According to Thomas and Shaw (1991a) the mekgacha have a distinct morphology, with the initial catchment area characterised by shallow and broad valley floors filled by a flat surface of clayey material. This initial morphology changes "downstream" to one of incised, rectilinear, flat-bottomed valleys with steep sides and abrupt valley heads. The valley sides are comprised largely of Kalahari Group sediments that have been calcretised, silcretised, or both, and are commonly capped by a layer of unconsolidated sand belonging to the Gordonia Formation. As the mekgacha leave the edges of the Kalahari Basin and approach its centre, their morphology again changes with the relief of the valleys decreasing until they become fairly flat, linear depressions, which may follow the interdune areas.

The mekgacha are considered to have been formed during wetter climatic conditions in the Late Tertiary or Quaternary (Grove, 1969). The mode of formation of the mekgacha is commonly supposed to be as a result of erosion during periods of increased precipitation. Shaw *et al.* (1992) showed that the Okwa/Mmone system may have flowed between 15000 and 12000 years ago into Lake Makgadikgadi and the deltaic sediments at the end of the Okwa (Cooke and Verstappen,1984; Shaw *et al.*, 1992) and the Groot Laagte (Thomas and Shaw, 1991a) are

examples of the effects of periodic flow along these valleys. Work on the morphology of the mekgacha by Shaw and de Vries (1988), Thomas and Shaw (1991a) and Nash *et al.* (1994a) does, however, show that the valleys were not formed by fluvial action alone, and that the influence of groundwater should be considered. The location of mekgacha appears to have been controlled by faults and fractures as well as deep weathering of rocks beneath the mekgachas, both of which accentuated the effect of groundwater movement along their reaches (Thomas and Shaw, 1991a; Nash *et al.*, 1994a; Nash, 1995). The thickness of Kalahari Group sedimentary rocks appears to have an effect on the extent to which geological structures and basement weathering influences valley alignment, however, with thicknesses of sediment of over 30 m appearing to prevent a relationship between valley and structural orientations (Nash, 1995). The presence of springs in the headwater sections has significantly contributed to mekgacha formation (Thomas and Shaw, 1991a) and processes of sapping and deep water weathering by groundwater have significantly aided mekgacha development (Shaw and de Vries, 1988; Nash *et al.*, 1994a,b). In spite of these factors influencing the formation of the mekgacha, the importance of tectonic uplift and river capture must also be considered in their formation and tectonic uplift was recognised by Shaw and de Vries (1988) as a mechanism for initiating groundwater sapping in the southeastern Kalahari.

5.3.6 Pans

5.3.6.1 Introduction

The pans of the Kalahari are generally less than 3 to 4 km in diameter, but in some areas can be up to tens of kilometres wide. They are generally fairly shallow features, up to 20 m deep, and may be flanked by dunes up to 30 m high (Lancaster, 1978b). In the Northern Cape Province Levin (1980) found that the northwestern rim of the pans was the steepest, whereas the southeastern edge of the pan showed no sign of erosion. Most pans in southern Africa occur on the arid side of the 500 mm mean annual isohyet, but are also extremely common in Hwange National Park in Zimbabwe where 2449 pans occur in an area that receives about 600 mm of rainfall annually (Goudie and Thomas, 1985). These pans are generally smaller than those found in the southern Kalahari. The largest pans occur in the Makgadikgadi and Etosha Basins which have been discussed earlier in the text and a broad band of pans stretches across southwestern Botswana.

5.3.6.2 Formation of pans

In general, the development of all pans has one factor in common, namely wind deflation of the pan surface. The factors leading to the original formation of the pan may, however, be more varied and complicated, with different geomorphological settings providing the conditions suitable for pan development.

(1) Disruption of drainage patterns

The disruption of drainage patterns by various processes has had a major influence on the formation and distribution of pans. Along drainage lines that are drying up for climatic or even tectonic reasons, sediment may block the old channels, or dunes from surrounding areas may move across them. In cases like these, the formation of shallow ponds of water may lead to salt accumulation, which in turn increases the possibilities of wind deflation, as is discussed later. Pans found along old drainage lines have been described in Botswana (Boocock and van Straten, 1962) and Namibia (Wellington, 1955). In the Gordonia district of South Africa they have been shown to have a definite relationship to palaeo-drainages (Levin, 1980) and are found along the old courses of the Koa, Eeenbeker and Tellerie Rivers (Malherbe *et al.*, 1986)(Fig. 5.27). Observations made during this study suggest that the Sewe Panne (Fig. 5.28, 5.29) in the Kalahari Gemsbok Park may have formed in a similar fashion and some of the pans are separated from either each other, or the remnants of the old drainage course, by nothing more the outer lunette dune. The control of drainage patterns by landforms will also have an important influence on the orientation of pans, and pans can commonly be found to be either associated with the occurrence of dolerite dykes which have locally disrupted drainage patterns (Goudie and Thomas, 1985), or are found in interdune troughs where rivers may once have flowed (Mallick *et al.*, 1981). Lancaster (1978a) found that the pans of southern Botswana mostly occur in a belt oriented along a watershed (the Kalahari Schwelle) separating the Okwa and the Molopo drainage systems (Fig. 5.29). The tectonic significance of the pans occurring along this watershed will be discussed in Chapter 8.

(2) Structural and erosional relationships

The common occurrence of pans overlying dolerite dykes, faults or other geophysical anomalies or linear features has been documented in Botswana (e.g. Mallick *et al.*, 1981; Farr *et al.*, 1981,1982). Mallick *et al.* (1981) suggested that pans can be formed by weathering taking place

as a result of the movement of groundwater along older sub-Kalahari features and Wormald *et al.* (2004) found a strong relationship between regional structural trends and pan distribution in Botswana. Where the underlying bedrock is a calcrete, the relationship between structure and erosion can be particularly strong and Bruno (1985) found that pans in the southern Kalahari are commonly formed in depressions, either overlying calcrete horizons, or along tectonic fractures formed in the calcrete surface. While Bruno (1985) believed that the depressions in the calcretes formed as a result of displacive growth of the calcite crystals, solution cavities or sink holes caused by the dissolution of calcretes or dolomites may also result in pan formation (Goudie and Thomas, 1985).

The pans in the Etosha Basin are thought to be developing on calcrete surfaces as a result of pan-edge retreat under phases of pluvial erosion during seasonal rainfall, and deflation by wind during dry periods (Rust, 1984,1985; Buch and Zöller, 1992). The Etosha Pan itself is believed to have formed as a result of the joining of several smaller pans following these erosive events (Rust, 1984,1985).

(3) The importance of salinity

One of the most important factors in the formation of pans is the presence of salt. Early views on the origin of pans centred around the formation of pans by animals congregating around deposits of salts (e.g. Alison, 1899; Passarge, 1904) and the presence of salt increases the rate of weathering of rocks through processes of hydration and crystallisation. The most important effect of high salinity, however, is that it discourages vegetation growth, with the pan sediments being exposed to the influence of wind, and dry materials being removed by deflation. The removal of exposed sediment by wind action is seen as the major cause of pan formation (e.g. Grove, 1969; Lancaster, 1978a; Goudie and Wells, 1995) and evidence for wind action can be seen by the alignment and elongation of pans along the prevailing wind directions (Grove, 1969), the difference in steepness of the pan edges on the windward side (where the slope is steep) and the leeward side (with a more gentle slope) (Goudie and Thomas, 1985), and the fairly widespread occurrence of lunette dunes on the leeward side of the pans. The nature of the lunette dunes themselves is discussed in section 4.2.5.3. Increased salinity in the soil or in shallow water in depressions can be achieved in several ways:

- By the action of termites on the edge of watering holes which not only results in the creation of sub-surface cavities, but also aids the accumulation of the salts (Thomas, 1988d).
- By capillary action moving salts contained in groundwater to the surface. Related to this may be the underlying bedrock and, according to Goudie and Thomas (1985), the greatest number of pans are found on the unconsolidated Kalahari sands and non-resistant Karoo Supergroup sediments like the Dwyka and Ecca shales. The weathering of Dwyka and Ecca shales appears to release larger quantities of salt or chloride than some of the other lithologies, and it has been found that saline groundwater is commonly associated with these lithologies (E.Van Wyk, *pers. comm.*)
- By the evaporation of water causing the concentration of salts over time. According to Bruno (1985) the water being evaporated does not necessarily have to be saline in the first place, as even the evaporation of fresh water over time can result in salinity in a pan.

5.3.6.3 Conclusions

It is important to note that all of the different mechanisms for the formation of pans are valid and that some pans form as a result of interaction between all of the factors. Strong regional structural trends would have controlled faulting and drainage patterns, which in turn controlled pan distribution and formation. As soon as a hollow had formed in which water could accumulate and in turn evaporate, so salinity would increase. The main affect of salinity would be to expose the pan to erosion by curbing vegetation growth and attract animals.

5.4 Conclusions

The geomorphological evolution of southern Africa over the past 150 million years has been influenced by many factors, with some influences inherited from geological events dating back to the Archaean. The influence on the topography of southern Africa during the period following the break-up of Gondwana has been marked, with the Cenozoic Era seeing the formation of the Kalahari Basin, and the evolution of the drainage systems into the form they exhibit at present.

The uplift of the margins of southern Africa in the middle and late Cretaceous had the affect of creating a dual drainage pattern characterised by short, energetic rivers draining down towards the sea and long rivers flowing through the interior of the continent. Although various possibilities exist for the location and even flow directions of rivers at this time, it is fairly evident that either downwarping or uplift in the late Cretaceous and early Cenozoic caused the back-tilting of many rivers into the basin and deposition of Kalahari Group sediments ensued. As rivers were captured and diverted away from the Kalahari Basin and the climate became more arid, so deposition largely ceased, and further uplift along certain axes in the Tertiary exposed some of the Kalahari Group sedimentary rocks to erosion. Seismic activity in the central parts of the Kalahari Basin has been linked to extensions of branches of the East African Rift System into the Kalahari Basin and this has further influenced the drainage patterns and resulted in continued deposition in the Okavango Delta.

CHAPTER 6: PALAEOCLIMATE

6.1 Introduction

The gathering of palaeoclimatic data from the Kalahari has in the past been hampered by the paucity of outcrops and the poor accessibility of much of the area. The dating of the arid and humid phases recognised in the sedimentary record has also been problematic, and the reworking, redeposition and post-depositional modification of Kalahari sediments has meant that it is difficult to obtain accurate and reliable dates for the original deposition of sediments. The dating of arid phases has been especially problematic and in many cases arid conditions have been assumed to be represented by gaps in a sedimentary succession. The advent of luminescence dating techniques in recent years has largely overcome this problem, however, and dating of Quaternary aeolian depositional events is now possible. Although there are still large gaps in our understanding of the climatic conditions during the early and middle Cenozoic, recent information from various localities around the Kalahari Basin has increased our understanding of the palaeoclimatic conditions present during the deposition of the Kalahari Group. Figure 6.1 shows the localities of some of the sites mentioned in this chapter where important palaeoclimatic data for the study area has been gathered.

Palaeoclimatic data for the late Tertiary and Pleistocene climates of the southern Kalahari region has come from spring and tufa deposits developed on the dolomites of the Ghaap Plateau. Depositional cycles preserved in these deposits provide a comprehensive record of changing climatic conditions dating back to the early Pliocene (Butzer *et al.*,1978). Sedimentary, palaeontological, and archeological evidence from nearby Kathu Pan (Beaumont *et al.*,1984) and Wonderwerk Cave (e.g. Avery,1981; Beaumont *et al.*,1984; Butzer, 1984a,b) and a chronology of flow changes in the Molopo River (Cooke, 1975b) has also provided evidence of climatic fluctuations which date back to the early Pleistocene.

A study of the growth rate and composition of speleothems (cave dripstones) has provided important evidence of palaeoenvironmental changes in the Kalahari. Speleothems are well suited to palaeoclimatic interpretations as uranium-series dating can be applied, they are sensitive to

climatic changes, and finally, they are terrestrial deposits, thereby complementing the marine records (Lauritzen and Lundberg, 1999). The dating of speleothems in the caves of the Kalahari (most notably Lobatse and Drotsky's Caves) has enabled several main phases of increased precipitation in the region surrounding the deposits to be identified (e.g. Cooke,1975a,b; Cooke and Verhagen, 1977; Cooke 1984; Shaw and Cooke, 1986).

Further evidence from the central Kalahari has come from analysis of the palaeolakes occurring in the Okavango- Makgadikgadi area, with several old shorelines of a palaeo Lake Makgadikgadi having been detected (e.g. Ebert and Hitchcock, 1978). Dating of palaeo-lake levels, usually with radiocarbon dating of associated inorganic and organic carbonates, (e.g. Street and Grove, 1976; Cooke, 1976, 1979, 1984; Heine, 1978, 1987; Cooke and Verstappen,1984; Shaw and Cooke, 1986) has led to several proposed dates of arid and more humid conditions, but as recognised by other authors (e.g. Cooke, 1980), inflow from rivers that originated on the basin margins, tectonic activity, and the effects of river capture, would have all had a major influence on the palaeo-lake levels. For these reasons very careful interpretations of palaeoclimatic data from Makgadikgadi are required, and lake levels cannot be attributed to only higher or lower precipitation.

The existence of the extensive linear dune fields, now vegetated, covering large portions of the Kalahari Basin has been seen as one of the strongest pieces of evidence for arid conditions having existed in the Kalahari during the late Tertiary and Quaternary. Connecting the dunes to periods of aridity is not without problems however. Dunes do not necessarily represent periods of very high aridity, and although precipitation is the most important variable, dune mobility is also affected by factors such as topography, vegetation cover and wind strength (Lancaster; 1981, 1987,1988,1990; Thomas, 1988c; Thomas and Shaw, 1991b; Wiggs *et al.*, 1996; Bullard *et al.*, 1997), with wind energy being of utmost importance on interdunes and lower dune slopes, and vegetation cover providing the main limiting factor for sediment movement on the upper slopes and crests of dunes (Wiggs *et al.*, 1995). Lancaster (1988, 2000) states that for conditions to be suitable for reactivation of dunes in the southwestern Kalahari, a reduction in rainfall of 50% at present temperatures or a 20% increase in the percentage of time that the wind is above threshold for sand transport would be required. Lancaster (1981) regarded the 150 mm isohyet, shown on figure 4.5 in its present position, as representing the limit of currently active dunes, and believes that when the dunes of the northern dune field were formed, this isohyet must have been some

1000-1200 km northeast of this position. As already discussed in Chapter 4, the availability of sediment for the dune building is an important factor, with more episodes of dune building occurring in areas where abundant deflatable fluvial sediment is available (Thomas *et al.*, 2000). Regional variations in climate should also be considered, and periods of aridity in the southwestern Kalahari are likely to both begin earlier and last longer than in the northern parts of the basin, resulting in a greater likelihood of the record of older dune events in the southwestern Kalahari being erased through subsequent reworking (Stokes *et al.*, 1997a; Thomas *et al.*, 2000). In addition to this, variations in sand activity in dune fields have been shown to occur episodically and in short-lived events of as little as 10 years (Bullard *et al.*, 1997). In general, however, the huge extent of the linear dune fields in the Kalahari suggests formation can only have occurred in regionally more arid periods, and as a result the dune fields do provide useful palaeoenvironmental information. Many of the recent ages retrieved from dunes are presented in section 4.2.5.

Evidence from the lunette dunes on the edges of the pans in the Kalahari should be looked at a little differently, with the outer dunes and the inner dunes (see Chapter 4) having developed under slightly different conditions. The outer sandy dunes, formed as a result of the erosion of unconsolidated Kalahari sands when conditions were fairly dry, vegetation cover minimal and wind strength strong (Lancaster, 1978b). The inner clay-rich dunes formed as a result of erosion of the pan floor itself, which would have initially required enough water in the pans to keep salinity high, and vegetation cover absent. This would suggest slightly wetter conditions than those experienced at the present day, being accompanied by high evaporation rates increasing salinity in the pans (Lancaster, 1978b). The subsequent drying up of the pans in a more arid period would have allowed deflation of the pan surface, and Deacon *et al.* (1984) believe that many of the lunettes were formed during drier conditions in the Late Holocene.

6.2 Evidence for climatic change

Following the warm, humid climate that prevailed over a large portion of Southern Africa for the first half of the Cretaceous (Dingle *et al.*, 1983; Rayner *et al.*, 1991), conditions became increasingly drier over much of the region (Partridge, 1998), and evidence from kimberlite crater-fill sediments at Orapa in Botswana suggests that drier and/or cooler conditions may have started replacing the warmer, wetter conditions by the mid Cretaceous (Rayner *et al.*, 1997; Bamford, 2000). The extensive duricrusts found over much of southern Africa (see Chapter 4), and in particular the formation of the silcretes are also believed to have been indicative of arid conditions in the early Tertiary, with some authors suggesting that a phase of their formation coincided with the early Palaeogene (Palaeocene and Eocene) formation of similar duricrusts at various other localities in the world (e.g. Partridge and Maud, 1989; Tyson and Partridge, 2000). The ages of duricrusts are not easily constrained, however. As described in Chapter 4, the complex relationships between calcretes and silcretes in the Kalahari suggest that various ages of silcretes occur, with the majority appearing to have been formed in the Quaternary. In addition to this, the climatic conditions of silcrete formation are not easily defined, and may have more to do with local conditions than regional climatic conditions.

The early Oligocene was characterised by lower sea levels (Haq *et al.*, 1987), probably caused by the establishment of the Eastern or Greater Antarctic ice sheet (K.C.A. Burke, *pers. comm.*). The late Oligocene and early Miocene were possibly characterised by wetter conditions, influenced by a period of global oceanic warming that occurred during this time (Tyson and Partridge, 2000). Faunal and floral evidence from alluvial deposits in the western parts of South Africa suggest woodland environments during the period between 19-17 Ma, interpreted as being indicative of fairly humid conditions (de Wit, 1990; Bamford, 2000; Vrba, 2000).

The expansion of the East Antarctic ice-sheet around 14 Ma was followed by the start of the cold upwelling within the Benguela Current off the west coast of southern Africa which is thought to have resulted in the termination of this wetter period, and the establishment of the desert conditions along the southwest coast of Africa (Tyson and Partridge, 2000). This together with the uplift along the eastern part of the subcontinent also resulted in a strong climatic gradient from east to west across southern Africa (Tyson and Partridge, 2000). From the mid Miocene, a greater seasonal variability in rainfall seems to have become entrenched (Partridge, 1990), and the onset of arid periods through to the Late Pliocene - Early Pleistocene comes from the examination of microfauna fossils preserved in northern Namibia (Pickford *et al.*,1994). Further evidence of an intensification of arid conditions has been recognised in the Lower Pliocene (Botha *et al.*, 1986, Segalan *et al.*, 2004) and Late Pliocene (Partridge,1993), with a global interval of Pliocene cooling and aridification occurring between 2.8 and 2.6 Ma (Partridge, 1990; Tyson and Partridge, 2000). It was this aridification that possibly resulted in widespread calcretisation and silcretisation of the Kalahari sediments and the development of sand dunes. Some evidence of more humid conditions towards the end of the Pliocene is, however, suggested from the Ghaap Escarpment to the south of the Kalahari Basin (Butzer *et al.*, 1978), from the rejuvenation of drainage systems in the Northern Cape Province of South Africa during this period (de Wit, 1993; de Wit *et al.*, 2000), and also from evidence of the development of grassland and forests in the mid- to late Pliocene hominid sites of Sterkfontein and Makapansgat to the east of the Kalahari Basin in South Africa (Lee-Thorpe and Talma, 2000).

The Pleistocene climate was strongly influenced by glacial/interglacial couplets occurring at higher altitudes, onto which complex patterns were superimposed, particularly in sub-tropical areas (Tyson and Partridge, 2000). Lake bed carbonates from the Namib Desert in Namibia suggest wetter conditions between 240-210 ka (Selby *et al.*, 1979), speleothem deposition took place at Drotsky's Cave at ~ 197 400 ± 41 300 BP (Brook *et al.*, 1990) and a synthesis of data from speleothems from Lobatse I, Drotsky's and Bone Caves (Burney *et al.*, 1994; Brook *et al.*, 1996, 1997, 1998) suggests wetter conditions at 200-186, and 133-131 ka. Oxygen isotope records from the south western Indian Ocean suggest that mildest conditions occurred at around 125 ka, with intermittent cooling occurring subsequently (Tyson and Partridge, 2000). Figure 6.2 is a summary of the wet and dry periods in the middle and southern Kalahari from 200 ka to the present. The greater density of data from the Holocene is shown in more detail in a later figure.

The synthesis of speleothems from Lobatse I, Drotsky's and Bone Caves also gave indications of wetter conditions between 111-103, 93-83, 77-69, 50-43, 38-35, 31-29, 26-21, 19-14, 12.5-11 ka (Burney *et al.*, 1994; Brook *et al.*, 1996, 1997, 1998), with speleothem material from the Otavi Mountainland to the east of Etosha Pan (Brook *et al.*, 1996, 1997, 1998) suggesting comparatively arid conditions in northern Namibia at 130-111, 103-93, 83-77, 69-50, 35-31, 30-27 ka. Warm wet conditions at 50-44 ka were recognised from speleothem formation in the Lobatse II cave in the southern Kalahari Basin (Holmgren *et al.*, 1995), and other radiocarbon and U-series dating from Drotsky's cave indicated wetter conditions at 45-37, and 34-29 ka (Cooke, 1984; Cooke and Verhagen, 1977; Shaw and Cooke, 1986). High lake levels existed near the Tsodilo Hills in northwestern Botswana between 40 and 32 ka, although diatom evidence and dune construction at 35-28 ka suggests more seasonal conditions from around 36 ka, with periods of drier and windier conditions (Thomas *et al.*, 2003). Further evidence of wet conditions around this time comes from

a high lake level at Makgadikgadi between 31 and 24 ka (Heine, 1978), and the two high lake level phases between 32 and 27 ka have been recognised on the banks of the Boteti River before it flows into the Makgadikgadi Basin (Shaw *et al.*, 1997), with further evidence coming from Tsodilo Hills where a lake was again present from 27-22 ka (Thomas *et al.*, 2003), and from Gi where lakes existed in the Dobe valley prior to 31 ka (Helgren and Brooks, 1983).

Further to the south, at Lebatse Pan, organic deposits dated about 29 ka suggest wetter conditions at that time (Holmgren and Shaw, 1997), and wetter conditions are inferred from pans of the southwest Kalahari for the period 35-27 ka (Lancaster, 1989). To the south of the basin work done on the Ghaap Escarpment (Butzer *et al.*, 1978) as well evidence from the nearby Wonderwerk, Equus, and Lobatse caves (Butzer, 1984a; Shaw and Cooke, 1986) suggests wetter, cooler conditions between about 30 and 26 ka with drier conditions becoming more dominant from around 24 ka. Conditions further to the west appear to have been drier throughout this period as is indicated by a period of dune building that lasted in the southwestern Kalahari from 28 to 23 ka (Stokes *et al.*, 1997a; Thomas *et al.*, 1997). The coldest conditions occurred during the Last Glacial Maximum around 20 ka (Shackleton, 1977), and although deposition of tufa occurred along the Ghaap Escarpment between 20.8 and 14 ka (Butzer *et al.*, 1978), and high lake levels occurred at Makgadikgadi at 20 990 BP (Street and Grove, 1976), conditions during the Last Glacial Maximum were generally drier than at present as is suggested by the analysis of a stalagmite from Lobatse II cave, which ceased to grow after 21 ka (Holmgren *et al.*, 1994, 1995). Figure 6.3 is a palaeoclimatic reconstruction of rainfall and temperature at the time of the Last Glacial Maximum at 21 to 18 ka using data gathered by the Palaeoclimates of the Southern Hemisphere (PASH) project of the International Union for Quaternary Research. From this data Partridge (1987) has shown that during the period of the Last Glacial Maximum, distinct spatial palaeoclimatic and vegetation gradients prevailed over South Africa. Rainfall over the Kalahari Basin was as low as 40 % of its present mean, and temperatures increased from south to north. From 18 to 15 ka the record from a high resolution sediment core from the continental slope off Namibia suggests drier conditions over the western parts of southern Africa (Gingele, 1996), although other evidence from the interior suggests more humid conditions. In the northwest of Botswana, a shallow lake is believed to have existed to the west of Tsodilo Hills with archaeological excavations suggesting high lake levels between 22.5 to 12 ka (Robbins *et al.*, 1994), and lake shore construction at 18-12 ka (Thomas *et al.*, 2003). Radiocarbon ages for lacustrine carbonates and diatom evidence

suggest a deep, permanent water body existed in the Tsodilo Hills Basin at around 17.5 ka, with a shallower, less permanent lake at around 15 ka (Brook *et al.*, 1992). Conditions in eastern Namibia were thought to be wet at around 17 ka (Deacon and Lancaster, 1988), with wet conditions indicated by spring deposits in pans near Gobabis that were dated at 17.2 and 11.8 ka (Butzer, 1984b). Increased deposition of speleothems to the northeast in Drotsky's cave from 16.2 to 13 ka (Cooke and Verhagen, 1977; Cooke, 1984) confirms wetter conditions during this time. In the southern Kalahari Basin there is evidence for wet conditions lasting from 17 to 15 ka (Kent and Gribnitz, 1985), with stromatolites at Urwi Pan in western Botswana dated at 16.26-15.61 ka suggesting lacustrine conditions during the same period (Lancaster, 1974, 1979).

Warming is believed to have occurred from about 15 ka (Partridge *et al.*, 1990, Partridge, 1993, 1997; Tyson and Partridge, 2000), and this was accompanied by further evidence of wet conditions in the Kalahari Basin. Increased flow in the Molopo River occurred from 16-13 ka (Heine, 1982), shell deposits in the Okwa and Xaudum valleys of central and northwestern Botswana suggest an increase in precipitation from 15 to 12 ka (Shaw *et al.*, 1992), and a high water stage at Makgadikgadi (Shaw, 1988) and at the Chobe-Zambezi confluence (Lake Caprivi)(Shaw and Thomas, 1988) also occurred during this period. Increased runoff in the Cunene River occurred between 15 and 10 ka (Gingele, 1996), and recharge of the aquifer at Stampriet in Namibia is thought to have occurred between 14 and 8 ka (Heaton *et al.*, 1983).

Evidence for drier periods in northern Namibia between 13 and 10 ka is also present, as well as from 10.6-8.5 and at 7.5 with fairly short lived wet periods at 12.5-11 and 8.2-7.9 ka (Brook *et al.*, 1996). The southwestern Kalahari may have been experiencing drier conditions at the end of the Pleistocene, with dune formation indicating that arid conditions prevailed in the southwestern Kalahari between 17 and 10 ka (Chapter 4; Stokes *et al.*, 1997a), possibly extending to about 8 ka on the basis of pollen data from South Africa (Scott, 1989). This may have been when the lunette dunes flanking the pans in the southern Kalahari were formed (Lancaster, 1978b). Figure 6.4 summarises the geomorphic and sedimentary evidence for environmental changes in the Kalahari for the period 50-10 ka.

Particular high temperatures over southern Africa occurred between 7 and 4.5 ka (during the Holocene altithermal), during which time temperatures were up to 2°C higher than at present

(Partridge *et al.*, 1990; Partridge, 1997), and rainfall over the Kalahari during this period was possibly also higher than at present (Figure 6.5). Increased runoff in the Kunene River between 6 and 5 ka was caused by high humidity in the source areas at this time (Gingele, 1996), and wetter conditions were also recognised in Drotsky's cave between 6 and 5 ka (Cooke, 1984), and again at 4 350 BP (Cooke, 1975a). Other evidence of more humid conditions during the Holocene altithermal comes from tufas on the Ghaap Plateau (Butzer *et al.*, 1978), from deposits containing micro mammals in Wonderwerk Cave (Avery, 1981, 1988; Thackeray, 1984), and from higher than average water levels in the Okavango Delta (Nash *et al.*, 1997).

The period following the Holocene altithermal was punctuated by alternating periods of warm, wet and cooler drier conditions, with detailed analysis of a stalagmite from a cave in the Makapansgat Valley southwest of the South African town of Polokwane revealing oscillations in the ^{18}O and ^{13}C records for the past 6600 years (Holmgren *et al.*, 1999; Lee-Thorp and Talma, 2000; Tyson and Partridge, 2000). In general decreasing values for ^{18}O and ^{13}C represent cooler drier conditions, and increasing values represent warmer, wetter conditions, and the values recorded show that the most rapid changes in the climate occurred at around 6 ka and between 4 and 3.7 ka (Tyson and Partridge, 2000). The period of 5 to 3 ka has been recognised by various researchers as an arid period favouring dune formation in the southern Kalahari with a peak of aridity around 4 ka (e.g. Heine, 1982; Van Zinderen Bakker, 1980, 1982, Beaumont *et al.*, 1984; Deacon and Lancaster, 1988), and as already discussed in Chapter 4, luminescence dating in the southwestern Kalahari shows periods of dune formation at 6 and between 2 and 1 ka (Thomas *et al.*, 1997), although wetter conditions over much of the region between 1.8 and 0.5 ka were recognised in speleothem material (Brook *et al.*, 1996, 1998). Warmer, wetter conditions did prevail from around 500 to 600 AD and 900 to 1300 AD, although drier periods may have occurred during the latter time frame, with dune formation in the Kalahari (Thomas, 1997; Thomas *et al.*, 1997). This period was followed by the cooler, drier conditions of the Little Ice Age which lasted from 1300 to 1810 AD. The lowest mean annual temperature was reached at 1700 AD during the last phase of the Little Ice Age (Tyson and Partridge, 2000). The high density of data from the Holocene summarised in Figure 6.6 shows that numerous climatic fluctuations occurred during the last 10 000 years, with at times remarkably different conditions occurring in various parts of the basin at any one time.

6.3 Conclusions

Palaeoclimatic evidence is a useful source of information when attempting to determine the depositional history and setting of Kalahari Group sediments, but climatic data can be misleading and may not necessarily be a good reflection of climatic conditions in the area of sedimentary deposition. The presence of fluvial sediments or even palaeo-valleys (see the Mekgacha mentioned in Chapter 5) may be an indication of humid conditions in areas adjacent to the basin and not, as has often been presumed, of climatic conditions within the Kalahari Basin. Climatic evidence from the Kalahari Basin for the early- and mid-Cenozoic is sparse and is open to much interpretation and although the relatively large amount of climatic data obtained from the Quaternary sediments has enabled many interpretations to be made, the climatic interpretations for this period are characterised by numerous contradictions and generalisations. There has been palaeo-climatic data forthcoming that enables some conclusions to be drawn, but all factors, including tectonic activity, climatic changes and local conditions need to be looked at together.

A number of arid and humid phases have been identified in the Kalahari during the course of the Cenozoic, but the conflicts between different research findings suggests that precipitation gradients from east to west and north to south appear to have had a large influence on local climatic conditions, certainly during the Holocene and possibly even earlier. It is therefore necessary when looking at climatic evidence from an area as vast as the Kalahari Basin to recognise that climatic conditions at any one time may differ substantially across the region, much as they do today, and lacustrine deposits may have accumulated in interdune areas during periods of dune stability (e.g. Lancaster, 1990). Anomalous situations caused by exotic rivers originating in areas of high rainfall influencing the local environment in more arid areas downstream further complicates interpretations of data. The swamps of the Okavango delta, occurring in an otherwise semi-arid area is a good example of this, with local conditions experienced there today not providing a true indication of climatic conditions of the region.

CHAPTER 7: MINERAL POTENTIAL OF THE KALAHARI BASIN

7.1 Introduction

Although an exhaustive inventory of mineral resources of the Kalahari Basin is not within the scope of this study, a mention of some of the known mineral deposits of the area gives some indication of the potential of the area for further mineral exploration and exploitation.

7.2 Archaean

7.2.1 Gold and Nickel

In the Archaean rocks of the Kaapvaal Craton, lode gold mineralisation has been found in the banded iron formation and schists of the Kraaipan Group (e.g. Hammond *et al.*, 1999; Kiefer, 2002; Kiefer and Viljoen , 2004; Hirner *et al.*, 2004) as well as in the Kanye Group and in quartz veins in the Gaborone Granite (Baldock *et al.*, 1977). Although Witwatersrand Supergroup rocks are not shown on the area covered by the sub-Kalahari map, exploration in Botswana has attempted to find the northwestern extent of the Witwatersrand Basin as the deposit contains more than half of the world's known gold resources. Gold is also found in the lower parts of the Ventersdorp Supergroup above the contact with the Witwatersrand Supergroup rocks.

The Phikwe Complex and the Baines Drift complex located in the Limpopo belt in Botswana contain nickel-copper sulphides (Baldock *et al.*, 1977), with a Ni-Cu mine at Selebe-Phikwe having remaining reserves of over 70 million tons (Mt) in 1987 (Department of Mines Annual Report, 1987 cited in: Carney *et al.*, 1994). Other mineral occurrences associated with the greenstone belts are antimony, zinc, iron, asbestos, talc, mercury, magnesite, barite and gemstones (CGS and CGMW, 1999).

7.3 Proterozoic

7.3.1 Iron and Manganese

The Transvaal Supergroup rocks host some of the largest deposits of iron and manganese in the world, with a large iron ore mine near the town of Kathu in the Northern Cape (Sishen Iron Ore Mine) boasting a potential deposit of about 1000 Mt of ore in its high-grade (65% iron) hematite deposit (Van Schalkwyk and Beukes, 1986) and another large deposit further south (Postmasburg) containing between 100 and 1000 Mt of iron. The Kraaipan rocks also contain large amounts of iron, with the Kraaipan occurrence to the west of Amalia containing 10-100 Mt of iron (CGS and CGMW, 1999). Iron has also been mined in southwest Angola from the banded hematite quartzites of the Cassinga Supergroup as well as from a "pebble" deposit thought to represent mid-Tertiary erosion and redeposition of Cassinga Supergroup rocks (Hood and Korpershoek, 1968).

South Africa has huge deposits of manganese constituting more than 20% of the worlds resources (Grohmann, 1995). The Kalahari Manganese Field, to the northwest of Kuruman is one of the biggest occurrences, with the Hotazel Formation (Voelwater Subgroup) containing over 13600 Mt of ore with a manganese content of over 20% (Taljaardt, 1982 cited in: Beukes, 1986). Almost the entire Kalahari Manganese Field is covered by Kalahari Group sediments, with the only outcrop being a small occurrence of Hotazel Formation at Black Rock. Other manganese mines are situated between Sishen and Postmasburg in the Postmasburg Manganese Field and are related to the unconformity between the Campbell Rand Subgroup and the Gamagara Formation in the Maremane Dome (Grobbelaar and Beukes, 1986).

7.3.2 Asbestos

Crocidolite Asbestos was extensively mined in the Northern Cape Province with four main fields having been identified. The Kuruman Crocidolite field which stretches from Severn, 115km northwest of Kuruman, southwards to near Danielskuil. In this field, large deposits situated in the Kuruman Formation in the Pomfret area, where eleven deposits are known in the vicinity of the Pomfret Mine with resources of about 2.3 Mt (Beukes and Dreyer, 1986). The Danielskuil field occurs from about 28°S down to Griquatown, the Postmasburg field stretches between

Postmasburg and Kathu, and the Koegas field, where 22 reefs have been recognised, stretches from Griquatown to Koegas (Ehlers and Vorster, 1998). In a continuation of the same band of occurrences, a deposit of crocidolite occurs at Knapdaar, near the Molopo River in Botswana (CGS and CGMW, 1999). The decrease in demand for asbestos has resulted in the closure of most of the mines, with the last remaining mine operating near Kuruman closing down March 1997 (Ehlers and Wilson, 2001).

7.3.3 Intrusive complexes

The Molopo Farms Complex is a layered Complex of about 12000 km^2 in area. It is a similar age to the Bushveld Complex which is well known for its massive ore reserves, and as a result has been the subject of several exploration programs. Nickeliferous pyrrhotite, chalcopyrite, and pyrite have all been reported from the Molopo Farms Complex (Gould *et al.*, 1987).

The Cunene Complex of Southern Angola and northern Namibia has occurrences of Fe and Ti, with smaller amounts of vanadium and platinum. The Gambos occurrence near the town of Chiange in Angola has an estimated 10 Mt of Fe and Ti (CGS and CGMW, 1999).

Magnetic anomalies associated with the Kalahari Suture Zone have been identified as possible exploration targets, with the Tshane complex, Tsetseng Complex to its east, and an anomaly to the south of the Tshane complex, "the Great Red Spot" having been preliminarily investigated for possible Olympic Dam-type Cu-Co or Ni-Cu-Co-PGM mineralisation (Brett *et al.*, 2000). An anomalous feature extending for 150 km from 25° S to 26°15' S was also targeted for exploration, and initial drilling into the so-called Mabua Breccia Trend (Brett *et al.*, 2000) intersected volcanic breccia (Meixner and Peart, 1984). Follow up drilling intersected altered dioritic hydrothermal breccias, and the Mabua Breccia Trend is considered to hold potential for magnetite / sulphide-associated Cu-Au mineralisation of Olympic-Dam type (Brett *et al.*, 2000).

7.3.4 Copper, Lead, Zinc

Copper mineralisation occurs in the Duruchaus Formation (Nosib Group) of Namibia, and in the Ghanzi Group of Botswana. There are numerous occurrences of strata-bound Cu-Ag and minor associated Pb and Zn sulphide mineralisation (Carney *et al.*, 1994; Modie, 2000), with one of the largest deposits (Zeta) having an inferred 20 Mt of ore, consisting of copper, silver and minor gold, zinc, and molybdenum (Carney *et al.*, 1994). In the Ngwako Pan area two localities containing fairly high grades of copper sulphides had their reserves calculated at 27 Mt at 1.0% Cu grade and 0.2% cut off value, and 17 Mt at 1.5% Cu grade and 0.5% Cu cut off value for the one area, and 95 Mt at 0.8% Cu grade and 0.2% cut off value and 49 Mt at 1.4% Cu grade and 0.5% cut off value for the second area (Siamisang, 1996). To the north and northwest of this, massive amounts of chalcopyrite have been described in dolomites southwest of the Koanakha Hills (Wright, 1958, in Carney *et al.*, 1994), and in the Roibok gneisses, the presence of pyrite enrichments and malachite and chrysocolla staining was reported (Lüdtke *et al.*, 1986). It is believed that these metamorphic rocks may have potential for stratiform, VMS deposits (Carney *et al.*, 1994). In the Aha Hills, high grade deposits of lead, zinc and silver, hosted in sandstones and dolomites have been proved by prospecting (Stalker, 1983, in Carney *et al.*, 1994). Volcanogenic massive cupreous pyrite deposits are associated with the Matchless Amphibolite Member of the Kuiseb Formation (e.g. Otjihase, Matchless) (Adamson and Teichmann, 1986; Miller, 1992c; Killick, 2000). Cu, Pb, and Zn are found near Tsumeb, Namibia, in the Tsumeb Subgroup of the Otavi Group, with a deposit of over 10 Mt (CGS and CGMW, 1999). The Tsumeb orebody is regarded as having developed in karsts at around 530-580Ma (Killick, 1986), and the whole of the Northern Platform and adjoining marginal areas of the northern and north-east trending branches of the Damara Orogen are considered target areas for Tsumeb-type and Mississippi Valley-type karst-related mineralisation (Miller, 1992c).

To the east of Kuruman at Peiring, a Zn and Pb deposit with a resource of 18 Mt of 3.6% Zn and 0.6 % Pb was found in the stromatolitic dolomites of the Campbell Rand Subgroup (Wheatley *et al.*, 1986), and mined until 2001. Another Zn-Pb deposit occurs at Bushy Park, about 34 km north-northeast of Griquastad, and although this deposit is yet to be developed, it was thought by Ehlers and Wilson (2001) to have good potential.

The Lufilian Belt is home to more than 800 mines and prospects, the majority of which are found in Zambia, with the western extent of the belt in Angola largely unknown and unexplored

(Premoli, 1999). The Lufilian Belt is characterised by three main types of mineralisation, namely stratiform, vein and skarn (Unrug, 1988). The stratiform deposits are the most impressive with major deposits as well as numerous minor occurrences of copper, copper-cobalt and uranium occurring in the Macondo Group of Angola and the Katangan Supergroup sediments of western Zambia (Unrug, 1988).

7.4 Palaeozoic

7.4.1 Coal

Coal deposits are fairly well known in the area covered by Kalahari Group sediments. In Namibia six areas with coal potential have been identified: Kaokoland; Huab Basin; Kavango and Caprivi; Owambo Basin; Waterberg Basin; Aranos Basin, with the latter four all covered by Kalahari Group sediments (Hegenberger, 1992). In southeastern Namibia, the coal in the Aranos Basin occurs in the Prince Albert and Whitehill Formations, with the individual seams in the latter being less than tens of centimetres in thickness (Cairncross, 2001). Total in situ tonnage of coal in the Aranos Basin is estimated at 371.9 Mt (Marsh and McDaid, 1986). Coal also occurs in the Owambo Basin of northern Namibia, where it is confined to the eastern section of a downthrown graben to the southeast of Ondangwa (Cairncross, 2001). In central Angola, the Lungue-Bungo deposit has an estimated reserve of up to 50 Mt (CGS and CGMW, 1999). In Botswana twelve prospected coal fields have been demarcated (Clark *et al.*, 1986; Chatupa, 1991). These areas stretch in a rough arc from the Namibian border, across to eastern Botswana and up to the east of the Makgadikgadi Pans to an area across the border from the Zimbabwean town of Hwange. In Botswana the best coal reserves are found in the southeast of the country near the town of Palapye, in the Moijabana/Morupule areas where in situ reserves of 9210 Mt of coal are present (Clark *et al.*, 1986). Some 5500 Mt of coal of inferior quality has also been discovered further to the southwest, in the Letlhakeng and Dutlwe areas, and to the east of Orapa, at Dukwe there is an estimated 50-500 Mt of coal (Clark *et al.*, 1986). In western Zimbabwe, in the Wankie (Hwange) coal field, in the vicinity of the town of Hwange, there is a deposit of 2100 Mt of mineable coal (Duguid, 1986).

7.5 Mesozoic

7.5.1 Diamonds

One of the greatest opportunities for exploration in the Kalahari Basin remains that for diamonds. In 1996 Botswana was the largest diamond producer in SADC, with 17,71 million carats produced (15,7% of the worlds production) (Cole, 1998). Most of these diamonds came from three kimberlite pipes, although 56 diamondiferous kimberlites are listed for Botswana (Cole, 1998). Largest of Botswana's diamondiferous pipes is the Orapa pipe, which is the third largest diamondiferous kimberlite pipe in the world, covering an enormous 110.6 ha (Carney *et al.*, 1994; Cole, 1998). Jwaneng is one of the richest kimberlite pipes with a grade of 150 ct/100t (Jennings, 1995). Both the Orapa and the Jwaneng pipes overly the cratonic areas, but many very large non-diamondiferous kimberlite pipes are also found in the southwest of Botswana, the most sizeable being the 200 ha, 77 million year old kimberlite pipe in the Tshabong field (Carney *et al.*,1994; Key and Ayres, 2000). Other significant kimberlite fields in Botswana include the Gope-Kikao, Lekgodu, Kukong, and Mabuasehube, the latter three of which are all situated in the southwestern part of Botswana. In Angola there are 105 known diamondiferous kimberlites (Cole, 1998), with one of the largest kimberlite pipes in the world, the Camfuca-Camazambo pipe of ± 150 ha (Khar'kiv *et al.*, 1992), occurring on the Chicapa River to the west of the town of Lucapa. Further south along the same river, the Catoca pipe contains a significant amount of diamonds (CGS and CGMW, 1999) and numerous other small pipes are also found in the area, as well as smaller alluvial deposits along the Chicapa and other rivers to the east and northeast as well as to the west along the Cuango River (CGS and CGMW, 1999). In South Africa the Finsch Mine lies to the south of Kuruman, and up until 1995 had produced approximately 93 million carats (Lynn, 1998). To the southwest of this a smaller kimberlite was mined at Peiserton Mine with some similar sized pipes at Sandrift to the northeast of Prieska (CGS and CGMW, 1999) and Makganyene 25 km northwest of Postmasburg.

An understanding of the pre-Kalahari structure of the region is of importance to kimberlite exploration, with kimberlite emplacement thought to be structurally controlled (White *et al.*, 1995). Cretaceous kimberlite pipes in Angola are distributed in a northeast-trending belt (Jelsma *et al.*, 2004) and are concentrated at the intersection of north-northwest-, east-northeast- and east-southeast-trending faults, and kimberlite dykes which are also controlled by the main structural grain of the area (Cole, 1998). In Botswana, the pipes at Orapa appear to be related to northwest-trending structures, and the Jwaneng kimberlite occurs near an intersection between northwest- and northeast-trending faults (Cole, 1998). In South Africa most kimberlite pipes appear to be related to the intersections of northeast- and northwest-trending structures (Friese, 1998).

Some diamondiferous kimberlites directly underlie younger sedimentary rocks. In Angola most of the diamondiferous pipes discovered are covered by either Calonda and Kwango Formation rocks, or by Kalahari Group lithologies (Cole, 1998) and in Botswana only one small diamondiferous kimberlite (Martin's Drift) crops out (Cole,1998) with the remainder being covered by Kalahari Group sedimentary rocks. In the Jwaneng field of southern Botswana, the 2424DK1 and 2424DK2 pipes were covered by 30m and 40-45m of Kalahari Group sedimentary rocks respectively (Carney *et al.*, 1994). In Botswana where Karoo Supergroup basalts underlie the majority of the Kalahari Group rocks, the effectiveness of aeromagnetic techniques for kimberlite exploration is limited. The exploration for kimberlite pipes, which may be buried beneath tens of metres of Kalahari Group sedimentary rocks, is thus heavily reliant on the search for indicator minerals as well as an understanding of sedimentary depositional processes that may have affected the eroded products of the kimberlite pipes. Orapa was discovered with the aid of indicator minerals in river beds which had been significantly affected by tectonic uplift in the Tertiary (Chadwick, 1983, in Cole, 1998). An understanding of how uplift may alter the courses and flow direction of rivers which may have eroded the kimberlites targeted is therefore of great importance.

Many of the southern African kimberlites, particularly those that may have been uplifted, have been eroded, with the diamonds being deposited along the west coast of southern Africa as well as inland along river systems. According to de Wit (1996) inland alluvial deposits in southern Africa had produced some 18 million carats up until 1996 and this is therefore seen as an important resource. Diamondiferous gravels are well known in South Africa from the Schweizer-Reneke and Lichtenburg districts to the east of Vryburg and diamondiferous gravels have been excavated about 64 km east of Kuruman, at Mahura Muthla, where diamondiferous gravels in palaeo-channels of up to 40m thick are sporadically mined (Ehlers and Wilson, 2001) and some 3 500 carats have been recovered (Ward *et al.*, 2004). The diamonds have both a primary origin, weathering from the Cretaceous kimberlites, as well as a secondary origin weathering from older sedimentary rocks. It has been suggested that diamonds which were originally eroded from older (pre-Karoo Supergroup) kimberlites may have been deposited along with Dwyka Group sedimentary rocks and that the subsequent erosion of the Dwyka rocks released many of the diamonds (du Toit, 1951; Stratten, 1979; Marshall, 1986; Van Wyk and Pienaar, 1986; Moore and Moore, 2004). In northeastern Angola alluvial diamondiferous placers have been found at several localities along the rivers flowing towards the Democratic Republic of Congo (Cole, 1998). Many of these alluvial deposits originate from weathering of the Cretaceous Calonda and Kwango Formations, which themselves contain diamondiferous palaeoplacers (Cole, 1998; Giresse, in press).

7.6 Cenozoic

7.6.1 Tertiary diatomites, clays and evaporites

Diatomite or Kieselguhr is primarily used as a filter aid, but is also used as a filler in plastics, paper and rubber, thermal insulator, carrier for catalysts and insecticides, anticaking agent in fertilisers and explosives, a pozzolanic admixture to cement, a mild abrasive, and as a source of reactive silica for the manufacture of sodium and calcium silicates (Strydom, 1998). In the Postmasburg and Kuruman districts of South Africa, numerous potentially economically exploitable reserves have been identified with in situ reserves of individual deposits vary between 20000 and 80000 tons, and a total volume of over 500 000 tons (Strydom, 1998). One of the largest of these deposits, Witberg, occurs about 65km west of Hotazel, and has an estimated in situ reserve of 150 000-170 000 tons (Oosterhuis *et al.*, 1991). A diatomite deposit of 60m wide and ~3km long is reported from the bed of the Klein Nossob River in Namibia where it has been exploited as a building stone (Schneider and Genis, 1992a). Figure 7.1 shows the distribution of 79 exploitable kieselguhr deposits in South Africa.

In Namibia, in a pan approximately 100 km southeast of Gobabis on the farm Nui-Sei 376, a deposit of authigenic sepiolite, a clay with super-absorbent properties, occurs in the form of

scattered porous aggregates of 5-30 cm in diameter, and as veins in a surface limestone (Schneider and Seeger, 1992). An estimated reserve of 4 Mt of pure and hard sepiolite, and 5 Mt of soft sepiolite occurs on the property (Schneider and Seeger, 1992), and additional deposits may occur in some of the numerous pans of the Kalahari. A chemical analysis of the sepiolite from Nui-Sei is shown in Table 7.1.

Table 7.1 - Analysis of sepiolite from Nui-Sei 376 (Levin, 1966).

	Weight %
SiO_2	55.6
Fe_2O_3	1.2
TiO_2	0.16
Al_2O_3	2.2
CaO	5.8
MgO	15.8
Na_2O_3	1.7
K_2O	1.7
CO_2	5
LOI	11.9

Salt is mined at some localities, for example in a band of pans stretching north northwest of Upington towards the Botswana border (Oosterhuis, 1998a), with Norokei and Groot Witpan Pans producing 60 000- 70000 tons p.a in 1981 (M.A. Thomas, 1981). At Sowa Pan in the Makgadikgadi Basin a reserve of more than 1000 Mt is mined (CGS and CGMW, 1999) along with soda ash, salt cake and potash (Gould, 1986). Soda is found in a deposit of up to 1 000 000 tons at Otjivalunda Pan near Etosha, and soda nitre ($NaNO_3$) is found in southeastern Namibia along the courses of the Auob, Olifants and Nossob Rivers where it occurs in calcrete, calcareous conglomerate and grit of the Kalahari Group. It is not commercially exploited (Schneider and Genis, 1992b). A gypsum deposit of between 5 and 100 Mt is mined at Fincham to the east-southeast of Upington in vleis and pans overlying the Nama Group (Oosterhuis, 1998b).

7.6.3 Heavy mineral deposits

Continental rifts are believed to be favourable sites for the accumulation of heavy minerals such as magnetite and ilmenite (Reid and Frostick, 1985). The main requirements for the concentration of the minerals in an economically viable deposit are an easily erodible source area and a method of concentration of the mineral grains (Frostick and Reid, 1990). One of the most effective methods of concentration is by the wave action along shorelines over a prolonged period and this has proved to be effective at Lake Turkana in northern Kenya (Frostick and Reid, 1990).

7.7 Groundwater

Water is a scarce and therefore very valuable commodity in the semi-arid to arid Kalahari Basin, and in much of the central and southern Kalahari, groundwater is the only source of permanent water. Proposals were made in the past to divert water into the Kalahari, the most famous being that of Schwarz, who in 1920 proposed a plan for irrigating the Kalahari with water diverted from higher rainfall areas in Angola and Zambia, hoping in the process to change the climate in the interior of the continent (Schwarz, 1920; Fig. 7.2). More recently, plans have been outlined for piping water from the Zambezi River to Gaborone (du Plessis and Rowntree, 2003), for diverting water from the Okavango river to Windhoek and for exploiting water from the Okavango Delta. The massive costs and potentially destructive environmental impact of these schemes is, however, likely to prevent further action and groundwater remains the most important source of water for the region.

In much of the area the Karoo Supergroup rocks are the main aquifers, with the Ntane sandstone of the Lebung Group and the Ecca Group sedimentary rocks providing much of the groundwater. In other regions the Kalahari lithologies themselves form the aquifers. It was found that in the western Hereroland region of Namibia the sandstones of the Kalahari Group, which equate with the Eiseb or Eden sandstones, form the best aquifer, the Middle Kalahari aquifer, (de Beer and Blume, 1985), while the Lower Kalahari , which includes the Budin and Wessels Formations was found to commonly contain brackish to saline water (de Beer and Blume, 1985). Further to the east in the Gam area Namibia Kalahari Group aquifers are an important source of water particularly

where faulting has lowered the Kalahari Group rocks below the water table (Simmonds and Smalley, 2000). In the southern parts of the southern Kalahari Basin, water from the Kalahari Group is of better quality than that from older rocks in the area, and only deteriorates when mixed with saline water from the Dwyka Group (Levin, 1980). The Wessels Formation gravels found in old palaeochannels can provide a good source of water with yields of up to $15m^3$/hr having been recorded (Molwalefhe, 1995).

Structural features associated with faulting are important targets for groundwater exploration and groundwater in the Dwyka Group rocks is easily found, as it occurs along horizontal and vertical structures (Levin, 1980). Structures are not, however, easily visible on the surface because of cover of Kalahari unconsolidated sands. Landsat and aerial photographs are therefore of limited use, but aeromagnetic data combined with the satellite imagery has been used successfully in detecting fault-related lineaments in sand-covered areas (e.g. Zeil *et al.*, 1991), as has electrical resistivity combined with magnetics (Peart, 1979) and gravity (Reeves and Hutchins, 1982). The depth of the groundwater below the surface is influenced by the thickness of Kalahari Group sediments, with shallow water tables occurring along watersheds and where the cover of Kalahari Group sediments is thin, and deep water levels occurring in areas where the Kalahari Group sediments are thickest (Levin, 1980, 1981).

Recharge of the groundwater is low because of low rainfall and high evapotranspiration, and subsiding water tables were described as far back as the 1950's (Wayland, 1953). Discharge of groundwater occurs from some saltpans by capillary action and evaporation of water (Levin, 1981). Boreholes are used to provide water to livestock, and in historical times, borehole water levels and yields have dropped through usage, with complete drying-up of boreholes occurring in some areas of the northern Cape Province during dry periods (Levin, 1980).

The unconsolidated sands at the top of the Kalahari Group are believed to impede rainfall infiltration, and according to Boocock and van Straten (1962), recharge of aquifers below thick deposits of sand is unlikely. De Vries (1984) believes that the last period of active recharge of Kalahari aquifers occurred at the end of a wet period about 12 500 years ago, but isotope observations in the Gordonia district of the Northern Cape Province have found that diffuse rainfall recharge can still occur over a large area (Verhagen, 1985). Aerial and lateral groundwater recharge of Karoo Supergroup aquifers has been shown to occur in southwestern Botswana (Molwalefhe, 2003) and some recharge is known to occur along the Kuruman River during floods (Levin, 1980, 1981; Meyer *et al.*, 1985; Verhagen, 1985).

7.8 Construction materials

Calcretes form an important source of aggregate for roads in the Kalahari, with compacted unconsolidated sands also being used on minor roads (Netterberg, 1998). In the Lobatse-Kanye area of Botswana, building aggregate is quarried and crushed from quartz porphyry and dolerite, and building stone comes from the ironstones and quartzites of the Transvaal Supergroup rocks in the south of the basin, and Karoo Supergroup sandstones and basalts and Quaternary silcretes around Maun in northern Botswana (Kreimeyer *et al.*, 1990). River sands are exploited from various localities in the Kalahari, although large deposits of these sands are fairly scarce, and largely limited to the eastern parts of Botswana, in particular around Francistown and Selebi-Phikwe (Kreimeyer *et al.*, 1990).

7.9 Conclusions

D.A. Pretorius (1979, p 414) once described the Kalahari "the last frontier for grassroots mineral exploration in the sub-continent". Despite the fact that a large amount of mineral exploration has been undertaken in the region in the last 25 years and many economically viable deposits have already having been discovered, the Kalahari Basin still has a large potential for the discovery of exploitable mineral deposits. The improvement in geophysical techniques, coverage and availability of data, as well as improved satellite imagery, better spread of geochemical surveys, and regional mapping programs has aided target generation. Geobotany has been found to be a useful tool, with the species *Helichrysum leptolepis* being used as an indicator of copper mineralisation in the Damara belt (Cole and Le Roex, 1978). The isopach, geological and topographical maps produced during this investigation have already been used as important exploration tools by various private companies, and will aid in the identification and exploitation of new mineral reserves. A better understanding of the geomorphic evolution of the area is also vital to exploration as geochemical sampling of stream sediments must take into account the timing of movement along tectonic axes.

CHAPTER 8: DISCUSSION AND CONCLUSIONS

8.1 Introduction

The Kalahari Basin is in many ways a unique area, with rocks deposited and emplaced over the past 3 500 million years exposed both within and around the edge of the basin. Several major tectonic events have occurred in the region, with each new event often exploiting older structural orientations and crustal weaknesses and in the same way, the subsidence, uplift and faulting that formed and shaped the Kalahari Basin and controlled the deposition of the Kalahari Group sedimentary rocks in the Late Cretaceous and Cenozoic was strongly influenced by basement structures and lithologies. In order to better understand the formation of the Kalahari Basin and the controls over subsequent sedimentary deposition, the influence of preceding events must therefore be recognised and understood. The main geological events shaping southern Africa from the Archaean to the present day have collectively defined the nature of the Kalahari Basin.

The dominant trends influencing the development of basins and orogenic belts since the Palaeoproterozoic have been oriented in approximately NE-SW and NW-SE directions. In many cases, the NE orientation represents the orientation of the rifts that formed, with the NW orientation often representing the faulting perpendicular to the rift orientation. There are exceptions to this, however, with NW-trending arms extending from apparent triple junctions having developed at various stages. The same NW- and NE-trending structures appear to have been reactivated at various times and during the Phanerozoic were important in controlling initially the Karoo and later the Kalahari sedimentation.

8.2 The influence of the pre-Kalahari geology on Kalahari basin development

The position of the cratons has been an important factor influencing the distribution of rifting in southern Africa, as generally, propagating rifts tend to avoid going through the cratons. The Congo, Kaapvaal and Zimbabwe Cratons form dominating stable, and these cratons are separated by various tectonic belts which record a history of break-up, accretion and collision. The Kaapvaal and Zimbabwe Cratons are separated by the Limpopo Belt and in turn are separated from the Congo Craton by the Damara, Irumide, Zambezi and Lufilian Belts. Not all of the boundaries of the cratons are clearly defined and the western flank of the Kaapvaal Craton is possibly marked by the Kalahari Line which joins with the NE-trending Makgadikgadi Line to form the Kalahari Suture Zone.

The main Palaeo- and Mesoproterozoic belts underlying the Kalahari Group sedimentary rocks are largely oriented in either northeast or northwest directions. The Magondi, Irumide, and Kibaran belts are all oriented northeast-southwest and the Namaqua Belt has a NW-SE orientation. The Koras-Sinclair-Ghanzi Rift developed along NW-SE and NE-SW arms, in the late Mesoproterozoic with the NE branch believed to represent the failed arm of a triple junction (Borg, 1988). During the Neoproterozoic break-up of Rodinia the same dominant trends were once again prominent, with the Damara rifting and orogeny occurring along both north-, south- and northeast-trending arms. The PanAfrican suturing which resulted in the final assembly of Gondwana, joined together the Congo and Kalahari Cratons in a suture zone marked by the Zambezi, Damara and Lufilian Belts. The Zambezi Belt is separated from the Lufilian Belt on its northern side by the northeast-southwest trending Mwembeshi Suture Zone which can be correlated to the southwest with the Okahandja Lineament Zone of the Damara Belt.

The assembly of Pangaea occurred firstly with the formation of the continent of Laurussia between 390 and 320 Ma and then with the collision of Laurussia with Gondwana (Burke and Dewey, 2002). The collisions are believed to have caused widespread rifting and strikeslip movement across Gondwana (Burke and Dewey, 2002) and during the Carboniferous-Permian the Botswana-Zambezi Basin formed (Visser and Praekelt, 1995), once again following a northeast orientation and involving some reactivation of the structures of the northeastern branch of the Koras-Sinclair-Ghanzi Rift (Borg, 1988). While the Botswana Basin possibly developed predominantly as an intracratonic sag basin as Johnson et al. (1996) believe, it is significant that sedimentation was influenced by NE-SW and NW-SE faults (R.A. Smith, 1984). During the Late Permian-Triassic the Cape Fold Belt was formed, with the collisional event possibly resulting in the formation of the Southern Trans-African Shear System (STASS) (de Wit *et al.*, 1995) which developed in the Damara, once again following a NE-SW orientation. Faulting, regional uplift and down-warping controlled Karoo sedimentation in the Late Permian and Triassic with reactivation of older faults occurring. In southern Botswana Karoo deposition was influenced by reactivation of faults bounding older grabens containing large thicknesses of Waterberg rocks and believed to have been active since the Palaeoproterozoic (Green *et al.*, 1980). Early Jurassic down-faulting of the mid-Zambezi Basin is thought to have followed older structures (McConnell, 1972; Lambiase, 1989) and the Luangwa rift possibly developed over a suture in the Irumide mountain belt (K.C.A. Burke, *pers. comm.*).

At around 180 Ma flood basalts were extruded over much of southern Africa and dolerite sills and dykes were intruded. The massive Botswana dyke swarm was emplaced along a NW-SE trend possibly related to a failed Jurassic rift, although the presence of older Proterozoic dykes with the same orientation within the dyke swarm suggests that once again older structures were exploited. The occurrence of both the Makgadikgadi and Okavango basins along the axis of the dyke swarm may be linked to later sag caused by the weight of the emplaced dykes in the crust. Derito *et al.* (1983) showed that dense loads in the crust, like basaltic dykes, can remain isostatically uncompensated until such time as any stress is applied to the lithosphere. Once stress is applied, subsidence along the dyke swarm will follow. Both the Okavango and Makgadikgadi occur at the intersection of the dyke swarm with NE-SW trending faults related to post-Karoo faulting as well as to Cenozoic rifts extending from the EARS. It is possible that this intersection resulted in subsidence along the dyke swarm.

The NE- and NW-trending faults were once again reactivated when the separation of Madagascar and the Seychelles from Africa at around 150-112 Ma resulted in the formation of grabens across southern Africa into which Karoo Supergroup rocks and the early Cretaceous Etendeka basalts were lowered (Raab *et al.*, 2002). Post-Karoo faulting displaced rocks several hundred metres in western Zimbabwe, in the Luangwa and Zambezi rifts and along NNW-trending faults in southern Botswana. Cretaceous kimberlite pipes probably also intruded along zones of structural weakness and this is evident in northern Angola where they occur in a northeast-trending zone (the Lucapa corridor) that is believed to follow a large basement structure (de Boorder, 1982; Jelsma *et al.*, 2004).

8.3 Cretaceous drainage

Although it is debatable whether the interior of southern Africa was elevated prior to the break-up of Gondwana (eg. Doucouré and de Wit, 2003; Partridge and Maud, 1987), there is evidence that by the end of the Cretaceous an uplifted margin existed that had resulted in a drainage pattern primarily consisting of short rivers flowing from the uplifted margin towards the sea and those flowing in the opposite direction, into the interior of the continent. Rivers at this time would have preferentially flowed along easily erodible structures and soft lithologies and commonly exploited the down-faulted grabens filled with Karoo rocks. The Karoo-filled Cabora Bassa, Mana Pools and Mid-Zambezi basins have been exploited by rivers which have begun to erode the relatively soft Karoo sedimentary rocks and the Harts River is partly controlled by valleys formed during Dwyka glaciation (du Toit, 1910). The Limpopo River has exploited a failed rift extending from a triple junction near Nuanetsi in southeastern Zimbabwe towards the east.

The interior of the continent was covered by the fairly flat topography of the African Surface. Limited amounts of uplift or erosion were believed to have been taking place in the mid- to late-Cretaceous, as is evinced by the generally good preservation of Cretaceous kimberlite pipes in Botswana (Hawthorne, 1975), although Rayner *et al.* (1991) believe as much as 50-100 m of rock has been eroded from the Orapa area subsequent to the emplacement of the Orapa kimberlite pipes at around 92 Ma. The rivers flowing across the interior of southern Africa followed a strong NW-SE course, parallel to the western coast of southern Africa and possibly following structures formed during the break-up of Africa and South America. The probable configuration of the Mid-Cretaceous drainage is summarised in Figure 8.1. The formation of the Kalahari Basin in the Late Cretaceous disrupted the existing drainage patterns and back-tilted some of the rivers into the interior of the continent.

8.4 Basin formation: Uplift or downwarp?

Deposition of the Kalahari Group sediments started when drainage patterns were disrupted by vertical changes in the southern African topography. The formation of the interior basin may have occurred due to the uplift of the areas surrounding the Kalahari Basin, by downwarping of the

interior, or by a combination of the two.

The Chad Basin is an example of a basin formed by uplift of the areas adjacent to it. It was formed largely as a result of the emergence of volcano-capped swells around its perimeter (Burke, 1976, 1996) and some 500 m of sediment has accumulated in it in the last 30 million years (Burke, 1976). Lake Victoria has also formed in a topographical low created between the uplifted flanks of the western and eastern branches of the East African Rift System (Fig. 5.20).

It has been shown that if land is uplifted across the course of a large, strongly flowing river, the river will merely cut down through this uplifted area, forming a gorge (e.g. Ollier, 1991). The presence of several large rivers flowing across the interior of southern Africa prior to the formation of the Kalahari Basin suggests, therefore, that uplift of adjacent areas may not have been the main cause of basin formation. If the Kalahari-Zimbabwe Axis had risen across the course of the Zambezi as has been previously suggested (eg. Moore and Larkin, 2001), then the Zambezi River would probably have cut a gorge through the uplifted axis and continued its course to the Limpopo. While the inability of a river to cut through a line of flexure can be explained by a change to more arid climatic conditions (e.g. Moore, 1999), or by river capture of its headwaters, this does not explain the large volumes of sediment deposited by the rivers in the newly formed basin (see isopach map). A more likely scenario is that the rivers were back-tilted by downwarp of the basin itself. The sub-Kalahari topographical surface generated as part of this study (Appendix E) provides evidence of subsidence, which may have been of varying degrees in different parts of the Kalahari Basin. Topographic profiles across the sub-Kalahari topographic surface along approximate paths of the southward-flowing Zambezi, Okavango and Kwando Rivers (Fig. 8.2) show that the base of the Kalahari Basin floor is depressed below the Okavango Delta and Makgadikgadi Basin. While some of this is subsidence is due to later rift related subsidence as well as sediment loading, it is probably largely due to Late Cretaceous downwarp of the interior of Botswana, which would have been enough to back-tilt the drainage. The low-altitude of the surface underlying the Kalahari Group sediments in the Etosha region (~600 m.a.s.l) also suggests substantial downwarp occurred there. An analogous situation of basin formation and drainage backtilting can be found in the Cenozoic Murray Basin in southeast Australia where subsidence of the Murray Basin resulted in the back-tilting of northward-flowing drainage and the separation of the Murray Basin from the Eromanga Basin to the north (Ollier, 1995). In the case of the Kalahari Basin, the epeirogenic flexure axes of

du Toit (1933) and others may have represented axes of relative uplift with subsidence on one side of the axes leading to the formation of the Kalahari Basin.

There are two possible mechanisms for the downwarp of the interior. The first model for the formation of intracratonic sag basins involves convective down-welling of the asthenosphere beneath the lithosphere. The development of a descending plume results in a depression of up to 600 m which can be further depressed when loaded with sediment, and if the descending plume is removed, the basin may then be uplifted and eroded (Middleton, 1989). An alternative hypothesis (e.g. Lambeck, 1983; Karner, 1986) suggests that in-plane compressive stress can result in peripheral uplift and downwarp of the central depression. While the first model suggests uplift occurred subsequent to downwarp and the second suggests that it occurred at the same time as downwarp, it is unclear which mechanism resulted in the formation of the Kalahari Basin. Post-depositional uplift has certainly occurred, but whether some of the uplift was related to the rebound following the removal of a descending plume or if it was all related to the sub-continental-scale formation of the African Superswell is unclear. The mechanisms for the formation of the African Superswell will be discussed later in this chapter. A characteristic of the in-plane stress mechanism is that the large peripheral uplift results in widespread clastic deposition in a basin with a general gradation of coarser material on the basin edges to finer material in the basin centre (Middleton, 1989). Borehole and outcrop evidence is insufficient to conclusively ascertain if there is a general coarsening of material from the edges of the Kalahari Basin to its centre, and as discussed above it does appear as if several sub-basins or depocentres, each with varying degrees of subsidence, may have formed. The subsidence may have reactivated the older structures discussed earlier in this chapter and as a result was largely controlled by their NE-SW and NW-SE orientation.

We can conclude that whichever mechanism of basin formation was involved, it is probable that basin subsidence was indeed the main controlling factor initiating Kalahari Group deposition. The subsidence probably involved formation and reactivation of faults and created the back-tilted drainage and accommodation space for sediment deposition. Later uplift of the periphery of the basin as well as along certain flexure axes within the basin would have accentuated the back-tilted/inward flowing drainage and increased sedimentation through the exposure of rocks to erosion and the generation of greater accommodation space.

8.5 Initial Kalahari Group deposition

Basal lithologies of the Kalahari Group in south Africa are similar to Cretaceous sediments found in Angola and it is probable that down-warping of the Kalahari Basin to the north of the Kalahari-Zimbabwe and Etosha-Griqualand-Transvaal (E-G-T) axes in the Late Cretaceous caused back tilting of the drainage away from the Limpopo and lower Kalahari Rivers respectively and into the newly formed Kalahari Basin where sedimentary deposition began. The sedimentary succession of the offshore Orange Basin shows the deposition of hemipelagic claystone beds occurred across the continental shelf in the Early Turonian (~ 93Ma) and mid-Coniacian (~ 86 Ma) (McMillan, 2003). These beds are believed to accumulate in parts of the basin where there is no great supply of coarse clastic (quartz sand) material (McMillan, 2003). McMillan (2003) suggests that their accumulation is a response to a decrease in the amount of clastic material reaching the coast, following the tectonic disruption of the continental interior and its drainage pattern. According to McMillan (2003) the bed-load of the rivers would be trapped in newly formed depressions and lakes in the interior. The Kalahari River is believed to have drained a large area of southern Africa prior to downwarp of the Kalahari Basin, and would have been responsible for a large proportion of the sediment deposited in the Orange Basin. Any Late Cretaceous disruption of the flow of the Kalahari River may, therefore, have removed some of the clastic component being deposited offshore. If the model of Moore and Moore (2004) is correct and most of the rivers in South Africa drained into the Kalahari River at this time, this would have had an even more significant influence.

In general, early deposition of the Kalahari Group sediments probably occurred in valleys with scree deposits accumulating at the base of slopes and alluvial gravels being deposited in channels. Alluvial fans may have formed due to episodic flooding, with some rounding and sorting of the upper gravels by the streams flowing over the fans and into the valleys. As down-warp continued, so river channels became choked with gravels and sand and conglomerates. At this stage the upper Zambezi and other rivers probably terminated in the Kalahari Basin in Palaeo-lake Makgadikgadi in much the same way as the Okavango River does today, and the proto-Upper Zambezi would at this stage have been depositing its full sediment load into the Kalahari Basin. The thick clay beds in the south and central Kalahari, reaching over 100 m in thickness in northern Namibia (Miller, 1992a), provide evidence of accumulation of fine-grained sediments in large, shallow, saline lakes and in sluggishly flowing braided stream networks, although in some areas the clays may be produced by *in situ* weathering of underlying beds (Far *et al.*, 1981; Bootsman, 1998). Gritty sandstones were deposited across a large area as material was washed in from the basin margins and interfingering of clay and sandstone layers and vertical and lateral gradation between the clays and the sandstones occurs throughout the basin, possibly representing channel and overbank deposition. Aeolian sands may have contributed to some of these early deposits. Generally, however, borehole evidence suggests that deposition of more sandy material continued for some time after the clays were deposited and in most cases the sandstones overlie the clays.

8.6 Regional uplift

Although southern Africa is known to have undergone episodes of Cretaceous uplift, the origin of the anomalously high elevation of the southern and east African plateau areas collectively called the African Superswell (Nyblade and Robinson, 1994) is debatable. The Kalahari Basin itself, occupies an elevated position, with the altitude of the lowest point of the floor of the basin floor still some 600 m above sea-level (m.a.s.l), and the flanks of the basin elevated to some 1300 m.a.s.l. Figure 8.3 shows east-west topographic profiles across the sub-Kalahari topographical surface of southern Africa. While uplift of the basin could possibly be explained by the removal of a descending plume beneath the basin, this does not explain the huge regional extent of the African Superswell. Two more likely possibilities are that either the basin developed on an already elevated interior (eg. Partridge *et al.*, in prep.), or that more recent uplift of the entire southern African region has occurred, along with the Kalahari Basin (eg. Burke, 1996).

According to Lithgow-Bertelloni and Silver (1998) there are two possible causes of large scale anomalous elevation of a continent: 1) changes in average density and/or thickness of the lithosphere, and 2) vertical motion of the continent in the absence of faulting or folding. The mechanisms for these main causes of elevation fall into three main groups: The first involves the heating of the lithosphere, the second involves processes of dynamic upwellings of the base of the lithosphere generated by flow in the underlying mantle, and the third involves uplift as a result of mantle density heterogeneity based on subduction history. The latter cause has generally been discounted because it has been seen to result in a very broad scale topographic high in the entire

Atlantic basin (Lithgow-Bertelloni and Silver, 1998), and would not cause a distinctive anomalous feature like the African Superswell.

The presence of widespread Cenozoic rifting and associated volcanism in eastern Africa shows the presence of a thermal anomaly within the lithosphere in that region and this is partly seen as evidence for a theory of lithospheric heating (Nyblade and Robinson, 1994). The high heat flow measurements taken from the mobile belts of southern Africa are also used as evidence for lithospheric heating rather than for crustal heat generation, and heat flow observation from the Atlantic Ocean to the southwest of the continent also suggest that the lithosphere beneath this area is thermally perturbed. Nyblade and Robinson (1994) cite earlier studies of bouguer anomaly data which suggested that the lithosphere in eastern Africa has been thinned both on the edges and in the centre of the East African plateau causing isostatic uplift. Further evidence for anomalous heat flow underneath the superswell may come from the fact that the African Superswell lies within the long-wavelength African-Atlantic geoid high, and above a deep mantle region characterised by low seismic velocities (Nyblade and Robinson, 1994). Burke (1996) suggests that the uplift of the African Superswell resulted from Africa having come to rest over a circulating mantle, with the swells forming as a response to plumes in the underlying mantle. Ritsema and van Hiejst (2000) believe, however, that apart from the East African Rift System there is insufficient evidence for the broad thermal anomaly in the lower mantle and for anomalous low-velocity structures in the upper mantle. They therefore could not accept that the entire African Superswell is the result of uplift caused by warm and low-density material in the upper mantle.

Dynamic topography is thought to occur where upwelling coincides with a high in the long-wavelength geoid, and as mentioned above, the low seismic velocities (showing probable upwelling) and the long-wavelength geoid occur within the African Superswell. According to Lithgow-Bertelloni and Silver (1998), dynamic topography refers to the deformation of the earths' surface, supported by the vertical stresses at the base of the lithosphere that are generated by flow in the underlying mantle. In this model, large, active upwellings are generated in the basal thermal boundary layer and induce a surface boundary of deformation (manifested as the African Superswell). According to Lithgow-Bertelloni and Silver (1998) the upwelling also constitutes a significant driving force for plates in the area. Ebinger (1989), using bouguer gravity as well as topography, suggested that although the eastern Africa plateau is partially isostatically compensated by thermal alteration of the lithosphere, dynamic compensation as a result of convective processes in the asthenosphere also plays a role.

In conclusion, the anomalously high regional topography of southern Africa may be a combination of inherited Gondwana topography and subsequent uplift related to the break-up of Gondwana and the later development of the African Superswell. While the development of the Southern African Plateau is perhaps not as easily explained by the mechanisms outlined above as the East African Plateau might be, there is nonetheless abundant evidence for episodes of uplift of the Southern African Plateau during the Miocene and Pliocene, both of which contributed to the continued sedimentary deposition in the basin. Subsequent uplift events in the Quaternary that affected the topography in and around the Kalahari Basin may have been largely related to the extension of rifting that was occurring along the East African Plateau.

8.6.1 Miocene uplift and continued sedimentation

In the south of the basin, the uplift of the Griqualand-Transvaal axis in the Miocene (Partridge and Maud, 2000) possibly rejuvenated some of the rivers flowing northwards into the basin and may have led to an increase in deposition of sediment in the Kalahari Basin. The uplift may also have resulted in the elevation the gravels in the upper course of the Mahura-Muthla River to the crest of the axis. Partridge (1993) suggests that the majority of the sandstones are substantially younger than the clays and may have been deposited in the Middle Miocene or later, possibly related to the uplift which started the Post African I cycle. This later uplift may have provided the erodible material and the accommodation space that enabled deposition of sediments beyond the original depocentres. The climate is thought to have been more humid during the early Miocene (Bamford, 2000). The layers of conglomerates and grits found throughout the sequence probably represent lag deposits as well as the periodic in-washing of material from elevated areas at the edge of the basin and within the basin and this may have been related to localised uplift events. Silcretisation and possibly calcretisation of the sandstones and clays possibly occurred in the Oligocene and Late Miocene during periods of tectonic stability. Evidence from northeastern Angola suggests that the deposition of the unconsolidated sands postdates the silicification of older Kalahari Group sediments in that area (Janmart, 1953).

8.6.2 Pliocene uplift and resultant deposition of the unconsolidated sands

The massive uplift of southern Africa in the Pliocene, in particular along the Ciskei-Swaziland Axis (Partridge, 1998; Partridge and Maud, 2000) but also to a lesser degree along the Griqualand-Transvaal and Kalahari-Zimbabwe axes (du Toit, 1933; Partridge, 1998) shaped the drainage patterns of southern Africa into ones resembling those at present. Uplift on the margins of the basins is evinced by the elevation of Kalahari Group sediments and their erosion off the elevated surfaces. Figure 8.4 highlights the main zones or axes of uplift, rifting and seismicity that influenced the Kalahari Basin during the late Tertiary and the Quaternary.

Eastward-flowing tributaries of the Limpopo River are eroding the Kalahari Group sediments along the edge of the Kalahari-Zimbabwe axis. The Kalahari Basin and sediments deposited in it may have extended further to the southwest, and subsequent to deposition were eroded off the uplifted area. The exposure of basal Kalahari Group sediments (Weissrand Formation) on the Urinaib Plateau appears to suggest this and the modern Molopo River may have been cut-off from the Orange by uplift along the NW-trending arm of Moore's (1999) Etosha-Griqualand-Transvaal axis. The relatively dry climate of the Pliocene would probably have meant that the Molopo had insufficient energy to cut through this flexure. The upper Fish was captured by a more aggressive coastal river and diverted towards the Atlantic. Further to the north, the Etosha Basin was possibly drained at some stage via the Hoanib River in the southwest, but further uplift to the west of the basin about 3 Ma is believed by Stuart-Williams (1992) to have resulted in the creation of a huge lake in the basin. On the western side of the Etosha Basin basal Kalahari Group rocks have been recorded at 1300 metres above sea level (Stuart-Williams, 1992, see section 5.6) which is some 200 m higher than the upper Kalahari Group sediments further to the east. Although the accuracy of thicknesses of Kalahari Group sediments in central Angola may not be good, it is nonetheless interesting to note that great Kalahari Group thicknesses are shown on areas elevated high above sea level. Uplift along the western and eastern edges of the "Bushmanland" depocentre also occurred in the Pliocene.

The erosion of older Kalahari Group and Karoo Supergroup sedimentary rocks resulted in the

accumulation of the massive amounts of unconsolidated sand in the basin during the late Pliocene or early Pleistocene**.** These sands are a combination of the products of in situ weathering of underlying rocks, fluvially transported sands, and aeolian deposits. During a wetter period in the Late Pliocene (de Wit, 1993) aeolian sand may have been fluvially transported and deposited in low-lying areas with aeolian processes dominating during periods of aridity. Drier periods in the Pleistocene allowed dunes to form and aeolian processes have imparted distinctive textural and mineralogical characteristics to the sands both during and since their deposition. The ability of wind action to impart an aeolian overprint on sands originally transported and deposited by streams and sheetwash (Grove,1969; Thomas,1987, Moore and Dingle,1998) should be recognised when attempting to make use of the Kalahari sands for palaeoenvironmental reconstructions. In addition to this, while very good dates for dune formation have come from the advent of luminescence dating techniques, the ages obtained do not necessarily translate to arid conditions at that stage, as wind strength and vegetation cover have been shown to be important factors. The extensive dune fields of the Kalahari are, however, a fairly good indication of the general climatic conditions that existed during the last 2 million years.

The drier conditions of the Pliocene (in particular around 2.8 million years ago) also resulted in the calcretisation and silcretisation of the Kalahari sedimentary rocks. Periods of semi-arid climate in the mid-Pleistocene, Upper Pleistocene and up until present day, were also possible periods of calcrete formation. Wetter periods are indicated in calcretes by solution holes filled with sandy soil and pebbles and lined with thin laminar calcite (Coates *et al.*, 1979). Silcretes and calcretes seem to be closely related in the field (see section 3.2.4), but the palaeo-environmental significance of silcretes may be even more complex than that of the calcretes, with opposite extremes of rainfall and temperature having been suggested for their existence. The palaeo-environmental significance of calcretes and silcretes were discussed in more detail in section 4.2.4.

8.6.3 Pleistocene and Quaternary uplift, rifting and erosion

In the late Tertiary and in the Quaternary the southwest-propagating branches of the EARS began to significantly influence drainage and sedimentation in the Kalahari Basin. The most obvious indication of the extension of these branches into the Kalahari Basin from the Western Rift Valley is the seismicity, as discussed in section 5.3.5.1. The four main zones of seismicity shown in Figs. 5.22 and 8.4 can be linked to topographical features in the Kalahari Basin and in particular to those formed by faults, movement along which post-dates the main periods of dune formation. The faulting commonly followed older structures and/or strike directions of pre-Kalahari rocks and the structures related to the southwestern extension of the EARS are dominated by NE- and NW-trending orientations, possibly defining lines of Euler longitude (NW-orientation) and latitude (SW-orientation) about an Euler pole situated to the southwest of Africa (C. Reeves, *pers. comm.*). While both the eastern and western branches developed in a zone of thinned lithosphere, they developed in different ways. The formation of the Eastern rift initially involved volcanism, which was followed by uplift and then faulting. The western rift involved initial subsidence, then faulting and formation of grabens and finally volcanism. There is no recent volcanism in the Kalahari Basin, with the most recent activity being the intrusion of kimberlite pipes in the Late Cretaceous. The propagation of the rifts to the southwest appears to be involving uplift along the sides and front of the propagating rift and associated faulting. Eventually this is likely to result in the formation of a graben or half-graben as the rift develops.

Just to the north of the Angola-Namibia border the Cuito cuts through an area of uplifted and exposed pre-Kalahari basement before joining with the Cubango River on the border itself. The river continues eastwards as the Okavango until it is diverted along a northwest-trending fault towards the Okavango delta to the southeast. The northwest-trending fault forms the current pan-handle in the Okavango Delta and corresponds with the orientation of a Euler line of latitude. The axis can be extended to the southwest where it may be linked to the Otavi axis as well as to the northeast into Zambia where it is associated with a zone of seismicity (Fig. 8.4). It is thought that this "Otavi-Caprivi-Mweru" axis of uplift is a possible southwestward extension of the Mweru-Tshangalele-Kabompo Rift (Gumbricht *et al.*, 2001). This axis of rifting and uplift may be following a much older structural weakness which can be linked to the Omaruru Lineament Zone (Corner, 2000; 2004) and also coincides with a NE-trending aeromagnetic high in western Zambia (Fig. 3.1).

To the north of this axis, another southwest propagating rift is suggested by the high seismicity of zone 1 (Figs 5.22, 8.4). This zone runs parallel to the Kibaran Belt from the top of Lake Tanganyika through the Upemba graben and into the Kalahari Basin. The Upemba graben is an older half-graben feature which is possibly being reactivated with uplift of its southeastern shoulder and subsidence in the half-graben to the west. This uplifted area extends into the Kalahari Basin where it is associated with an area of uplifted basin floor (see sub-Kalahari topographical surface). Further along the line of this axis to the southwest this feature may be related to older northeast-trending structures bordering the northern side of the Etosha Basin. To the north of this axis uplift along a northeast-trending zone (corresponding with the Lucapa kimberlite corridor described earlier) may have occurred in the Tertiary with subsequent erosion of Kalahari Group rocks off the elevated areas.

The more diffuse seismic zone 3 (Fig. 5.22) is possibly related to reactivation of faults on the sides of the Kafue Basin and uplift on its northeastern side. The faults can be extended towards the southeast where they can be correlated with faults like the Linyanti and Chobe faults which divert the Kwando River towards the northeast and into the Zambezi as well as the Gomare and other faults which mark the upper reaches of the delta. Further to the southwest, similar NE-trending faults may have been responsible for deflecting the Cubango-Cuito River to the east as well as creating an uplifted ridge across the rivers' path. Further to the southeast along the same trend is the Kwango Axis which is also a known zone of uplift (Fig. 8.4).

Seismic Zone 4 clearly extends from Lake Malawi down through the Luangwa valley and along the Zambezi Valley beneath Lake Kariba (Figs 5.22; 8.4). It can be continued to the southwest into Botswana as is shown by the Kalahari Seismicity Axis of Reeves (1972a) and where it corresponds with the Trans Southern Africa Lineament Zone, an aeromagnetic feature identified by Corner (2003, 2004) roughly following the Makgadikgadi Line (Fig. 3.1). It possibly also corresponds to the STASS of de Wit *et al.* (1995). This zone represents reactivation of rifts that formed during deposition of Karoo sediments as well as those formed during the separation of Madagascar from Africa. The northwestern side of this reactivated rift coincides with the Irumide mountains as well as with an uplifted area to the northwest of Lake Kariba. Some of the high topography on the northwestern side of the Kariba-Luangwa rift may be related to uplift resulting from the extension of the Kariba-Luangwa rift from the EARS. The Bangweulu and Lukanga swamps of Zambia have formed between this uplifted area and the high topography to the northwest flanking the Mweru-Tshangalele-Kabompo Rift, on the Bangweulu block. The uplifted western flank of the Luangwa-Kariba rift can be extended through ridges of uplifted basement to the Ghanzi ridge and the faults marking the distal end of the Okavango Delta. The faults have followed older faults in the Ghanzi Group rocks as well as following the strike of the Ghanzi Group formations. Upthrow along the southern side of faults extending to the northeast of the Ghanzi ridge prevented the Okavango River from flowing towards the south and subsidence occurred on the northern side of the ridge. The resultant depression on the northwestern side of these faults has been partially filled by water and sediments and current seismicity in the area indicates that the faulting is still active and/or that sediment-loading is taking place. Overflow from lakes created in the Ngami and Mababe sub-basins at the distal end of the Okavango Delta followed the course of the Boteti River into the Makgadikgadi sub-basin which lies along an extension of the Luangwa-Kariba rift and is flanked on its southeastern side by faults and the uplifted Kalahari-Zimbabwe axis. Neotectonic faulting has disrupted dunes and drainage lines on the northern side of the Makgadikgadi Basin.

The headward erosion of the Okavango River was probably accentuated by uplift along the Otavi-Caprivi axis and it captured the headwaters of the Cuito and Cubango Rivers at some time during the Pliocene, diverting their flows towards the east. The lower courses of these rivers dried up and were buried under the aeolian sands. The Zambezi during this time was cutting back into the interior and had possibly already captured the Luangwa by the Oligocene. The Kafue was probably captured around in the Upper Pleistocene. Uplift along the Khomas and Otavi axes also influenced drainage patterns in the area and uplifted Kalahari Group sediments started being eroded from these axes.

The Upper Zambezi was captured by the lower Zambezi in the early Pleistocene and the giant lakes filling the Makgadikgadi Basin and extending over much of the Okavango-Ngami-Mababe-Caprivi areas started shrinking as rainfall was not high enough at this time to sustain the high lake levels. Periodic diversion of the Zambezi into the Okavango and Makgadikgadi Basins may have occurred at various stages during the Pleistocene, however, as is evinced by various younger shorelines and diatomaceous deposits. On the western side of the Kalahari Basin, further uplift of the western margin of the Etosha sub-basin in the upper Pleistocene resulted in the capture of the Upper Cunene by the Lower Cunene approximately 35 000 years ago. As water was diverted from the Etosha sub-basin, so the giant lake covering the basin dried up.

In the southwestern Kalahari, renewed uplift along the Kalahari Schwelle is suggested by the massive concentration of pans extending along its axis (see section 5.2.7, Fig. 5.29). The relatively small rivers that originally flowed across this area were unable to incise through the uplifted axis and became choked with sediment and pans formed in the old drainage channels.

The dry valleys or Mekgacha formed and developed through complex groundwater processes. During wetter periods in the Holocene, pans became filled with water and deposits of diatoms accumulated along with pan sediments.

8.7 Final conclusions

In the past many authors have viewed the development of the Kalahari Basin in terms of regional subsidence or interior downwarp (e.g. King, 1963; De Swardt and Bennet, 1974; Thomas, 1988b; Thomas and Shaw, 1991a), probably related to continental rift margin uplift following the break-up of Gondwana and the opening of the south Atlantic Ocean (ten Brink and Stern, 1992) and the ensuing continued uplift along the continental margin as a result of isostatic compensation (Summerfield, 1985). While interior downwarp is best illustrated by the failure of some of the interior rivers to cut channels through the basin rim, it none the less is apparent by the anomalously high elevation of the Kalahari Basin and indeed of most of southern and eastern Africa that uplift was an important factor in shaping the Kalahari Basin and controlling the deposition of the sediment in it. Much of the uplift does, however appear to have occurred subsequently to the deposition of the lower Kalahari Group sediments and possibly only since the mid-Tertiary. The later uplift resulted in rejuvenation of drainages around the edges of the Kalahari Basin possibly eroding some of the Kalahari Group sediments.

The pre-Kalahari geological history of southern Africa therefore provides abundant evidence of a history of reactivation of older structural orientations or weaknesses over time. According to Burke and Dewey (2002), once a rift has formed in a continent, any stress change can result in it being reactivated as either a topographical high or a depression. The likelihood of rifts forming and continents splitting along zones of weakness is increased when continents are in deviatoric stress when assembled into large "supercontinents" like Pangaea (Dewey, 1988). Not only did reactivation of older rift orientations occur at several times prior to the Late Cretaceous formation of the Kalahari Basin, but it appears that some of the same orientations may have been reactivated during the formation of the Kalahari Basin and are being exploited by rifts propagating from the East African Rift System today. The formation of the Cenozoic rifts in Africa can possibly be explained by the replacement of the deviatoric compressional regime of the African Plate by deviatoric tension in areas of compensated continental uplift (Dewey, 1988). There is abundant evidence of uplift having occurred in southern and eastern Africa, providing the setting for tension and extensional structures within the African Plate.

The geological and tectonic evolution of the area over the past 3.5 billion years has played an important part in determining the directions and extent of the axes of uplift and this study has shown that in order to fully understand how the Kalahari Basin evolved it is necessary to understand preceding tectonic events and their effect on what we see today. Even some of the most recent sedimentary deposits in the basin, the pan sediments, partly owe their existence to the tectonic disruption of drainage courses by axes of uplift which themselves are influenced by much older structural trends. The current deposition of sediment in the Okavango delta is occurring because of faulting along northwest-southeast and northeast-southwest structural trends that have been active at various times in the past during Koras-Sinclair-Ghanzi, Damara, Karoo and Gondwana break-up rifting and are now possibly being exploited by an extension of the East African Rift System.

The lack of dates from the Kalahari Group sedimentary rocks has meant that we cannot pinpoint exactly when in the last 70 million years the deposition of the Kalahari Group occurred. By looking at the tectonic evolution of the area prior and subsequent to Kalahari Group deposition we can, however, begin to recognise and understand the events that would have influenced the depositional processes. As we constrain some of these events through new geomorphological evidence as well as by the development of new dating techniques we will be able to constrain further the chronology of events that influenced the deposition of the Kalahari Group. While this study has been regional in extent and has to a large extent focussed on the macro scale characteristics of the Kalahari Basin, it provides a base from which more detailed research can be conducted and shows some of the complexity involved when understanding a basin that has been influenced by geological events that began over 3 billion years ago.

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APPENDICES

APPENDIX A

Table A.1 - XRF Analyses of a red clay from Sishen Mine (Ehlers and Wilson, 2001).

Sample	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	LOI	H_2O-
KGS/1	56.84	0.9	13.87	9.14	0.16	5.35	3.18	<0.02	0.2	0.03	10.53	7.18

Table A2 - XRF analysis of a borehole to the west of Sishen Iron Ore Mine, Northern Cape. Sample numbers represent the depth of sample from surface in metres.

Sample No.	%SiO_2	%TiO_2	%Al_2O_3	%Fe_2O_3	%MnO	%MgO	%CaO	%Na_2O	%K_2O	%P_2O_5	%LOI	%TOTAL	%$CaCO_3$ (calculated)
4	14.01	0.15	0.93	0.75	0.08	1.9	44.38	0.06	0.1	0.02	37.98	100.36	79.14
10	13.67	0.16	0.85	0.91	0.08	7.29	37.23	0.06	0.09	0.01	38.72	99.07	66.39
14	13.94	0.15	0.93	0.7	0.07	12.87	31.83	0.04	0.08	0.01	39.85	100.47	56.76
16	17.38	0.18	1.04	1.03	0.07	13.4	28.69	0.05	0.12	0.01	37.89	99.86	51.16
22	12.18	0.14	1.02	0.63	0.06	15.09	29.57	0.04	0.08	0.01	41.6	100.42	52.73
30	12.25	0.17	1.3	0.81	0.07	12.84	32.39	0.05	0.1	0.01	40.49	100.48	57.76
32	15.15	0.18	1.96	1.21	0.08	12.81	29.17	0.05	0.13	0.01	39.59	100.34	52.02
34	14.42	0.18	1.71	1.18	0.06	11.58	31.86	0.05	0.15	0.01	39.24	100.44	56.81
36	16.96	0.19	2.51	1.26	0.05	12.6	27.04	0.05	0.15	0.02	39.44	100.27	48.22
38	17.42	0.24	1.9	1.47	0.07	13.67	26.49	0.05	0.12	0.02	39	100.45	47.24
42	26.46	0.32	3.16	2	0.08	13.2	22.33	0.05	0.2	0.02	32.31	100.13	39.82
44	13.55	0.2	2.02	1.3	0.05	16.38	24.48	0.06	0.07	0.02	40.66	98.79	43.65
46	18.91	0.27	2.76	1.75	0.07	15.61	22.28	0.04	0.12	0.01	36.9	98.72	39.73
50	21.77	0.31	3.41	2.16	0.09	13.7	20.53	0.05	0.15	0.01	36.74	98.92	36.61
52	24.61	0.37	4.54	2.75	0.14	13.7	18.66	0.06	0.19	0.02	33.75	98.79	33.28
56	43.16	0.67	9.65	6.07	0.11	8.78	10.12	0.06	0.22	0.04	20.93	99.81	18.07
58	38.65	0.62	9.16	5.6	0.09	9.36	10.95	0.08	0.2	0.04	24.67	99.42	19.53
60	41.13	0.66	10.12	6.06	0.07	7.84	8.81	0.09	0.22	0.03	25.24	100.27	15.71
62	36.66	0.59	8.42	5.21	0.12	9.16	10.4	0.09	0.19	0.03	27.94	98.81	18.54
64	32.1	0.51	7.74	4.45	0.13	10.22	12.32	0.08	0.19	0.05	30.85	98.62	21.97
66	37.48	0.62	8.96	5.49	0.15	8.65	9.4	0.08	0.17	0.05	27.66	98.71	16.76
76	42.86	0.58	8.34	5.66	0.11	7.05	11.38	0.09	0.33	0.05	22.89	99.34	20.29
78	41.96	0.65	9.06	6.49	0.1	6.34	1.81	0.09	0.37	0.06	23.5	100.43	3.23
80	41.86	0.89	8.99	10.04	0.09	5.29	12.62	0.26	0.73	0.12	19.6	100.49	22.50

Table A3 - XRF analysis of a borehole to the west of Sishen Iron Ore Mine, Northern Cape. Sample numbers represent the depth of sample from surface in metres.

Sample No.	ppm Rb	ppm Sr	ppm Y	ppm Zr	ppm Nb	ppm Co	ppm Ni	ppm Cu	ppm Zn	ppm V	ppm Cr	ppm Ba
4	14	56	<4	48	<6	<6	<6	8	20	14	<12	84
10	12	128	<4	54	<6	<6	<6	11	16	17	<12	111
14	9	168	<4	45	<6	<6	<6	6	14	14	<12	63
16	12	164	<4	57	<6	<6	<6	7	16	20	12	80
22	9	138	<4	31	<6	<6	<6	<6	14	13	<12	45
30	12	168	<4	37	<6	<6	<6	6	15	13	<12	49
32	12	154	<4	46	<6	<6	<6	8	16	19	13	93
34	12	138	<4	45	<6	<6	<6	7	19	18	13	78
36	13	147	7	48	<6	<6	<6	10	22	21	18	92
38	14	139	<4	55	<6	<6	<6	11	26	22	24	82
42	18	135	7	88	<6	6	7	12	27	35	29	111
44	11	300	<4	39	<6	6	5	10	31	44	18	84
46	15	234	7	67	<6	7	4	12	36	39	29	100
50	17	305	9	67	<6	12	12	18	36	44	33	92
52	22	244	11	84	6	13	16	23	33	61	45	113
56	22	64	16	147	11	23	52	48	38	175	137	230
58	20	150	24	127	9	21	45	43	40	138	109	241
60	19	204	21	122	9	19	47	46	41	125	96	195
62	18	233	17	109	7	22	34	39	31	105	86	241
64	14	261	16	92	6	17	25	32	27	84	61	236
66	18	210	16	114	8	22	32	37	32	111	74	229
76	27	96	14	107	7	19	39	40	40	108	74	280
78	27	100	18	113	7	23	31	55	51	15	59	333
80	36	59	28	115	7	29	10	123	99	177	17	330