A REVIEW OF THE KALAHARI GROUP: AN AID TO KIMBERLITE EXPLORATION IN THIS MEDIUM

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ABSTRACT

The Kalahari Group sediments cover vast portions of the Archean Kaapvaal and Congo cratons that are considered highly prospective for economic kimberlites. In southern Africa, the term Kalahari refers to a structural basin, a group of Cretaceous to recent terrestrial continental sediments and an ill-defined desert, all of which have been grouped together as the Mega Kalahari by Thomas and Shaw (1993). The Mega Kalahari grouping includes sediments stretching from South Africa in the south to the Democratic Republic of Congo in the north, and from eastern Namibia to western Zimbabwe. This sand sea, at 2.5 million km^2, is the largest on earth and presents significant obstacles and challenges to the kimberlite explorationist attempting to locate bedrock-hosted diamondiferous kimberlite bodies.

The Mega Kalahari sediments represent an ancient depositional environment with a complex history in which the stratigraphy and age of the deposits are not particularly well constrained or understood. Low fossil content, limited exposure, poor differentiation of the dominant surficial Kalahari Sand and a limited comprehension of an extensive duricrust suite has delayed the understanding of the sedimentological and environmental history of the basin.

This sequence of sediments has accumulated and evolved through fluvio-deltaic, aeolian and groundwater processes, with characteristics due to primary deposition and subsequent modification being difficult to distinguish. Deposition in the Kalahari Basin has been subject to tectonic influences, changes in drainage directions and source areas of sediments, river capture and numerous large and small climatic fluctuations both in the basin and surrounding areas. It bears the imprint of recurring cycles during which the same sediments were reworked, sometimes by different agencies, all of which exacerbate attempts to correlate sedimentary units across the sequence.

The Mega Kalahari is a series of contiguous Phanerozoic sedimentary basins situated within the African Superswell. The Superswell has dominated the gross geomorphology of southern Africa and contributed significantly to the present character of the Mega Kalahari and the evolution of the drainage systems. Overall, the tectonic framework established in southern Africa by the division of Gondwanaland led to the creation of a dual drainage system, with the hingeline acting as a watershed between a coastally-orientated exoreic system and an endoreic system draining

into the interior. Deposition of sediments started in the late Cretaceous. Neo-tectonic activity expressed in the rifting in central Botswana, further influenced sedimentation rates and exerted a strong control over paleo-drainage directions

This review presents the complexities of the Kalahari cover sequence. The most important geomorphological and sedimentary factors to be considered when designing and implementing kimberlite exploration programs within the Mega Kalahari environment are outlined and discussed. New data from exploration drilling programs are presented on the thickness of the Kalahari within portions of northern Namibia, western Zambia and Botswana.

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1. INTRODUCTION

In southern Africa the name Kalahari refers to a structural basin, a group of terrestrial continental sediments and an ill-defined desert, all of which are intimately linked yet also possess distinct characteristics. There is little doubt that environmental changes have played an important part in all three of these contexts, but over differing time-scales and spatial extents. The Kalahari sediments represent an ancient depositional environment with a complex history in which the stratigraphy and age of the deposits are not particularly well constrained or understood. Collectively the sediments are referred to as the Kalahari Group (SACS, 1980). SACS (1980) distinguished 6 major lithological types in the Kalahari Group: conglomerate and gravel, marl, sandstone, alluvium and lacustrine deposits, Kalahari Sand, and duricrusts (mainly calcrete and silcrete).

In the early 1990s, Thomas and Shaw (1993) coined the phrase "Mega Kalahari" to include all the Kalahari Group sediments stretching from South Africa in the south to the Democratic Republic of Congo in the north and from eastern Namibia to western Zimbabwe. This term has been adopted for the purposes of the review.

This extensive Jurassic-to-Recent Kalahari system has been the subject of scientific interest since the early 20th century. Research in the last two decades has largely centred upon the late Quaternary history of the sediments in southern half of the Kalahari sediments.

Improved access into the remote areas of the Kalahari thirstland, development of remote sensing techniques and application of radiometric dating methods have all permitted new data to be gathered from surface sediments and landforms and have encouraged recent investigations. The result has been greater insight into the complexities of the succession, the nature and controls of landform development together with the chronological reconstruction of climatic change within the temporal range of carbon dating. Despite these advances consensus has not been a reached about the chronology of landform development, nor a parallel growth of research and understanding regarding the deposition of the Mega Kalahari.

There are several reasons for this. First, Mega Kalahari sediments contain little of economic value (except water and soda ash), so drilling and exploration has generally concentrated on what lies beneath, rather than within the sequence. Exploration companies are generally reluctant to make detailed drilling data available for public review. Consequently, with the low relief of the Mega Kalahari plateau yielding few natural exposures, little new data have been forthcoming about sub-surface Mega Kalahari sediments.

Research on late Quaternary landform development has been hampered by the lack of dateable material in some spatially extensive sediments, notably those which are aeolian in origin, and the poor environmental control of some published dates. Furthermore there has been an unconscious tendency to view the surface landforms and sediments separately from the subsurface sediments. Most of the mapping and research has concentrated on isolated parts of the basin, and to date there has been no basin-wide synthesis of lithological, stratigraphic, mineralogical and paleoenvironmental characteristics of the Mega Kalahari. As a result many questions concerning the stratigraphy and age of the deposits remain.

The aim of this study is to comprehensively review available literature on the evolution, age, deposition, development and characteristics of the Mega Kalahari sediments. The review considers some of the key factors in the long-term evolution of the Kalahari, highlighting the most significant characteristics of evolutionary events. These include development of the basin following on upon the fragmentation of Gondwanaland, sub-continent drainage development, climatic change and post-deformational modification. In doing so, the major geomorphological and sedimentary processes operating in the Mega Kalahari can be identified, and misconceptions and difficulties considered. It is the author's opinion that much understanding is to be gained by viewing the Mega Kalahari surface landforms and sub-surface sediments as a whole.

The author takes a particular interest in kimberlite exploration within the Mega Kalahari and is involved in kimberlite exploration programs throughout much of the Mega Kalahari. The Mega Kalahari sediments cover vast portions of both the Kaapvaal and Congo cratons that are highly prospective for economic kimberlites. This young, extensive veneer covers some 2.5 million km^2, of which approximately 1.6 million km^2 lie directly over cratonic crust that is prospective for diamond-bearing kimberlites. This presents formidable obstacles and challenges to the kimberlite explorationist attempting to locate bedrock-hosted diamondiferous kimberlite bodies. New data are presented on the thickness of the Kalahari within portions of northern Namibia, western Zambia and Botswana. It is hoped that a review of the complexities of the cover sequence will result in an improved kimberlite exploration strategy for prospecting in this medium.

Heavy mineral sampling is a major kimberlite exploration technique in the Kalahari and is an extensive subject on its own. Sampling in the Mega Kalahari is not discussed in any great detail in this review. Rather, it is hoped that an improved understanding of the evolution of the Mega Kalahari will assist in better interpretation of sampling results within this medium.

2. PREVIOUS WORK

In 1904 a German geomorphologist, Dr Passarge first described the Kalahari sediments of northern Botswana. He identified five major units: The Botletle beds (oldest); Kalahari limestone; Kalahari sand; deck sand and alluvium (youngest). Maufe (1915) and Rogers (1936) revised this scheme and grouped the beds described by Passarge as a Kalahari System and recent alluvial deposits, implying a chronostratigraphic relationship between the different units. Rogers (1936) felt the Kalahari System should be included as one of the great stratigraphical groups of South Africa. Due to the lack of internal fossils (Robert, 1942: Stagman 1978), inter regional correlations, and age determinations of the different lithological units were based on their relationship to erosion surfaces of supposed sub-continent-wide extent (King, 1947, 1962). Leppersonne (1952) was therefore inclined to relate beds in the Congo basin with those in the Kalahari proper, whilst Mabbutt (1955) gave the oldest units a post-Cretaceous age, attributing the Kalahari limestone to the Early Tertiary and the Kalahari Sand to the Middle or Late Tertiary.

There have been numerous descriptions since by Cooke (1957), Mallick et al. (1981), Wright 1978 and Jones (1982). More recently Partridge (1987, 1998), Thomas (1988), Thomas and Shaw (1990, 1993, 1996), Moore and Dingle (1988), Moore (1999) and Haddon (2000) have described and considered the Mega Kalahari in the greater context of tectonic setting and geomorphic evolution.

3. DISTRIBUTION AND NATURE OF THE MEGA KALAHARI

The surface units of the Mega Kalahari sediments cover over 2.5 million km^2 of central and southern Africa from the Orange River at 29°S to 1°N in the western Congo and southern Gabon (Figure 1). It is the world's largest continuous sand body and extends far beyond the ill-defined geographic area known today as the Kalahari, which is approximately coincident with Botswana (Goudie, 1970: Stokes at al., 1997). The Kalahari sediments are thus found in a range of climates today. In Botswana, conditions are predominantly semi-arid, with almost all rainfall confined to a six-month long, hot, wet season. Mean annual amounts range from about 250 mm in the south-west (more arid in neighbouring parts of South Africa and Namibia), to 650mm in the far northeast. Further north, conditions get progressively wetter and less seasonal, from sub-humid in the drier parts of the Zambezi plain to humid tropical conditions in the DRC and eastern Gabon.

In some areas the sand is formed into extensive areas of linear dunes of different ages. Elsewhere, it forms a gently undulating sand sheet. Over most of its distribution, the Mega Kalahari sand is covered with vegetation ranging from sparse scrub savannah in the south west to savannah woodland in the in the north and even tropical forest in the equatorial belt. Kalahari sand is composed of fine-to-medium-sized, dominantly (>90%) rounded to sub-rounded quartz grains, and has a red brown to ochre colour in many areas (Lancaster, 2000). In many other parts of the Mega Kalahari (Western Zambia, Eastern Angola, northeastern Namibia) the colour is typically a pale-light grey/white. It has commonly been assigned an aeolian origin, based on sedimentary characteristics (e.g. rounding, particle size) and associated landforms, but fluvial transport may be important in some areas (Moore and Dingle, 1998). Some of the "aeolian" characteristics of the sand may be derived from older (e.g. Karoo-age) aeolian sandstone. Thickness of the Kalahari sand varies widely. In the south-western Kalahari, it is 20-30 m thick and rests on an extensive limestone (calcrete) surface, whereas in northern Namibia as much as 300m of sand overlies lower Kalahari Group gravels.

Studies of Kalahari sand composition in Botswana (Baillieul, 1975) Zambia (Savory, 1965) and Zimbabwe (Thomas, 1984) indicate that the sand is derived from many local sources, including in situ weathering of bedrock, with extensive redistribution of material by wind and water, leading to spatially distinct groupings of sands of different origins. The age of the Kalahari sand is uncertain and estimates have ranged from Miocene to Lower Pleistocene (Thomas and Shaw, 1991, Haddon, 2000).

The area covered by Kalahari sediments forms an extensive sand-covered plain or plateau with only limited relief and is noted for its almost constant altitude around 1000 m.a.s.l., a result of its setting in a long-term continental depositional environment. Although bounded by mountains in the south east, south and west, the Kalahari plateau has an elevated appearance in some locations, where drainage lines have incised into the surface or where head-ward erosion is causing its margin to retreat. This is no better illustrated than in Western Zimbabwe, where steep gradient tributaries of the middle Zambezi are cutting westwards into the Kalahari sediments, exposing underlying Karoo beds and creating a distinct slope (Thomas and Shaw 1988).

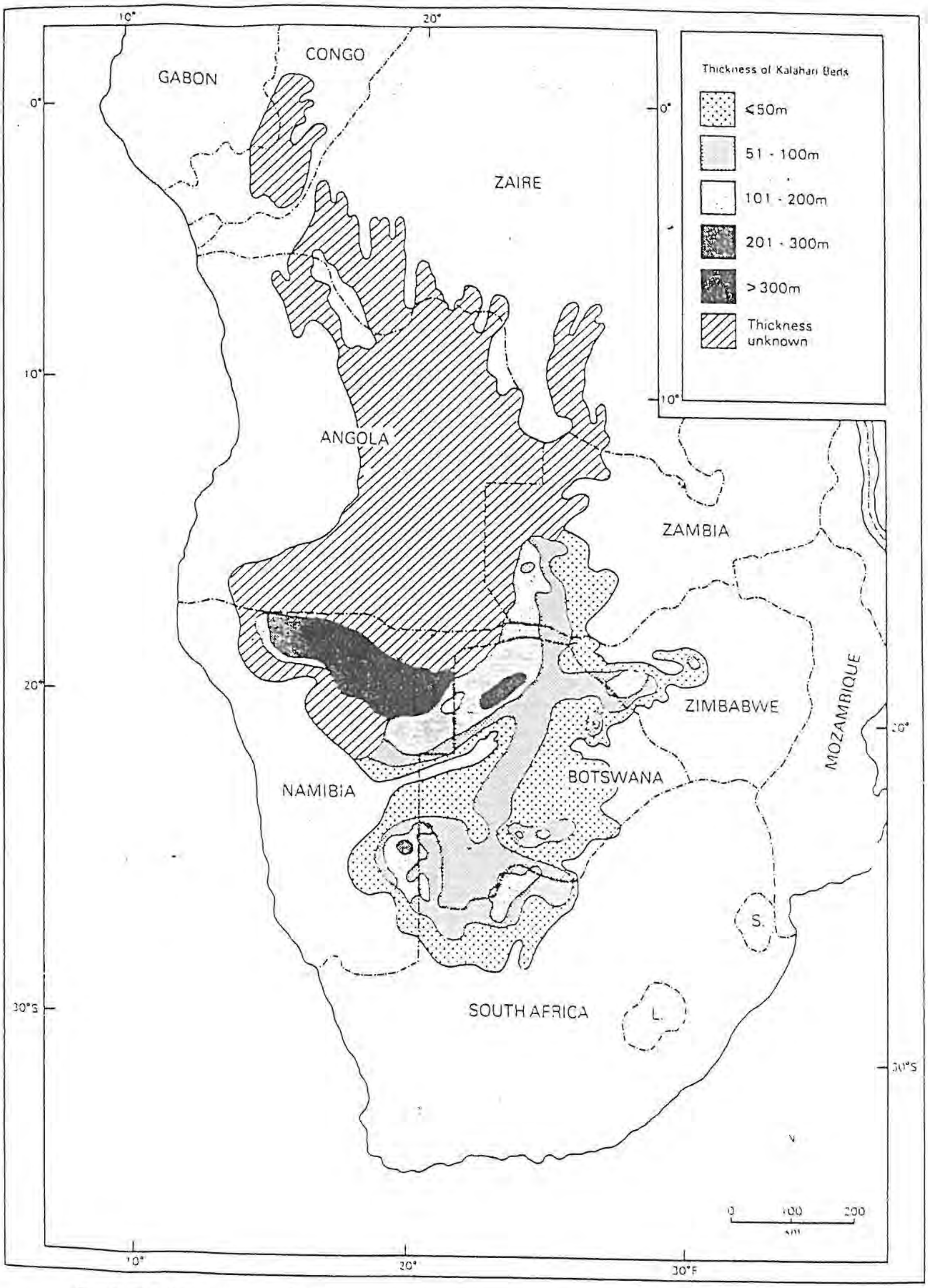


Fig. 1. The distribution and thickness of Kalahari Group sediments (modified after Thomas 1988).

3.1 DEPTH AND STRUCTURAL BACKGROUND

Borehole logs from holes drilled for hydrological and mineral prospecting purposes provide some information on the thickness of the Mega Kalahari beds and depth of the sub-Kalahari floor. They also allow identification of the general lithology of the Kalahari strata encountered in different areas.

Unfortunately, the available borehole data are restricted to an area south of approximately latitude 15° S, but this does cover around half the area mantled by the Mega Kalahari in the sub continent: in Botswana, Namibia, South Africa, Zambia and Zimbabwe, and in the extreme south of Angola (Figure 2).

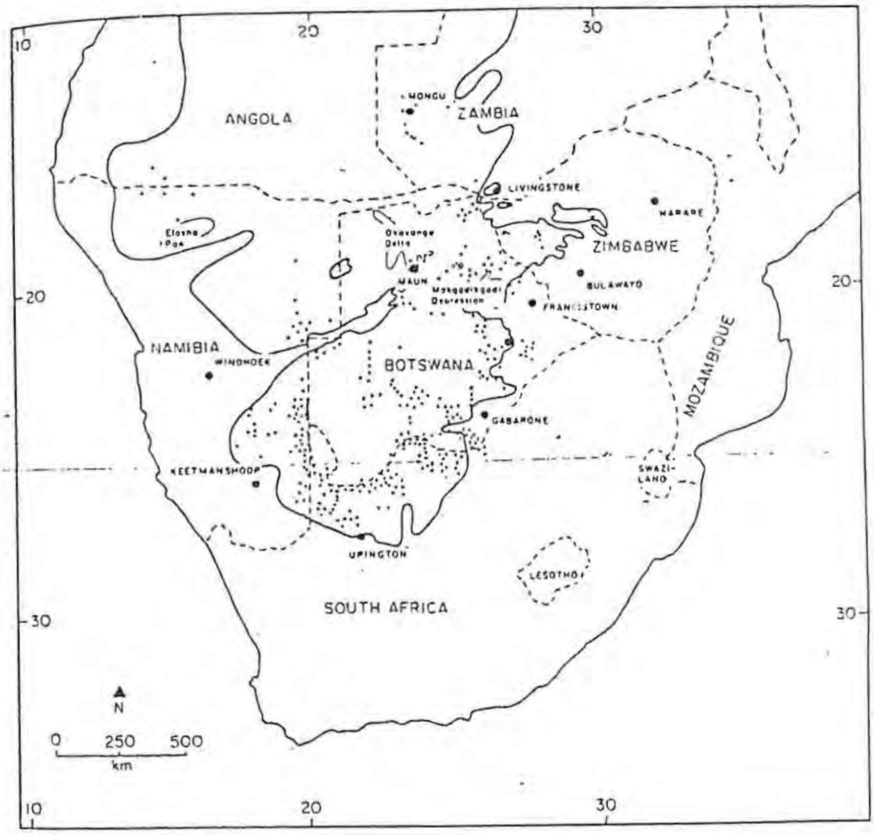


Figure 2. Location of boreholes providing data for the analysis of depths of the Kalahari Beds after Thomas (1988)

A total of 327 borehole records were used by Thomas (1988) to establish the depth of the sub-Mega Kalahari topography. A further 110 from north-eastern Namibia, where the Cainozoic beds attain great thickness (frequently in excess of 200m), did not penetrate the Kalahari/pre-Kalahari interface but were taken into account in the subsequent analysis.

The 416 records obtained from Botswana, Nambia and South Africa were obtained from the respective Geological Survey departments. Two from Angola and 8 from Zambia came from published sources (Halpenny, 1957; Money, 1972). Zimbabwean data came from an unpublished report by P. Cochran (1969). The main area for which little data were available is in central and southwestern Botswana, in the remote Kalahari Dune Desert. Based on this data, Thomas (1988) compiled an isopach map of the Kalahari beds (Figure 1) and Haddon (1999) updated and extended this isopach map further north (Figure 3).

From the isopach maps it is evident that the Mega Kalahari beds are highly variable in thickness. This cannot be attributed to surface topographical variations, since the sub-continental interior is an area of markedly low relief. Nor is it simply the result of decreasing thicknesses of sediments towards the rim of the intracontinental basin. Much of the variation in thickness can be attributed partly to recent, relatively minor rifting in the central Mega Kalahari, the relief of the pre-Mega Kalahari surface and the position and aerial extent of such depocentres as the Etosha, Aranos, and Angolan "sub-basins" (Figure 4). Whilst this region is now known to be less tectonically stable than was once supposed (Reeves, 1972), there is no evidence to suggest that the Mega Kalahari beds themselves were affected by extensive post-depositional faulting (Wright 1978) which could have influenced the preservation of differential thicknesses of sediments. In their discussion of fault and fracture lineaments in Botswana, Mallick (1981), Habgood et al. (1981) noted their low frequency within the Kalahari. This is probably due to the relatively young age of these sediments and, therefore, in the current absence of more compelling evidence, this view of limited post-deformational tectonic activity is maintained for this review. Persistent localised intracratonic rifting has, however, affected depositional rates in the Okavango Delta and Makgadikgadi Depression (Reeves, 1972; Scholz et al., 1976).

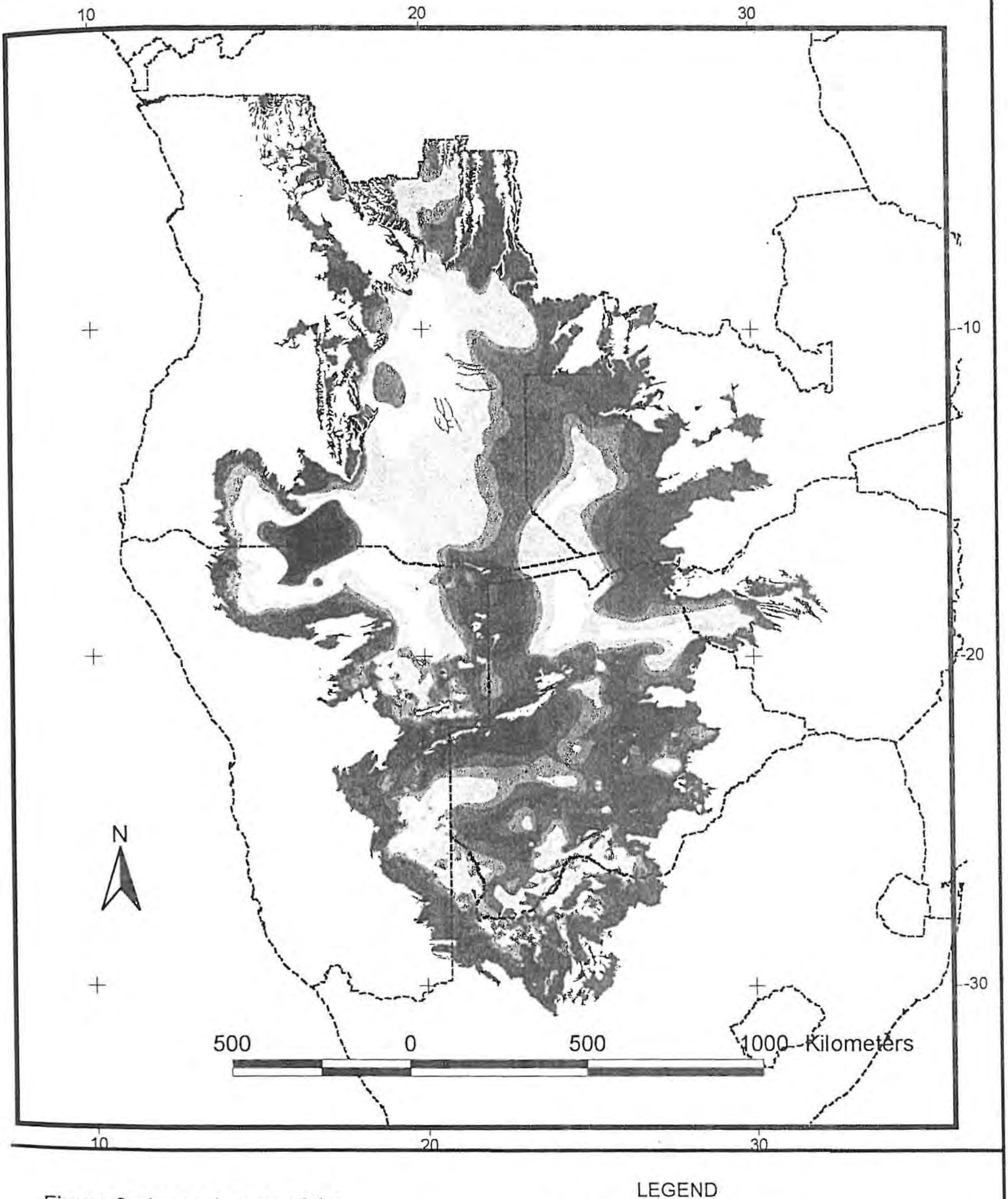


Figure 3. Isopach map of the Mega Kalahari after Haddon (1999)

LEGEND

0
1 - 30
31 - 60
61 - 90
91 - 150
151 - 180
181 - 210
211 - 270
271 - 300
301 - 330
331 - 390
391 - 450
Borders

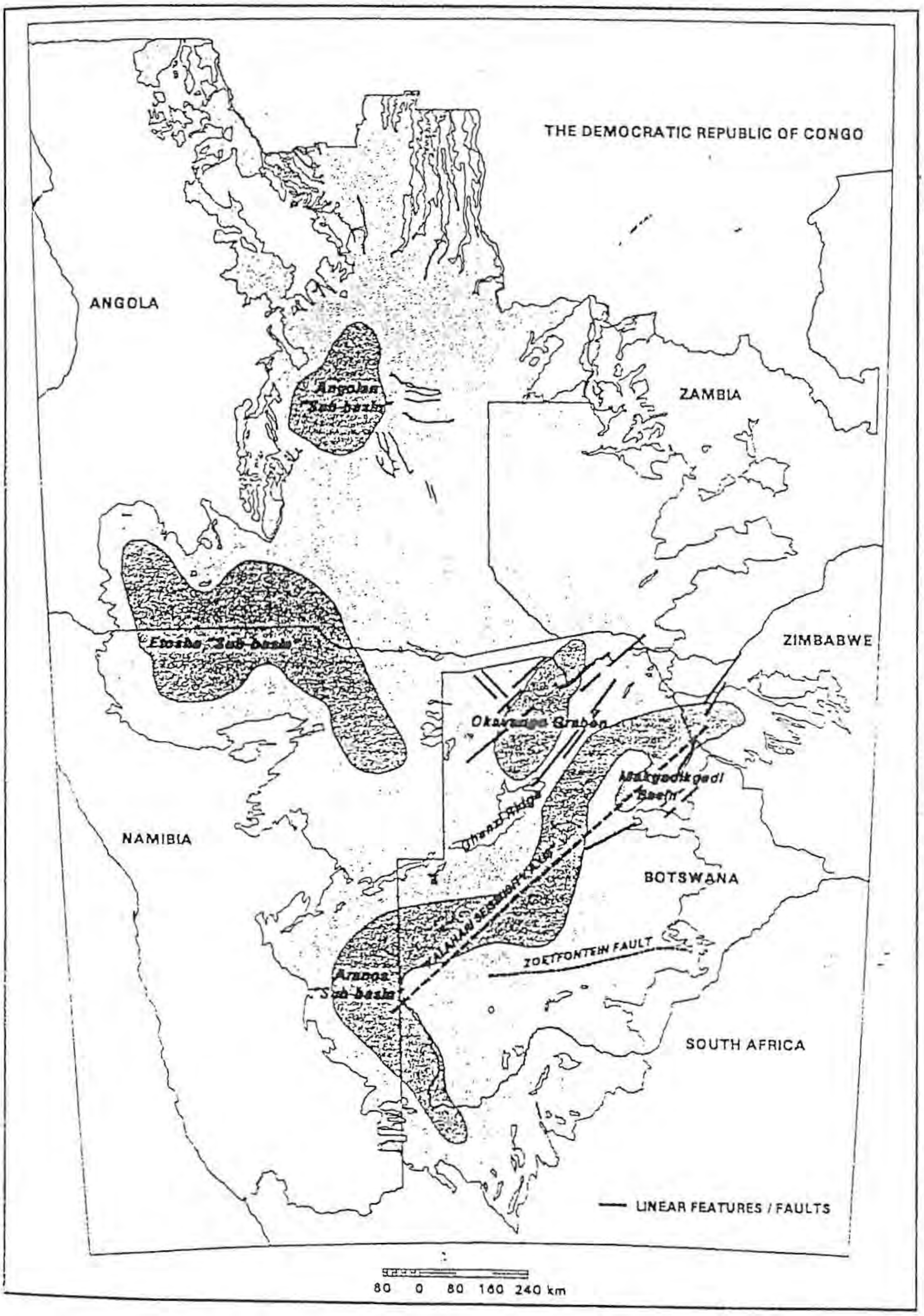


Figure 4. Principal structural features of the Kalahari Basin. Shaded areas correspond with the location of the thickest Kalahari sediments (after Haddon, 2000).

The distribution identified in the isopach map is therefore interpreted as indicating that the Mega Kalahari beds accumulated in a series of contiguous, structurally controlled Phanerozoic sedimentary basins (Figure 5): respectively the Congo basin, centred on the Democratic Republic of Congo (DRC), the Cubango/Barotse basin (Angola/N-Namibia) and the Kalahari basin, approximately coincident with the area known as the Kalahari Desert. Sedimentation in these different component basins has not been consistent through space or time, with the oldest sediments in the southern Kalahari basin being Permo-Carboniferous (Visser, 1983) while the Congo basin did not receive sediments before the Mesozoic (Burollet, 1984). In the Mega Kalahari context these basins appear to have been sedimentologically linked.

Thomas (1988) cites 3 sub-basin structures to the south of the Ghanzi ridge. The westernmost basin straddles the Namibian-South Africa-Botswana borders and contains up to 200 m of Kalahari sediments resting upon Ecca Beds. In the south-eastern basin the Kalahari Beds attain a maximum thickness of 187 m, whilst there is a third, still shallower, basin centred on 25°S 25°E. Despite the paucity of the borehole data from south-western Botswana, those that are available suggest that the deepest parts of the western basin are discrete from those in the east. Furthermore Visser (1983) identified a pre-Karoo basin in the same western part of the Southern Kalahari and suggested that the pre-Kalahari topography reflected this older Carboniferous structure.

Thicker accumulations of Kalahari sediments occur to the north of the Ghanzi ridge, with over 270 m in the Etosha area of northern Namibia. The absence of borehole data from most of Angola precludes determining whether these sediments accumulated in a discrete local basin or as part of a more extensive DRC basin in which 4000 m of Phanerozoic sediments are present. The corridor of thicker Kalahari sediments running north-eastwards through central Botswana to western Zimbabwe, may be due to increased sedimentation in a graben structure which Reeves (1978) identified in the pre-Cambrian basement rocks of the area.

Figure 5 Sedimentary basins in Africa. After Burollet (1984) and others. , Phanerozoic sedimentary basins; , Inter-basin swells; , major rifts.

3.2 NEW DEPTH DATA FROM EXPLORATION BOREHOLES

Depth data from exploration drilling in the Kalahari in northeast Namibia, western Zambia and central Botswana is presented in Figures 6b, 6c and 6d. All data is overlain on Haddon's (1999) isopachs for the Mega Kalahari. In northeast Namibia actual drilling intersected much thicker Kalahari successions (85-206 m) than anticipated from the isopach map (30-90 m). In western Zambia there is a reasonable correlation between drilling data and the isopach map except for the extreme west. In the far west the Kalahari was found to be 160-180 m thick compared to the isopach estimates of 30-60 m. Within central Botswana the drilling results also indicate actual Kalahari thicknesses greater than those indicated by the isopach map. Significant variability in depth of Kalahari across 10's of kilometres is evident in all three areas. In northeast Namibia and western Zambia the depths between holes vary by as much as 70-120 m. In Central Botswana the variability between holes was less, around 30-60 m.

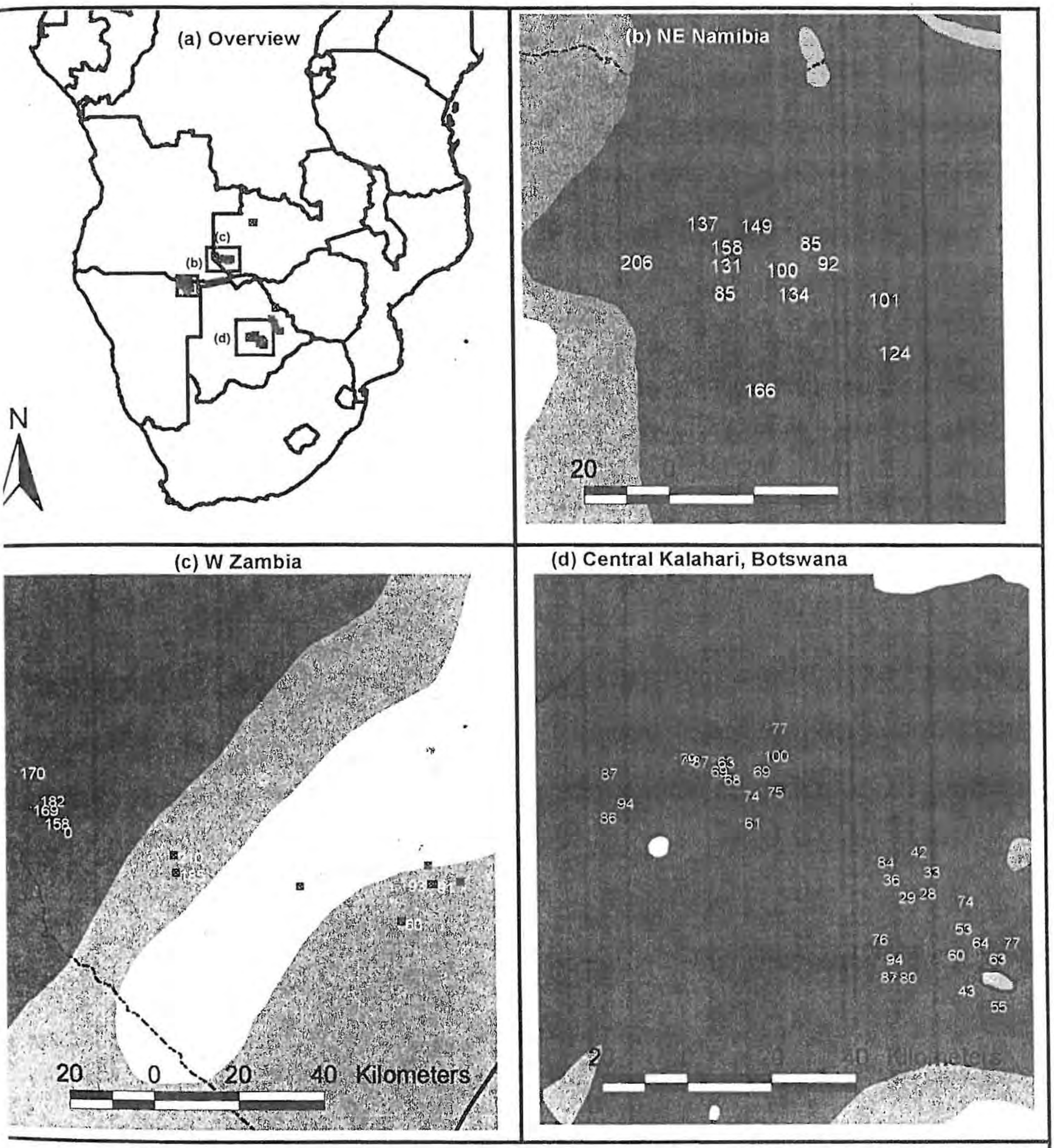


Figure 6. Exploration boreholes with depth to base of Kalahari on Kalahari isopach map: (a) overview; (b)- NE Namibia;(c)-W Zambia; (d)-Central Kalahari Botswana (Isopachs after Haddon 1999)

----- Africa Borders

Exploration drill holes, depth to base Kalahari

Kalahari isopach contours

0
1 - 30
31 - 60
61 - 90
91 - 150
151 - 180
181 - 210
211 - 270
271 - 300
301 - 330
331 - 390
391 - 450

3.3 LITHOLOGIES OF THE MEGA KALAHARI

Attempts to develop a basin wide stratigraphy have been limited by the quality of borehole data, and the absence of fossils in the lower units has made correlation over large areas very difficult. Further complications, in the form of tectonic activity, the imprint of climatic fluctuations, and the erosion, reworking, redeposition and post depositional modification of the Kalahari sediments have resulted in a sequence that is complex. The most common lithologis are described below. A schematic cross-section through the Mega Kalahari in the south of the basin is shown in Figure 7 (Haddon, 2000).

In his observations of the Kalahari, Rogers (1936) recorded conglomerates, sand, marls, ferricretes, calcretes and silcretes as the major units present.

Basal sediments

The basal Kalahari sediments are commonly gravels, conglomerates, scree deposits, and grits and cemented reworked rubble derived from underlying units, all of which rest unconformably on pre-Kalahari topography: for example, in Zambia on Precambrian to Cretaceous rocks, and infilling relict river channels in the southern Kalahari (Jones, 1982). The composition of the gravels reflects regional source area bedrock characteristics, Basalt and agate pebbles are common in Zambia (Money, 1972), often

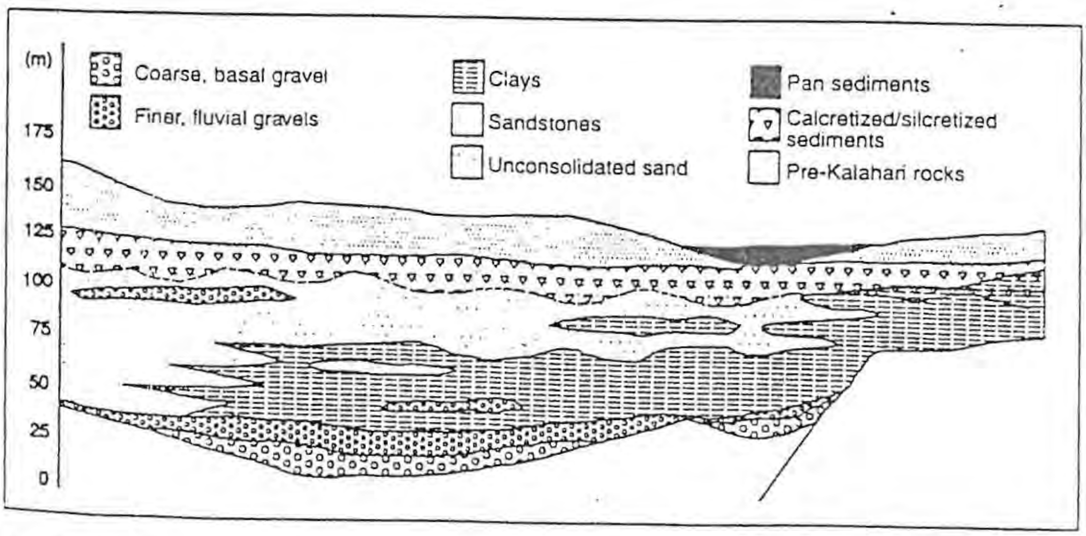


Fig. 7 Schematic, idealized representation of the Kalahari lithostratigraphy in the southern portion of the basin.

cemented into a conglomerate, and Du Toit (1954) reported that over 90 m of cemented gravels in the upper Auob area of Namibia are composed of granite, gneiss, quartz and Fish River Sandstone pebbles. These basal deposits vary widely in composition, thickness, and degree of cementation, but occur throughout most of the basin, probably reflecting the positions and sizes of late Cretaceous drainage systems and, possibly the locations of fault-bounded depressions. In general the gravels are coarser and more angular at the base with finer gravels, indicating more distant fluvial transport, overlying them. Fine fluvial gravels have also been described higher up in the sequence, where they commonly occur as thin beds. The matrix and general lithology tends to be clay-rich in the south and more sandy and gritty to the north. Evidence of reworking of this basal unit has been observed in northern Namibia by Albat (1978), and calcretisation and silcretisation have also been described (Shaw and De Vries, 1987). The gravels reach thicknesses of up to 100 m in some areas (Smit, 1977) but commonly thin and pinch out over relatively short distances. Their thickness and distribution appear to be strongly controlled by the pre-Kalahri topography.

Marls

Marls are often present above the basal gravels in borehole records in the southwestern part of the Kalahari region, as noted by Boocock & Van Straten (1962). They are also present in the Grootfontein area boreholes of northern Namibia, alternating with bands of clay and sand, but are not found in the records in Zimbabwe or Zambia, nor were they found by Cahen and Lepersonne (1952) in DRC.

Botletle Beds

Units equivalent to Passarge's (1904) Botletle Beds are found throughout the Mega Kalahari. These consist essentially of silcretised sandstone (orthoquartzite) interspersed with silcrete, calcrete, clay and ferruginous horizons. Thicknesses vary considerably, reflecting local circumstances. Equivalent names are Gres Polymorphes or Etage Moyen in DRC (Cahen and Lepersonne, 1952; Claeys, 1947), Pipe Sandstone in Zimbabwe (Maufe, 1939) and Barotse Formation in Zambia (Money, 1942).

Clays

In the south-western areas of the Kalahari calcareous clay deposits form an important component of the Mega Kalahari and reach thicknesses of up to 100 m (Smit, 1977). In the central and northern areas records of the clays are sparse, but their presence has been noted in Botswana by Boocock and Van Straten (1962), and in Namibia by Thomas (1988) and Miller (1992). The base of these clays is sometimes characterised by pebbles supported in the fine matrix and gritty and sandy lenses also occur within the clay beds (Smit, 1977). The clay may grade upwards to a clayey calcrete, and in some cases thin calcretised layers have been found within the clay.

Sandstones

Red, brown, greenish or, most commonly, yellow sandstones have been identified throughout the basin and appear to have a far greater aerial extent than the underlying lithologies (Haddon, 2000). The sandstones are poorly consolidated, display a gradational contact with the underlying clays (where these are present), and may also occur as lenses within the clay (Smit, 1977) or in troughs on the pre-Kalahari surface (Levin, 1980). The sandstones contain intercalated duricrust horizons and frequently become more calcretised and calcified towards the top, where they may be indistinguishable from the overlying calcretised sand and duricrusts. They are medium to fairly coarse grained and contain thin grit or pebble layers or an interlocking network of tubes filled with sand. Primary sedimentary structures are generally rare or absent, though Thomas (1981), Money (1972), and Miller (1992) have recognised cross-bedding in the sandstones and gritstones.

Duricrusts

The calcretes of the Mega Kalahari are amongst the thickest and most impressive of their kind in the world, representing pedogenic episodes in a semi-arid climate during Pliocene to recent times. Despite the gross inter-regional similarities in the units of the Kalahari Beds, examination of the borehole records shows that there are considerable local and regional variations. Silcrete and calcrete duricrusts cover huge areas in the Mega Kalahari and are found in multiple horizons and in a wide variety of stratigraphic connotations. This indicates that Boocock and Van Straten's (1962) identification of two main periods of duricrust formation, one either side of the deposition of the Kalahari Sand, was an oversimplification. Helgren (1984) confirmed this by identifying over a dozen silcrete facies in the Makgadikgadi basin.

The Kalahari duricrusts are modified sediments that have undergone chemical alteration, encouraged by factors such as long-term tectonic stability of parts of the Kalahari, climatic variations about a semi-arid mean, endoreic surface drainage, an active groundwater regime and the availability of silica and low valency solutes. Varying degrees of calcrete maturity are related to a number of independent factors (Watts, 1980): time, climate, host materials, carbonate source, geomorphological position, organic influences, sedimentation (or erosion) rate and various localised conditions. This interplay of such a number of parameters over an area as large as the Mega Kalahari obviously results in a highly diverse suite of calcrete types, but such variations are frequently observed even on a single outcrop. Consequently broad conclusions must be circumspect.

The predominantly pedogenic calcretes of the Kalahari possess a wide range of physical characteristics, ranging from powdery states, through nodules, to a honeycomb form, to dense hardpans with lamellar rinds, to solutionally disrupted and extremely dense boulders (Watts, 1980). In composite profiles several types of calcrete or, indeed, several examples of the same type of calcrete, may be found (Figure 8). There are some regional differences Thus, boulder calcretes are particularly common in the relatively high rainfall zone of eastern Botswana, such as at Artesia, Mookane and east of Francistown, whilst silica-toughened calcretes are common in the highly saline Makgadikgadi depression. Calcrete sequences are often in excess of 50 m thick (Watts, 1980). Simple profiles are generally < 2.5 m thick but may be thicker where sediment accretion has been constant and sufficiently slow. Borehole data indicate composite calcretes at least 3 m thick, with most being greater than 6 m. Some thick composite calcretes in the south (some are non-pedogenic) may be Pliocene in age but most of the intermediate thickness forms (6-15 m) are probably Quaternary.

Although some calcretes are developed in and on bedrock, the majority occur in superficial Kalahari sand. The thickness of the sand overlying the calcretes is rarely greater than 30 m and is usually less that 3 m (Grove, 1969).

As duricrusts can develop in a variety of situations, many pedogenic (Goudie, 1973) calcretes, silcretes and ferricretes in the Kalahari should, perhaps, not even be considered as lithofacies. In the middle and southern Kalahari this has resulted in the formation of an extensive duricrust suites, particularly in the calcrete-silcrete spectrum, whilst ferricretes are common on the eastern margin of the basin, and in the wetter north.

In general terms, calcrete predominates in the southern Kalahari and on the Bakalahari Schwelle. Calcrete profiles of 60 m thickness along the Molopo valley and 100 m at Khakea, in southwestern Botswana, have been reported (Goudie, 1973). In the middle Kalahari, silcretes are associated with the distal end of the Okavango Delta, along rivers with annual flow regimes and in lower parts of the Makgadikgadi basin, where calcretes occur in association with the shorelines of the paleo-lake system (Shaw, 1987). The increasing salinity of both ground and surface water towards the Makgadikgadi sink, culminating in the massive alkaline beds of Sua Pan, plays an important part in the distribution of calcretes in these areas.

(a)

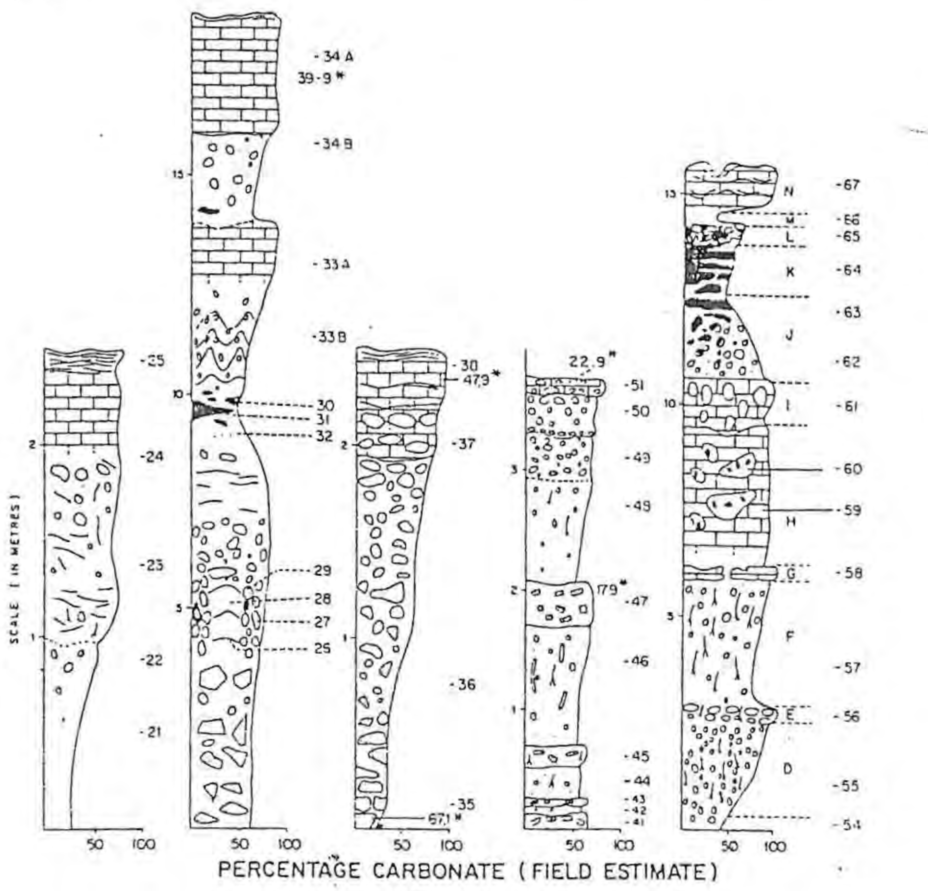


Fig. 8 (a) Selection of logged calcrete profiles (simple and composite) from the southern Kalahari. Key to all logs is shown in Fig. 3(b). Localities: profile 21–25: borrow pit on west side of road, Phephane Valley, just above confluence with Molopo, 21 km from Vorsterdorp; profile 26–34B (see Fig. 2); profile 35–38: cliff exposure on Precambrian quartzite, south side of Molopo Gorge, near Bogogobe; profile 41–51: cliff exposure, south side of Molopo Gorge, 8 km west of Bogogobe; profile 54–67 (see Fig. 2). Numbers with asterix are Schmidt Hammer readings taken at outcrop, and other numbers refer to field sampling locations.

(b)

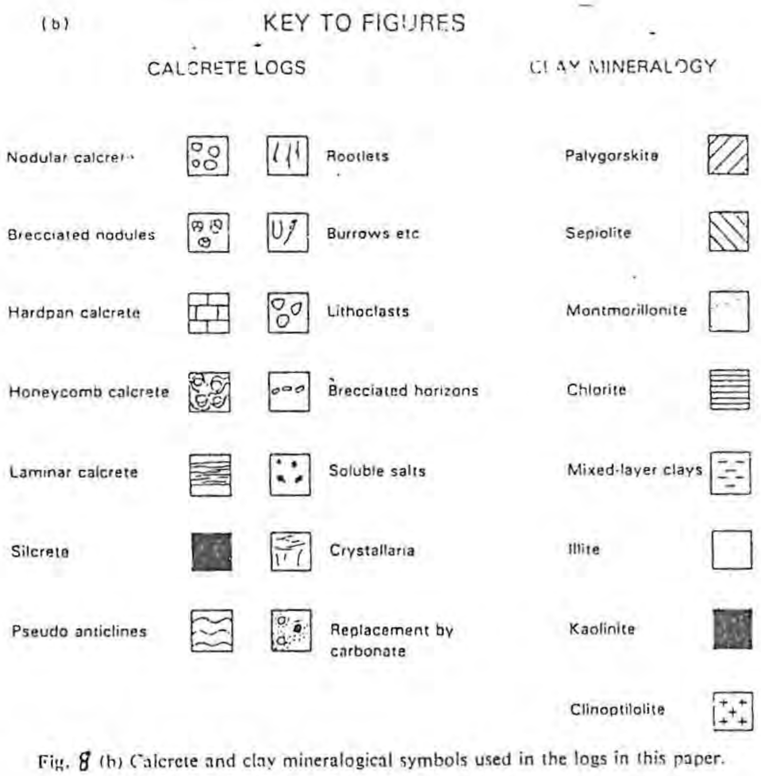


Fig. 8 (b) Calcrete and clay mineralogical symbols used in the logs in this paper.

Despite this, the presence of calcrete, silcrete and fericrete together at some sites suggests that lithological and hydrological variations are important at the local level.

More commonly, however, calcretes overlie the sandstones which grade upwards from a sandstone to a calcareous sandstone and finally to a sandy calcrete (Thomas, 1981). The calcretes contain layers of conglomerate and grit, which represent occasional fluvial influxes from the edge of the basin, and calcretised lag or scree deposits. More than one age of calcrete is represented and recent nodular or powder forms, occurring at shallow depth or at surface between dunes (Levin, 1980), suggest that calcretes may still be forming. Silcretes are not as prominent in the south of the basin but have been reported to be widespread in the central and northern Kalahari where the climate is wetter. Pedogenic horizons intermediate between calcrete, silcrete and ferricrete occur throughout the basin, and calcrete is often replaced by silcrete and vice versa.

Sand

Throughout much of the Mega Kalahari, the surface unit of the Kalahari Beds is the unconsolidated Kalahari Sand. Equivalents are the Plateau Sand (Maufe, 1939), Sables Ochres (Cahen and Lepersonne, 1952) and Mo̊ngu Sand (Money, 1972). Locally, other sediments can assume importance at the surface of the Mega Kalahari, notably in northern Botswana, the Okavango alluvium and Makgadikgadi pan and lacustrine sediments.

The Kalahari Sand is also present in a number of other stratigraphic relationships in the upper horizons of the Kalahari beds from all the basins. This unit has been more fully studied than any other unit in the Mega Kalahari. Nonetheless, debate on its origin, mode of deposition, age and environmental significance continues. The Kalahari sands vary in colour, composition, thickness and age and in many areas of the basin have been reworked and redeposited more than once in their history. The sands are generally red, but may be whitish, particularly in poorly drained areas where hematite coating of the grains has been removed by shallow groundwater action, and may even be black in the vicinity of dolerite outcrops (Thomas, 1981). The sands are mostly fine to medium in grain size, are usually well sorted, and consist of sub-angular to sub-rounded grains. The dominant mineral is quartz with subsidiary feldspar, rock fragments and biotite. Occasional silcrete fragments and grains of calcite and dolomite occur with, on a larger scale, sporadic blebs and stringers of brown and green clay.

The red colour of the sands is due to haematitic and limonitic coats around the detrital grains. This iron is attributed to in-situ weathering of ferro-magnesian minerals (e.g. biotite, hornblende)

It is generally accepted that the sands have an aeolian origin (Bond, 1948) but the importance of in situ weathering of Karoo bedrock as a major contributor to the release and formation of sand in some areas has been recognised (Baillieul, 1979). Locally derived sand was probably supplemented by fluvial and wind transported detrital material (Baillieul, 1979; Schlegal et al., 1989; Moore and Dingle, 1998).

In the Etosha basin 280 m of uninterrupted sands lie over Karoo Beds (Figure 9).

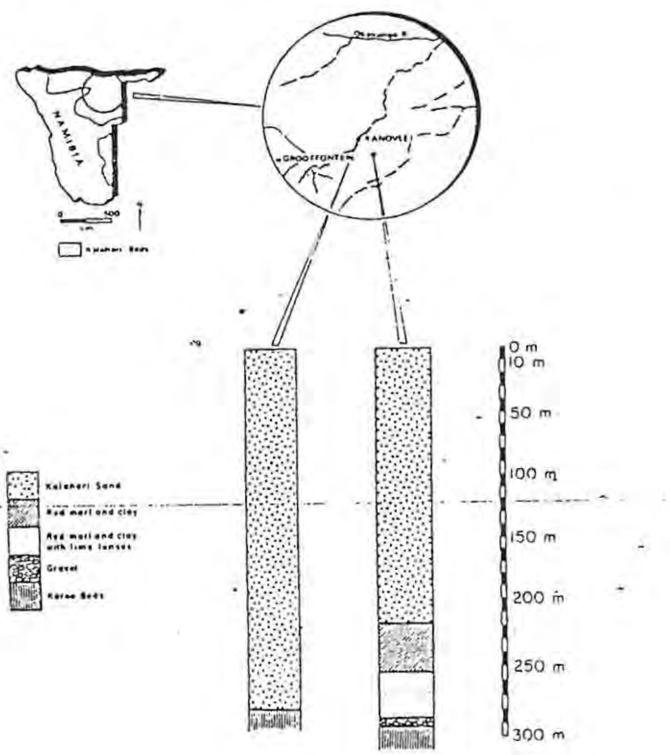


Figure 9. Variations in Kalahari Bed Stratigraphy in Groottfontein area, Namibia after Thomas (1981)

In western Zimbabwe, multiple horizons may be observed to be interbedded with calcretes and silcretes. Although one name is generally applied to this sediment facies over a considerable lateral extent, closer examination shows that it is variable on a number of counts. Poldervaart (1957) commented on colour variations whilst Baillieul (1975) identified four types of sand in Botswana after considering, amongst other factors, the relationships of surface sands and underlying non-Kalahari strata. Trapnell and Clothier (1957) identified three types in Zambia. Variations in heavy mineral content have been used in attempts to differentiate the provenance of the sand (Bond, 1948; Poldervaart, 1957).

Over large tracts of the interior of southern Africa, the Kalahari Sand has been shaped into three systems of Quaternary paleodunes of different ages and spatial extent (Lancaster 1981; Thomas, 1984). These aeolian landforms probably represent episodes of sand reworking rather than original deposition, which has variously been dated as Tertiary (Maufe, 1930,1939), Miocene (Cahen and Lepersonne, 1952), Pliocene (Mabbutt, 1955), Plio-Pleistocene (King, 1962) and Pleistocene (Bond, 1948). A more detailed description of the unconsolidated sands occurring in the Kalahari Desert of Botswana is provided in Section 3.4.

Pans

Shallow closed depressions, which are only intermittently inundated, known as 'pans', are found throughout the Kalahari and, to a much lesser extent, also in other parts of Botswana. Pans are a characteristic feature of the Kalahari and some contain fairly complex sequences of sediments, with contributions from flooding, aeolian activity, groundwater movement, weathering and duricrust formation. The sediments mostly consist of laminated grey and white silts and clays, occasionally containing pockets, fissure infillings or thin layers of more sandy material. They are frequently capped by calcretes, silcretes or evaporite deposits. During one or more wet intervals of the Quaternary, diatoms and molluscs lived in these pans and contributed to the formation of diatomaceous limestone or tufa deposits (Netterberg, 1980). These diatomaceous limestones are grey in colour, of low density, fairly soft and contain many hollow tubules, the latter resulting from root growth and the activity of burrowing organisms.

The Kalahari Group received review by Thomas and Shaw (1991) and their stratigraphy for the group is shown in (Figure 10).

Passarge (1904) Kalahari (Botswana)	Maufe (1939) Victoria Falls	Cahen & Lepersonne (1952) Zaire	Money (1972) Zambia
Alluvialle Bildungen (Alluvium) 1. In swamp zone 2. In sandveld Decksand			Zambezi formation (3. Limestone and clays on pan floors) (2. Various interspersed duricrusts)
Kalahari Sand (Divided into 4 subgroups) Basal river gravels	Kalahari Sand Carstone Nodule Bed	Sable ochres	1. Mongu sand member (of Zambezi formation)
Kalahari Kalk (Limestone) 2. Young marls and Pan deposits 1. Calc – sinter and calcareous sandstone	Pipe Sandstone	Lower Barotse	(Upper Barotse formation) (Middle Barotse formation)
Botletleschichten (Botletle Beds) 3. Pan sandstone and sandy limestone	Kalahari-Chalcedony	Gres polymorphes Silicified sandstone Chalcedonic limestone	Lower Barotse formation 2. Cemented sandstone with pipes
2. Chalcedonic sandstone 1. Cemented weathering debris of underlying geology			
↓		Kamina series Gravels and sandstones	1. Basal conglomerates

NOTES
Horizontal lines indicate implied correlation between areas

↓ Alternative correlations from literature

( ) no direct correlations implied

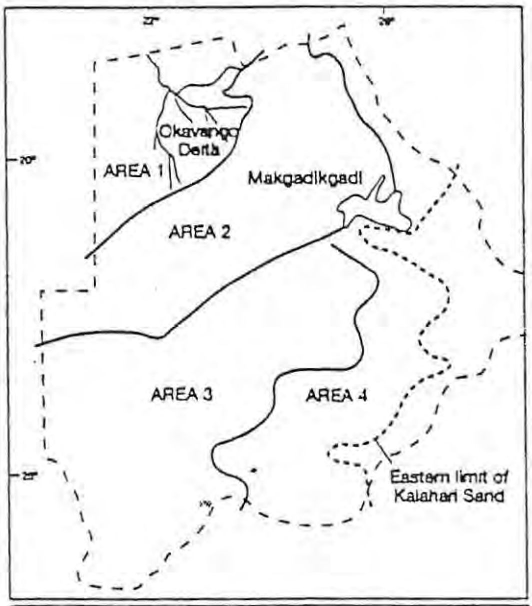
Pipe sandstone has been correlated with the Gres Polymorphes, and therefore the Botletle Beds, by Cahen & Lepersonne (1952), but with the lower unit of the Kalaharikalk by Maufe (1939)

Figure 10 Stratigraphy of the Kalahari Group sediments (after Thomas & Shaw, 1991*a*). Horizontal lines indicate implied correlation between areas; ↓, alternative correlations from literature; ( ), no direct correlations implied.

3.4 KALAHARI DESERT

3.4.1 Extent and Character

Thomas and Baillieul (1975) have shown that the sands of the Kalahari Desert in Botswana are of four major types, rather than being homogenous as was once thought (Figure 11). Sand type 1 is a fine grained quartz sand found in an area of old longitudinal dunes in the north-western part of the country. The encroachment of the Okavango Swamps has caused the erosion of the dunes over a large area. Type 2 is a sand derived from two sources: the decomposition of the Ghanzi Sandstone, and a transported aeolian component. Sand type 3 is very similar statistically to type 2 except that it lacks the distinctive feldspar component in the finer fractions. It is assumed to be derived from the underlying, fine-grained Karoo sandstones. Type 4 sand occurs in the eastern part of the country where the sand cover is the thinnest (less than 15 m). Here, the characteristics of each sub-variety of sand are determined by the bedrock that acts as its dominant source rock.



Area 1:	Pure quartz sand of eolian origin
Area 2:	Eolian quartz sand plus fine component derived from weathering of underlying Ghanzi sandstone
Area 3:	Well-sorted sand with component from underlying Karoo sandstone
Area 4:	Sands of fluvial and reworked eolian origin

Figure 11. Kalahari sand types in Botswana (after Baillieul, 1975)

Sand types 1, 2 and 3 contain aeolian dune features indicative of a period when Botswana was much more arid than today. It was during such arid periods that the Kalahari Desert derived its aeolian aspects. Except for one or two restricted areas the desert today is stabilised by vegetation and aeolian processes are no longer active. Modification of the sand distributions in these areas now occurs by fluvial processes. Sand type 4 displays an entirely fluvial origin and distribution, and any resemblance to an aeolian sand is due to derivation from a source rock that is of Aeolian origin.

It has been shown that a statistical and microscopic analysis of a surface sand sample can be used to determine the nature of the underlying bedrock in areas where the transport of source material has been areally limited. This is especially true where sand cover is less than 15 m thick (type 4 sand) Beyond this depth the effects of bioturbation are less and the amount of material exchange diminishes.

3.4.2 Dune Systems in the Kalahari Desert

Extensive systems of dunes, mostly of linear form occur in the northern and south-western Kalahari Desert. The dunes can be divided into three groups (Lancaster, 1981; Thomas, 1984) an eastern group centred in western Zimbabwe and adjacent areas of Botswana; a northern group situated west and north of the Okavango Delta; and a southern group, in the southwestern Kalahari. (Figure 12).

The eastern group consists of degraded linear dunes up to 50 km long with a ridge width of 500-2500 m and a spacing of 1500-2500 m (Flint and Bond, 1968; Thomas, 1984). Dune alignments change systematically from ESE-WNW in the east and north of the area to ENE-WSW in the west (Stokes et.al., 1998). The dunes are generally less than 10 m high, although heights of as much as 25 m have been recorded (Thomas,1984). The dune ridges are covered with an open savannah woodland, whereas grassland dominates the interdune areas. Small barchanoid and transverse ridges located west of Makgadikgadi depression (Grove, 1969) are also included in the eastern group of dunes.

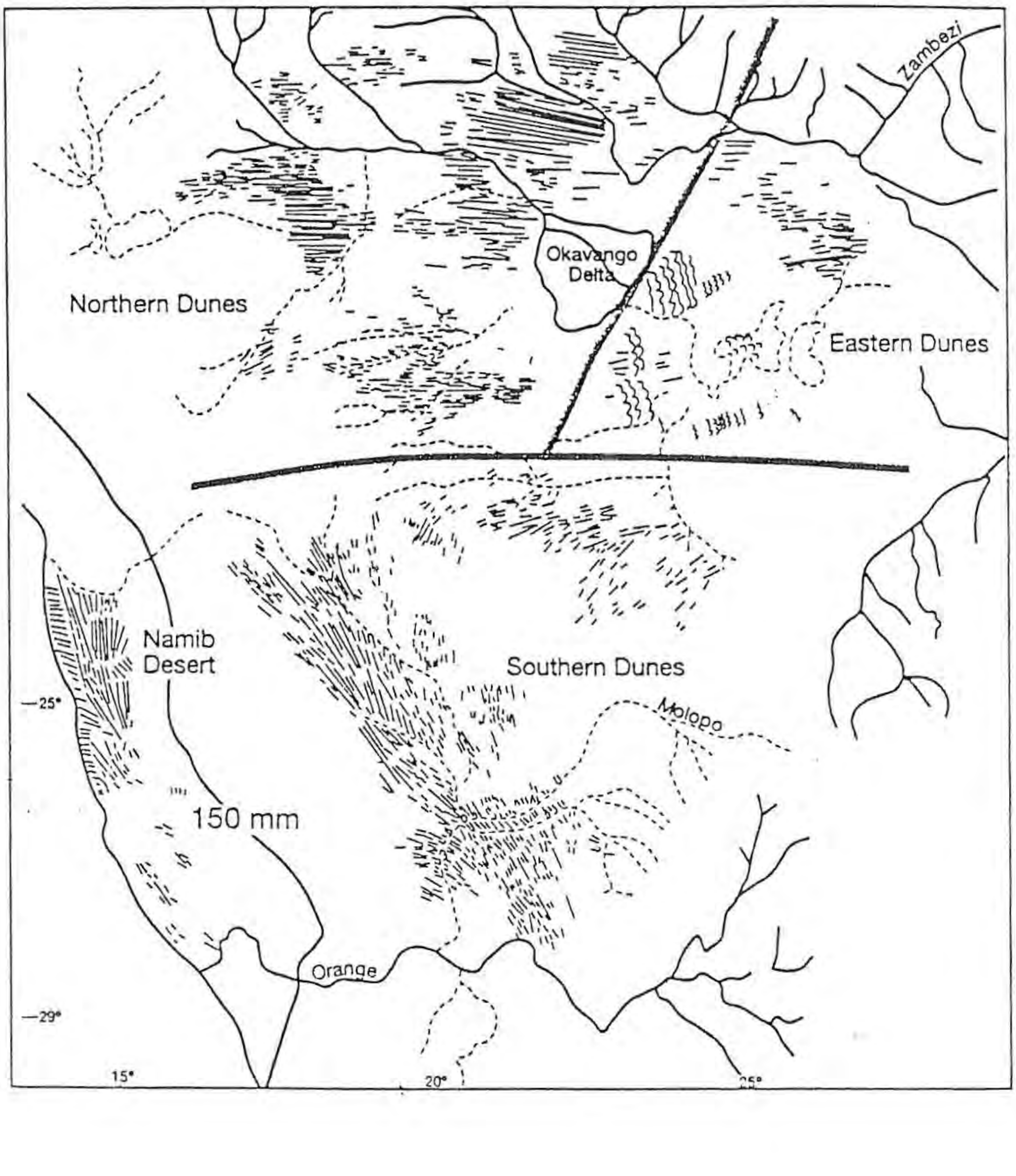


Figure 12. Dune systems in theKalahari (after Thomas, 1884 and Lancaster 1981).

The northern dunes consist of broad, rounded linear ridges on E-W to ENE-WSW alignments that extend to 13°S in Angola and may reach the Shaba plateau in the DRC at 11°S. In northern Botswana and Namibia, the dunes are up to 25 m high, with a spacing of 1500-2500 m, and some ridges extend for a distance of 200 km. As in the eastern group, savannah woodland dominates the dunes, with grassland in the interdune areas. Formation by easterly winds is indicated by deposition of sand against the eastern sides of the Aha and Tsodilo Hills and dune free areas down wind (Grove, 1969). The dunes are truncated by the Okavango system and also cut by other major rivers in this area. Degradation of the dunes increases northward, and ridges in western Zambia and Angola north of 14°S are discontinuous (Williams, 1982).

The southern dunes are the best-studied part of the system. They consist of 2-15 m high, straight to slightly sinuous, simple and compound, partly vegetated linear ridges on NNW-SSE to WNW-ESE alignments. Dune width is commonly 150-250 m with a spacing of 200-450 m (Lancaster, 1988), although some dunes are as much as 2000 m apart. Several varieties of linear dunes have been identified in the region (Goudie, 1970) (Figure 13). The most widespread are straight sub-parallel ridges 10-20 m high, with a spacing of 300-400 m, with rare to scattered Y-junctions that cover much of the central parts of the sand-sea (class 2 dunes of Bullard et al., 1995) and closely spaced (300 m apart), low (10 m high), narrow near dunes, with common Y-junctions, located south-east of the Molopo River valley (class 3 dunes of Bullard et al. 1995). Discontinuous and reticulate dune patterns occur toward the north and east of the dune field. In addition to the linear dunes, there are small areas of parabolic dunes that may represent a reactivation of the system (Eriksson et al. 1989).

The dunes occur in a 100-200 km wide belt that extends from the highlands of Namibia to the Orange River and rests on an extensive calcrete surface. Vegetation cover on the dunes consists of grasses, with scattered trees and shrubs, but the dune crests are commonly unvegetated and active sand movement takes place episodically. The amount of vegetation cover increases, with rainfall, to the north and north-east of the dune field. Process studies and climatic indices of dune mobility suggest these dunes are episodically active today (Wiggs et al., 1995).

The southern dunes are composed of well-to-moderately sorted medium-fine sand with a mean grain size of 170 – 340 μm. Dune crests tend to be slightly coarser, but better sorted, than adjacent dune bank and interdune areas. Overall the sand becomes finer and better sorted in a south easterly direction parallel to the net direction of sand transport (Lancaster, 1986). (Figures 14 &15).

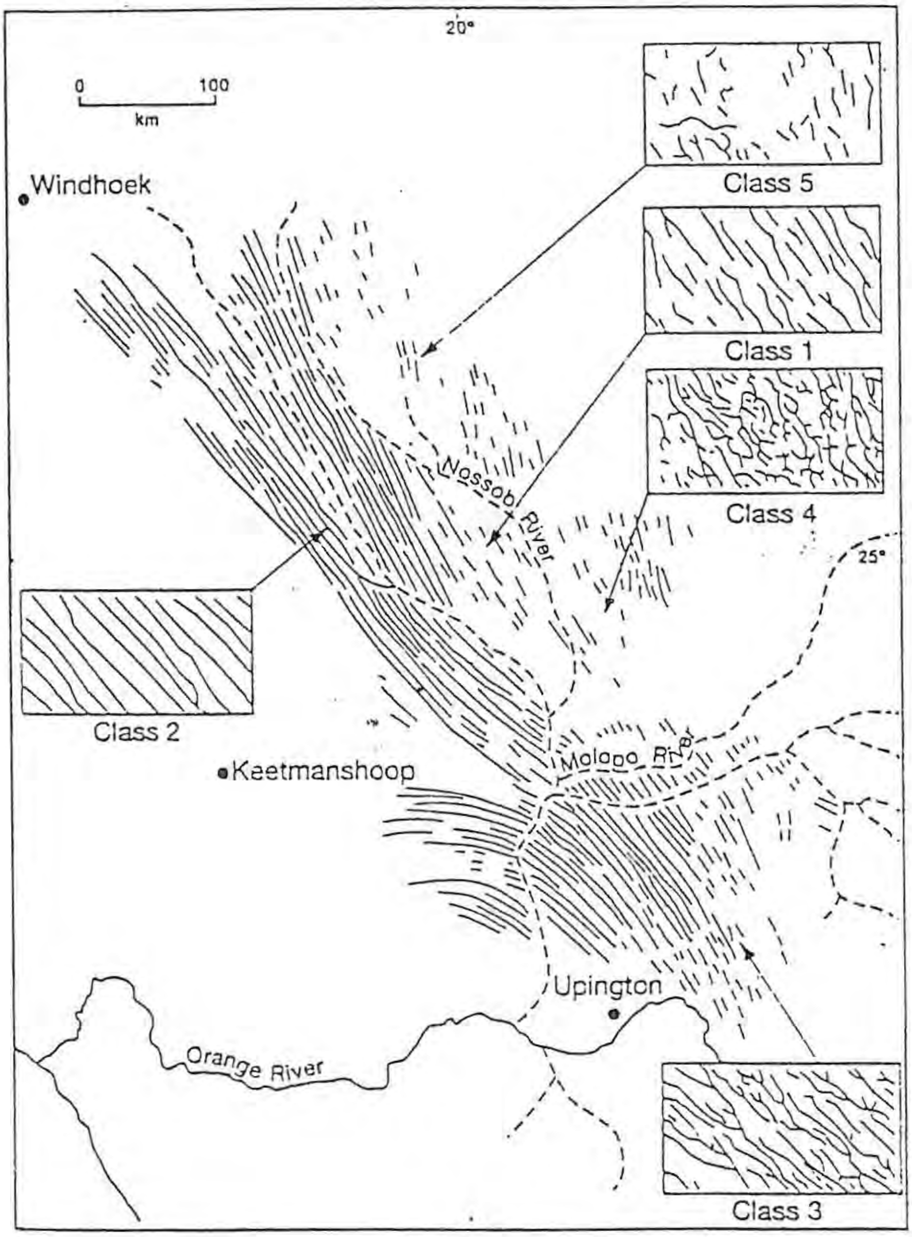


Figure 13. Dune systems in the south western Kalahari (after Lancaster, 1988 and Bullard et al., 1996)

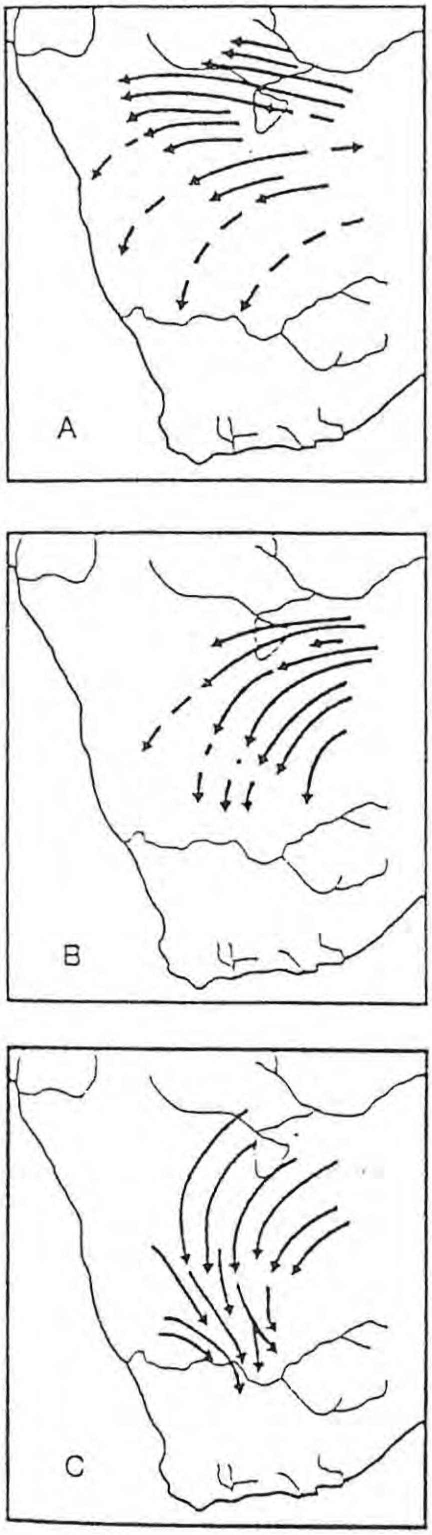


Fig. 14 Paleowinds inferred from the trends of relict dune systems in the Kalahari (after Lancaster, 1981).

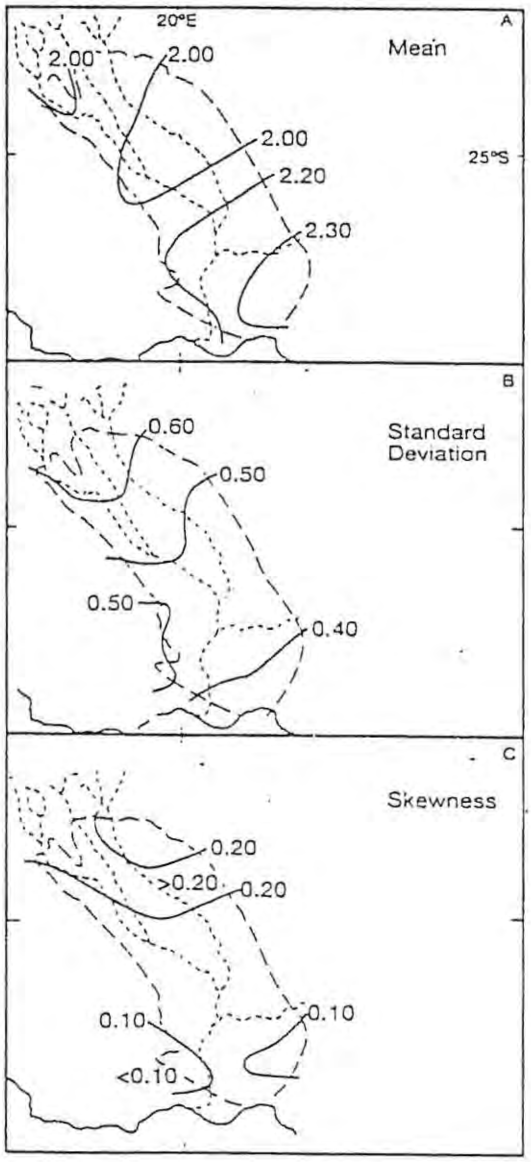


Fig. 15 Grain size and sorting variations in the south western Kalahari (after Lancaster, 1986a).

3.4.3 Mineralogy and Textures

Within Botswana the Kalahari Group sediments lie in two main depocentres separated by the northeast-southwest Ghanzi ridge (Thomas, 1988). In the south, the thickest deposits (up to 200 m) lie in the extreme southwest of Botswana, and extend into south-eastern Namibia. Thinner (generally < 100 m) sequences are present in south and central Botswana. Although the sedimentary succession of the Kalahari Group varies considerably across the basin, the upper part is largely composed of sands. These Kalahari Sands have characteristics such as grain surface textures which indicate aeolian deposition. However, it is often difficult to distinguish between wind and water lain units within the Kalahari (Thomas, 1987).

The earliest regional study of the Kalahari cover sands was by Poldervaart (1957), who focused on the distribution of the associated heavy mineral fraction. He concluded that the proportions of kyanite and staurolite decrease from east to west, while zircon proportions vary in the opposite sense. Tourmaline was a ubiquitous component of the heavy mineral suite, but did not show a clear distribution pattern.

Subsequently, Ballieul (1975) compiled an overview of the Kalahari sands by sampling along widely spaced roads and tracks. His analyses allowed a subdivision of the cover sands into four main types as discussed above in section 3.4.1. Ballieul concluded that some of the sands in Botswana were derived from local sources, originally by aeolian agencies, and subsequently modified by fluvial activity.

Thomas (1987) carried out a comprehensive analysis of Kalahari sands from most of central and northern Botswana, with the view to discriminating sedimentary environments. Control samples from known landforms were compared with material with no landform association. He concluded that discrimination was not possible by statistical means because aeolian processes have played a significant role in the development of grain and texture shape, and that processes responsible for non-aeolian landforms have not caused significant modification'. They have merely 'reworked the already wind-sorted sand' (Thomas, 1987 p.302).

Schlegel et al. (1989) reported regional grain-size and heavy-mineral studies over selected parts of the southern Kalahari sand cover. They concluded that their fineness, better sorting, and more positive skewness in comparison with what they termed 'mixed sands' could readily distinguish dune crest sands. The latter are predominantly fluvatile in nature, but contain some aeolian grains. This distinction is clearest in the southwest of the area studied by Schlegal et al. (1989), parallel to the longitudinal dunes of the Nosib and Auob valleys that stretch from Windhoek to Upington. Elsewhere in their survey areas, the distinction was less clear because of encroachment of the Okavango Delta sediments and erosion and degradation of the older Kalahari sands.

Moore and Dingle (1998) describe a helicopter-supported reconnaissance soil-sampling program conducted by Falconbridge in 1980 (Figure 16). The size of the area sampled was approximately 78 500 km^2, covering relatively inaccessible parts of the central Kalahari which had not been previously investigated. The results of this study show that there is not a consistent correlation between sand texture and sub-Kalahari geology, as envisaged by Ballieul (1975). While sub-Kalahari geology may therefore locally influence the texture of the cover sands, it does not satisfactorily explain many patterns apparent in the Kalahari (Moore and Dingle, 1998).

Over much of the Mega Kalahari there is unequivocal evidence from the presence of relict dunes of the dominance of aeolian processes during the deposition of the cover sands (Grove, 1969; Mallick et al., 1981; Lancaster, 1974; Thomas 1987). However, a number of observations from the study by Moore and Dingle (1998) suggest that this is not the case for many parts of the Kalahari within his study area. Moore and Dingle (1998) identified heavy mineral distribution patterns that were inconsistent with local prevailing wind patterns. Moore found that the textural and heavy-mineral distribution patterns were at variance with aeolian deposition controlled by prevailing wind directions. Rather, the patterns of deposition were dominated by fluvial influences such as sheet wash and ephemeral streams. A greater correspondence of sediment to underlying lithologies was more evident in the vicinity of the basin margins of the study area.

Basal Kalahari beds in southwestern and northern Namibia are inferred to be fluvial in origin (Mabbutt, 1955; Miller, 1992). The evidence for deposition of the cover sands in the area studied by ephemeral streams and sheetwash processes suggests that much of the Kalahari succession in Botswana may therefore be waterlain (Moore and Dingle, 1998). In other words, aeolian processes may have only resulted in a local secondary imprint on an essentially fluvial sedimentary sequence during recent arid episodes such as those inferred by Lancaster (1974).

Chatupa and Direng (2000) conducted a geochemical orientation survey in the nothwest Ngamiland that is extensively covered by the Kalahari sandveld. The most prominent geomorphological features of the region are the longitudinal rolling sand dunes and floodplain sediments adjacent to the unique, mid-continent Okavango Delta. Most of the sand dunes were deposited in the late Quarternary, and show evidence of both pluvial and arid environments.

The sandveld regolith comprises 70% quartz sand (-1000 + 63 μm) and 10-20% silt/clay (< 63 μm) material. There are no size variations between material from dune crests, dune depressions and down the soil profile to a depth of a metre. Boring animals (termites, ants, rats, etc.) have effectively homogenised the uppermost (0-1m) sandveld regolith. The most proficient being the termites, who build remarkable, architectural mounds that jut out of this essentially featureless terrain. Eroded relict mounds are also present in the interdune depression areas. Chatupa and Direng (2000) clearly identified termites as a significant agent for the mobilisation and dispersion of both heavy minerals and trace elements. The termites evidently follow the fluctuating groundwaters, and invariably transfer water and some of the deeper material to the surface. Termites have been observed to have gone down to horizons as deep as 70 m in the Jwaneng diamond field (Lock, 1985). Hence they have been considered to be one of the main bioturbation agents for transporting kimberlitic indicator minerals to the surface.

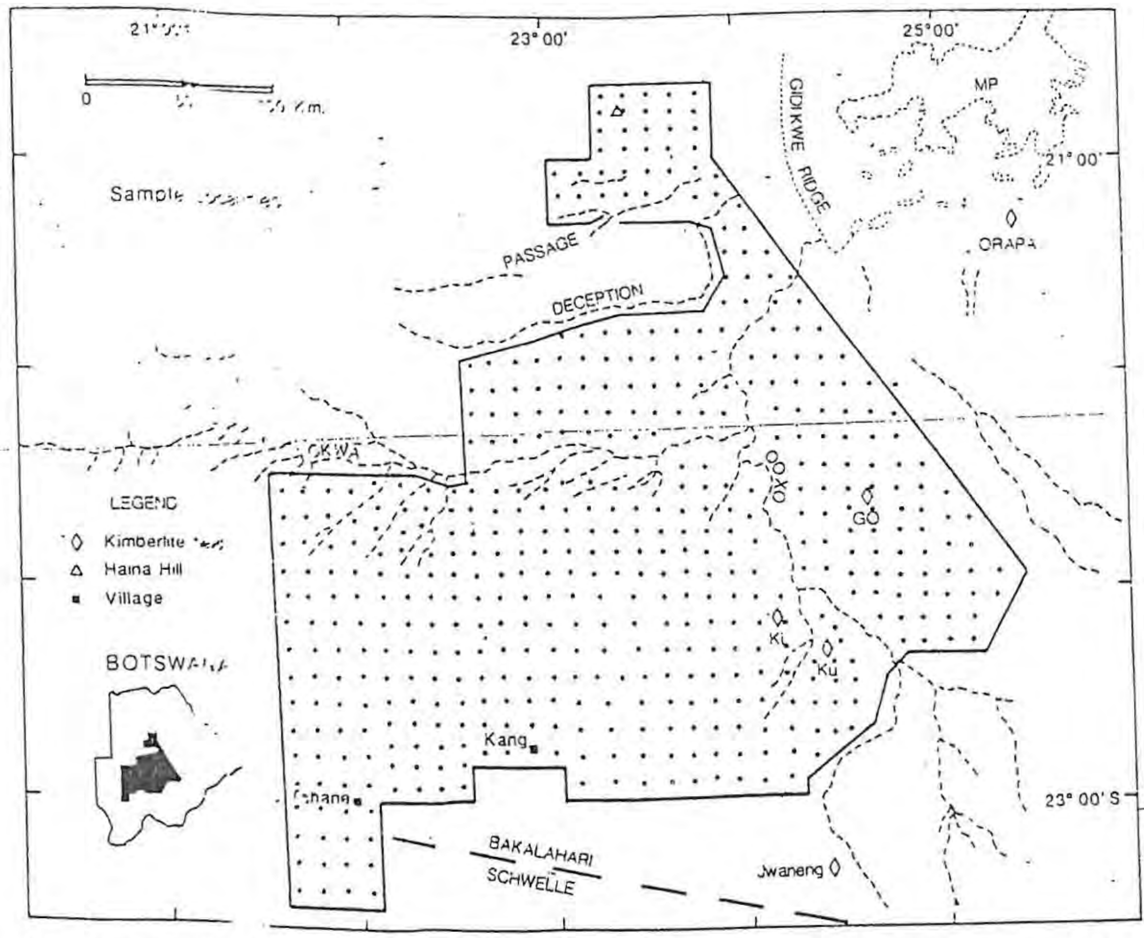


Figure 16 Sampling grid. Ki = Kikao kimberlites. Ku = Khutse kimberlites; MP = Makgadigadi Pans complex.

3.4.4 Paleoenvironments

Today, the northern and eastern dunes are situated in an area that receives 400 mm or more of rain per year, and are clearly relict forms. By contrast, the southern dunes are episodically active in an area where mean rainfall is 150-200 mm/yr. Wind energy in current conditions is also too low for widespread aeolian sediment transport in the current Kalahari region (Lancaster, 1998). Drier and windier climates are therefore implied for dune building periods in the Kalahari, but the effects of local sediment sources should not be discounted. The magnitude of the changes required for dune building and/or reworking in the Kalahari can be assessed using climatic indices of dune mobility (Lancaster, 1988). These indicate that reactivation of dunes in the southern part of the system requires a drop in rainfall of 50 % at present temperatures or a 20% increase in the percentage of time the wind is above threshold for sand transport. Much greater changes are required for reactivation of the northern and eastern dunes.

The atmospheric circulation patterns in the Kalahari region are dominated today by winds outblowing around two anticyclonic cells: the South African anticyclone situated over the eastern part of the sub-continent and the South Atlantic anticyclone off the west coast. Lancaster (1981) identified three paleo-circulation patterns based on the pattern of dune alignments (Figure 15): the northern dunes formed when anticyclonic circulation was much larger in radius; the eastern dunes formed during a period when the South African anticyclone was 2° N of its present position; and the southern dunes formed during a period when the year-round circulation resembled the modern October pattern, with a stronger South Atlantic anticyclone.

The strong SW-NE rainfall gradient in the Kalahari is a product of the contrast in sea-surface temperature (SST) between the Indian and South Atlantic oceans at the latitude of southern Africa (Stokes et al., 1997). Large differences of SSTs tend to result in an eastward movement of convective activity and dry conditions over the Kalahari, as a result of reduced transport of moisture from the western Indian Ocean, which is the source of most of the rain at the present time. Small differences in SST tend to increase convective activity and rainfall over the interior of the sub-continent. The control for the system appears to be the SST in the SE Atlantic. It can therefore be argued that periods of dune formation are the result of enhanced SST contrast between the SE Atlantic and the Indian Oceans. Stokes et al. (1988) argue that periods of dune formation correspond to millennial-scale cold sea-surface temperature events in the south east Atlantic. These events have been linked to sub-Milankovitch climate changes recognised in the northern hemisphere oceanic and cryosperic environmental archives covering the same period. The increased importance of the Atlantic as a moisture source during glacial times has been demonstrated by Stute and Talma (1998), whose study of dissolved noble gases in the Stampriet Aquifer of eastern Namibia revealed a 5°C temperature drop at the Last Glacial Maximum. Also evident from their analyses was an enrichment in ^{18}O in the water, compared with Holocene values. The authors attribute this anomaly to an increase in the contribution of nearby sources in the Atlantic to recharge of the aquifer during the last Glacial Maximum; this is consistent with the northward expansion of the westerlies at that time.

3.4.5 Characteristics of the Kalahari today

The area that is often referred to today as the Kalahari Desert and was included in Passarge's (1904) *Die Kalahari* is part of the Mega Kalahari approximately bounded by a line running from the Etosha Pan in Namibia to the Okavango Delta. This is also the area that has been the focus of concern in Quaternary paleoenvironmental studies during the last two decades since Grove (1969).

Given that it extends over more than 10 degrees of latitude, it is not surprising that the Kalahari Desert is climatically variable today, with a strong north to south-west mean annual precipitation gradient from approximately 800 mm to 150 mm and a concomitant increase in inter-annual rainfall variability. Although this has an influence on the composition of vegetation communities resulting in distinct spatial variations (Weare and Yalala, 1971), there are some notable environmental characteristics that prevail throughout the region and which are a direct result of the factors controlling the evolution and development of the area.

The constant altitude around 1000m.a.s.l. and low relative relief as well as the absence of any perennial and ephemeral drainage systems are quite striking within the Kalahari. Dry valley systems are found throughout the Kalahari, while the Middle Kalahari (Passarge's terminology) of northern Botswana is notable for its paleo-lakes. Together with the two other notable landforms suites of the Kalahari, the pan and dune systems these features provide potential sources of information for interpreting the Quaternary evolution of central southern Africa (Figure 17).

The Kalahari is also notable for its extensive cover of unconsolidated sand. At 2.5 million km^2, extending across the length and breadth of the Mega Kalahari, this represents the most extensive sand sea on earth. The sand possesses three major systems of linear (longitudinal) sand dunes that despite an absence of direct radiometric dates have been attributed to distinct periods of Pleistocene aridity (Lancaster, 1981) of varying spatial extent.

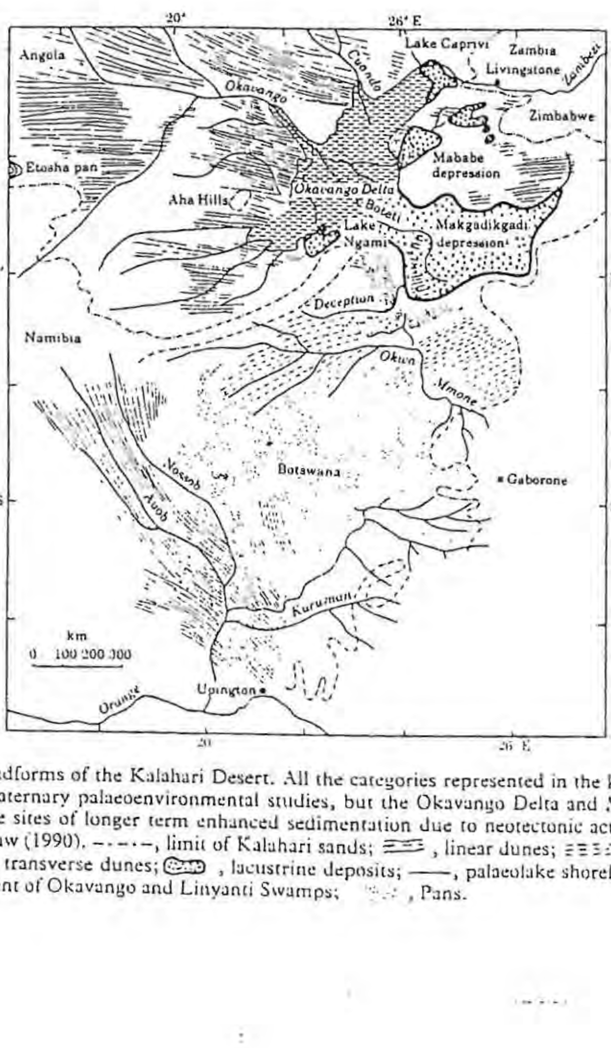


Figure 17 Landforms of the Kalahari Desert. All the categories represented in the key have been utilized in Quaternary palaeoenvironmental studies, but the Okavango Delta and Makgadikgadi Depression are sites of longer term enhanced sedimentation due to neotectonic activities. From Thomas & Shaw (1990). – - – - –, limit of Kalahari sands; , linear dunes; , feint linear dunes; , transverse dunes; , lacustrine deposits; ——, palaeolake shorelines; , maximum extent of Okavango and Linyanti Swamps; , Pans.

4. GEOLOGY OF THE MEGA KALAHARI

Despite the difficulties with chronostratigraphic interpretations and correlations it cannot be denied that there are widespread gross similarities and inter-regional equivalences in the lithostratigraphy of the Kalahari Beds from different areas of the Mega Kalahari (Thomas, 1988). This is illustrated in Figure 18 , which has been compiled based on information from borehole data and published sources (Thomas, 1988). These equivalences reflect deposition and erosion under environmentally similar conditions within the continental interior and on its flanks from which material was derived. There is, however, no evidence of temporal coincidence.

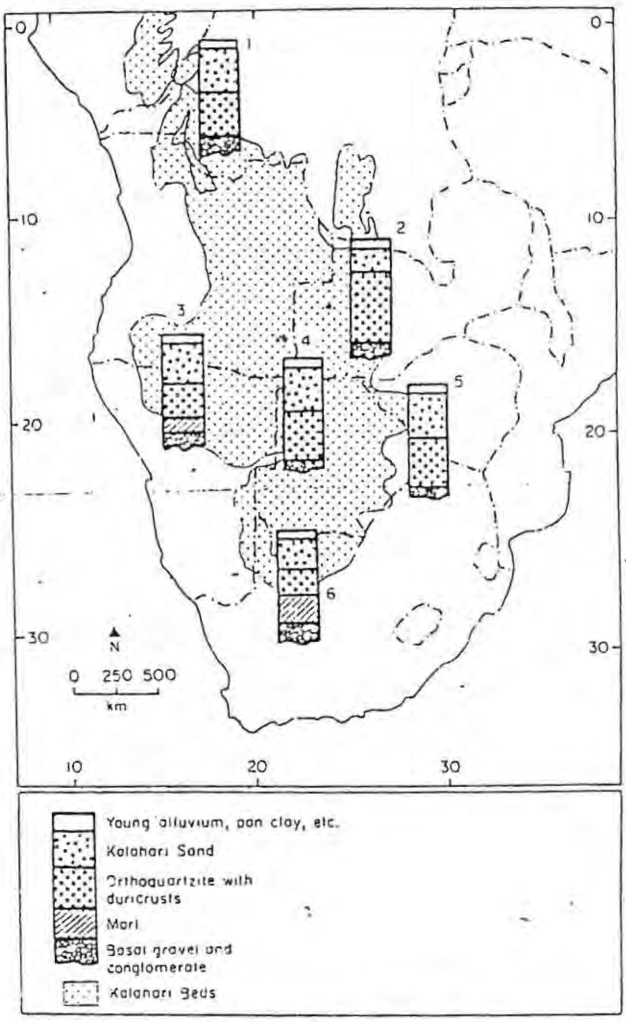


Figure 18 Generalised lithostratigraphy of the Kalahari Beds. 1, Zaire (Cahen & Lepersonne, 1952); 2, Zambia (Money, 1972); 3, northern Namibia (borehole data); 4, northern Botswana (Wright, 1978); 5, Zimbabwe (borehole data); 6, southern Botswana (Du Toit, 1954) and South Africa (borehole data). Schematic diagram only, not to scale.

It is also evident that, despite broad similarities, significant regional lithological variations exist both in the occurrence of different units and in their thickness (Figure 19). The problem of poor exposure and their extensiveness have not assisted in the establishment of a definitive stratigraphy. This section considers the general stratigraphic and lithological characteristics of the Mega Kalahari on a country-by-country basis. Mention of Angola, Gabon and the DRC are omitted due to the lack of published literature, particularly in English.

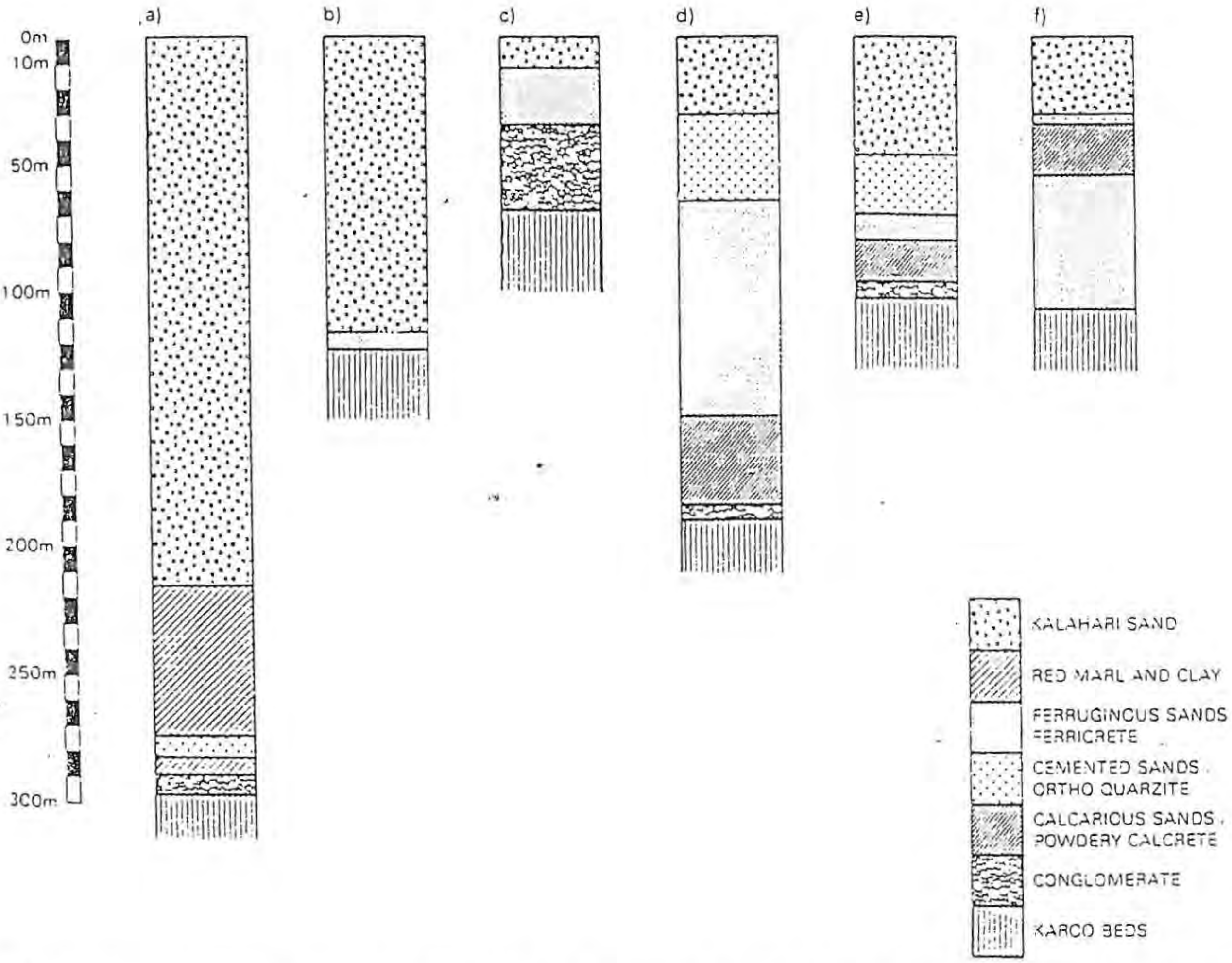


Fig. 19 Examples of Kalahari Group stratigraphy, from borehole records. A: East of Grootfontein, N. Namibia. B, C and D are from western Zambia. B: from the south-eastern area, C: from NE of Mongu and D: from the Zambezi plain. E and F are from western Zimbabwe. Note that the key is not a chronostratigraphic column.
Fig. 3. Illustration de la stratigraphie du Groupe Kalahari, d'après des resultats de sondages. A: A l'est de Grootfontein, Namibia du Nord. B, C et D en Zambie de l'Ouest: B: au Sud-Est, C: au Nord-Est de Mongu et D: de la plaine du Zambèze. E et F a l'Ouest du Zimbabwe. Noter: La légende n'est pas une colonne chronostratigraphique.

4.1 BOTSWANA

The Kalahari beds comprise the most aerially extensive geological unit in Botswana (Figure 20). It is also one of the most prospected, as it forms a relatively thin mantle over a number of kimberlite diatremes, some of which are highly diamondiferous. Their present-day distribution in Botswana probably indicates depositional models for much of the Mega Kalahari.

There have been few systematic studies of the Kalahari beds in Botswana. The unit represents a complex system of laterally impersistent facies and can be sub-divided only on a local scale. Furthermore deposition within a shallow, terrestrial basin during climatic fluctuations and tectonic upheavals would have resulted in a tendency towards successive reworking of the sediments (Coates et al., 1980).

4.1.1 General Lithologies

Certain workers have proposed partial successions. For example, McConnell (1959) suggests a central Kalahari succession commencing with the cemented Botletle beds of Passarge (1904), which then pass up to calcretes and lacustrine limestones ('Kalahari limestone, of Passarge) overlain in turn by 100 m of unconsolidated sands. In the southern Kalahari he recognised a similar sequence - starting with red marls, sandstones and conglomerates which are overlain in turn by 'Kalahari limestone' and then the sands. The following lithological summary is in part based on the account by Jones (1980).

Marls

Marls or clays with pink or red colours are described from the base of the Kalahari beds in the southern Kalahari basin. Although these deposits are laterally impersistent they can nevertheless form accumulations of up to 60 m in thickness. Green marls are associated with the fluvio-lacustrine sediments of the Okavango and Makgadikgadi.

Gravels

Gravels were reported by Rogers (1936) and have frequently been encountered during mining or drilling operations. These poorly sorted, structureless deposits in part represent infills to river valleys incised in the previous Kalahari surface. They are known to constitute important aquifers in the North Cape region of South Africa, although some are completely cemented (Gould et al., 1987). Gravels form the prominent fossil shoreline ridges of the Makgadikgadi Depression.

Aeolian sands

These form a conspicuous mantle over vast areas of the Kalahari landsurface. Typically they are reddish-brown or buff to white in colour and are fine grained. The surficial deposits are seldom thicker than about 30 m but they pass down to further unconsolidated sands with interlayers of duricrusts. At certain depths, individual sand grains become cemented by carbonate or silica to produce calcareous or quartzitic sandstone.

Dune features are present throughout the Kalahari basin. Although many are now degraded and most are vegetated, regional variations in dune orientation are still apparent (Figure 20). These were attributed by Mallick and others (1981) to two periods of aeolian activity, the first producing east-west trending dunes and the younger period giving rise to southerly or southwesterly-trending features. These are only the most recent wind directions to have been involved in sand distribution, however. Older sand deposits are recognised by Poldervaart (1955) and Baillieul (1975). Quartz is the main constituent with feldspar a persistent accessory. Kyanite has been recorded from the oldest sands in the northwest, whereas those in the southeast have a distinct heavy mineral assemblage of garnet, epidote, sillimanite, and andalusite suggesting derivation from the southeast.

Soil sampling undertaken for diamond exploration has shown that the various kimberlitic indicator minerals have migrated upwards through the superficial cover by a process of bioturbation (Lock 1985). Surface accumulations of these minerals are not always discrete anomalies over kimberlite pipes, however, and it is therefore probable that recent aeolian activity may have obscured the effects of upward bioturbation in certain areas.

Duricrusts

Duricrusts found in Botswana comprise calcretes, silcretes and ferricretes (respectively formed mainly of fine-grained low-magnesium calcite, silica, and iron oxides and hydroxides), and some intermediate types such as those formed by chalcedonic replacement of calcrete. Calcretes are widespread throughout the Kalahari and in other parts of Botswana. Silcretes are most commonly found only in the Kalahari beds. Ferricretes locally replace soils in the east of the country, outside the Kalahari beds outcrop, and so are not considered further here. Calcrete is the only widespread rock-type in Botswana that is often suitable for road construction. Its engineering properties have therefore been studied in some detail (Netterberg, 1980).

Generally speaking, calcretes can be separated into pedogenic and non-pedogenic varieties, although these seem to develop broadly similar textures. Pedogenic calcretes form by the progressive replacement of soil by calcite precipitated from vadose water. According to Goudie (1983), the formation of non-pedogenic calcretes can occur in drainage channels at the surface, or just above the water table (or even below it) thereby forming at depth if the substrate is sufficiently permeable. The suggested conditions for calcrete formation typically involve the interplay of biological and chemical pedogenic processes, groundwater conditions near surface and at depth, and rates of evapo-transpiration and capillary rise (Goudie 1983, Klapp 1983).

During the development of pedogenic calcrete, the soil first becomes calcified and is then gradually replaced in turn by powder calcrete, nodular calcrete, honeycomb or sheet calcrete, and ultimately by hardpan calcrete. These textures can all be observed in some mature calcrete horizons, in sequence downwards from a layer of hardpan calcrete. Thin laminar calcretes can occur on top of hardpans, although this may be typical of non-pedogenic calcretes formed just above the water table (Semeniuk and Meagher, 1981). Partial dissolution or disintegration of hardpan calcrete, either at of below the surface, leads to the formation of boulder calcrete and calcrete breccias. These may become re-cemented in due course, so forming complex textures (Netterberg, 1980). Little erosion may be required to remove the thin residue of soil overlying pedogenic calcretes, so partly exposed calcrete hardpans are common in the Kalahari. Complex calcrete profiles, including laminar forms, have been described in the Kalahari beds by Goudie (1983).

Three types of silcrete have been distinguished by Wopfner (1983); Types I and II formed respectively of crystalline and cryptocrystalline quartz and are associated with kaolinisation in a moist, acid environment. Type I silcretes are diagenetic, whereas type II silcretes involve pedogenic processes. Type III silcretes are composed mainly of cristobalitic-tridymitic

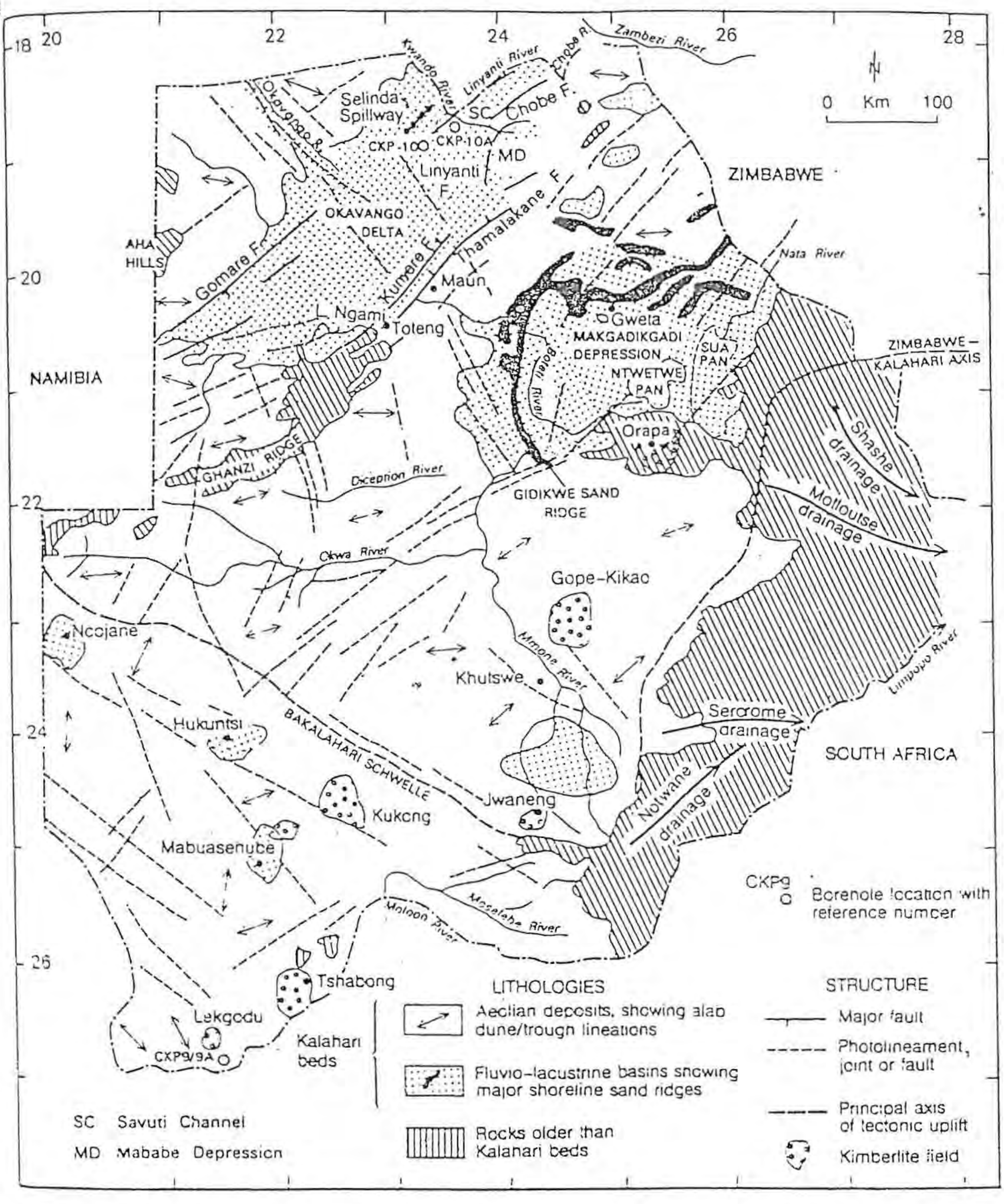


Figure 20 Distribution in Botswana of the Kalahari beds and of the major structures affecting them. Modified from Mallick and others (1981), Hutchins and others (1976), and Main (1987).

silica in arid, alkaline environments. Silcretes associated with pan deposits in Botswana are probably of Type III, but those elsewhere in the Kalahari beds could include examples of type II.

The Kalahari in Botswana has probably seen numerous periods of duricrust formation and replacement, linked with changes in climate and rate of deposition of sediment. In some areas, duricrust formation may continue at the present day. Figure 21 depicts a typical duricrust sequence from the Kalahari plateau.

Pans

Within Botswana some pans lie within dry valleys, in interdune hollows, or adjacent to rocky outcrops, but most seem to be placed at random. Regionally there is a concentration of pans on the 'Bakalahari Schwelle' (Figure 20), reflecting a broad area of very local surface drainage between the Molopo and Makgadikgadi drainage basins. The pans are flooded only after local rain but may then retain surface water for many months. Even then, the water table may remain accessible to hand-dug wells. Pans are thus extremely important to the ecology and economy of the Kalahari.

Various explanations have been advanced for the formation and distribution of pans but while no single model seems to explain all cases, their location can be related to the hydrology of their sub-strata, and in so many instances marks previous drainage channels or water bearing zones within the bedrock or the Kalahari beds. Such places would have developed ephemeral pools of surface water, leading to vegetation removal and, as the pools dried out, exposure of the soil to aeolian deflation.

Typically pans are circular or ovoid and range from a few tens of metres up to a few kilometres in diameter, but there are a few much larger examples. The pan surface is level, and can be grassy, muddy, or silty or stony, and not infrequently has a saline crust. The typical 'Kalahari pan' has a fringing lunette dune or dune complex on its western or southern margin, composed mainly of fine-grained material blown from the pan surface, partly stabilised by calcrete (Mallick et al., 1981).

Pan deposits commonly consist of grey-green clays or dark grey clayey sands (Jones, 1980). Studies by Farr and others (1980) show that in clays from the Mogatse Pan (Cenral Kalahari, Botswana) montmorillonite, quartz and dolomite are major constituents, with minor sepidolite and calcite and traces of halite and illite.

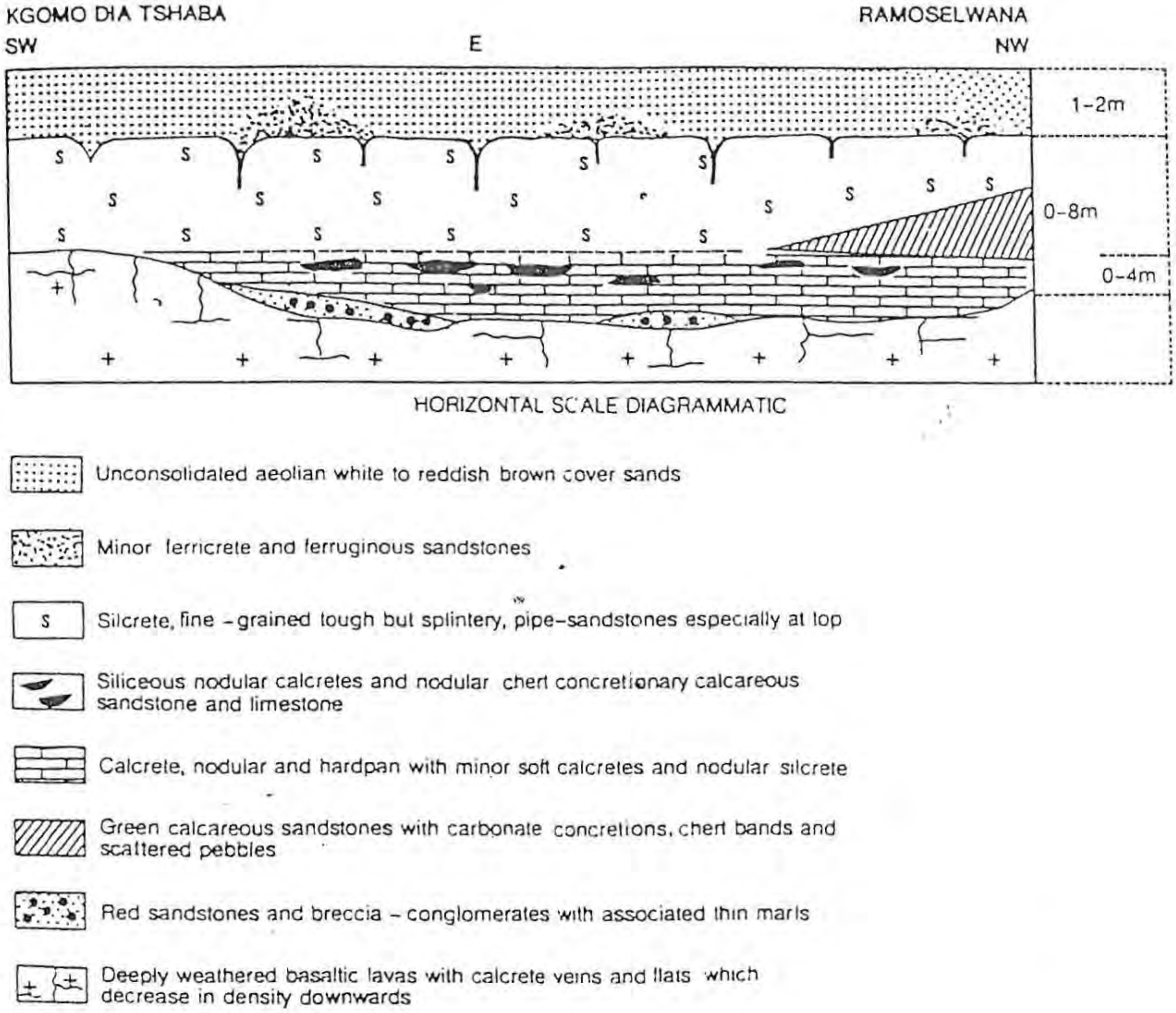


Figure 21. A duricrust sequence from the Kalahari Plateau surface in the vicinity of th Serorome drainage system. From Williamson in prep.

4.1.2 The Makgadikgadi Depression and the Okavango Delta

In Northern Botswana, the Kalahari is dominated by a complex system of inland drainage basins. Water draining east from the Angolan highlands reaches Botswana in the Okavango River which, most unusually for a river of its size, terminates in a huge inland delta, the Okavango Delta (Figure 20). A similar but much smaller system lies just to the northeast, where the Kwando River drains into a delta at the head of the Linyati River. Most of the water entering the Okavango Delta leaves the system either by evapo-transpiration or by drainage into aquifers, but ephemeral surface outflow can pass southwest to Lake Ngami, south east into the Boteti River and the western Makgadikgadi, or northeast via the Selinda Spillway into the Linyanti Delta. Depending on relative flood levels in the two inland deltas, water can apparently flow in either direction in the Selinda channel. In past times the Kwando would also have flowed via the Savuti channel southeast into a lake in the Mababe Depression, but it now mostly drains northeast via the Linyati and Chobe Rivers into the Upper Zambezi. In the eastern Makgadikgadi, Sua Pan regularly receives inflow via the Nata River from the northeast. Otherwise modern inundation of Lake Ngami and the Makgadikgadi Pans is probably more likely to be the result of local rainfall than inflow from the major rivers.

There is little information about the sedimentary sequences deposited in these drainage basins. Geophysical studies and deep boreholes have shown that they contain the thickest successions of the Kalahari beds in Botswana, and that they have had a complex history, involving both fluvatile and lacustrine deposition (Meixner and Peart, 1984). Present day morphology suggests however, that, whereas the Makgadikgadi Depression is a former lake basin, the Okavango Delta is now a fluvially-dominated complex bounded by a northeasterly-trending rift system.

In the Makgadikgadi Depression, former levels of the Greater Lake Makgadikgadi are indicated by shoreline gravel or sand ridges, some with cobble beaches, which lie at elevations of 912, 920, 936, 940 and 945 m above mean sea level. At its greatest extent, the lake may have covered and area of between 60 000 and 80 000 km^2 (Dingle et al., 1983), including the Makgadikgadi Pans themselves, most of the Boteti Valley, including Lake Ngami, the Madebe Depression and a third of the Okavango Delta. The volume of water then held by the lake has been calculated at 40 to 50 km^3, which after taking account of losses through evaporation, is many times that which could have been sustained by the present outflow from the combined Okavango and Kwando Rivers. It would have required the additional ingress of the Upper Zambezi River (Thomas and Shaw, 1988, Main 1987).

The Okavango, Kwando and Zambezi rives are thought to have originally flowed through what is now the Kalahari to join either the Limpopo or Orange rivers (Bone 1963). The Greater Lake Makgadikgadi came into existence as a result of ponding of this drainage to the north and west of the Bakalahari Schwelle and Zimbabawe-Kalahari axis of Miocene and Plio-Pleistocene uplift and faulting (Figure 20).

The thickness of sediment deposited in the resulting basin, or complex of basins, indicates the amount of subsidence that has taken place. The Kalahari beds could be about 200 m in thickness to the northeast of Gweta (Jones, 1980), suggesting a considerable former extension of the Makgadikgadi basin across what is now the Zimbabwe border (Figure 20). This is supported by the 160 m thick Kalahari beds that have been intersected in a borehole near Gweta. Beneath a 20 m-thick calcretised capping, the post-Karoo deposits consist of alternations of friable dark grey-green (when wet) calcareous muddy siltstones and similarly dark laminated marls (Carney and Dowsett, 1991). XRD analysis shows that the main constituent of these sediments is calcium carbonate. Quartz is a secondary component, with minor feldspar, micas, amphibole, dolomite and possibly sepiolite. Unfortunately core preservation was too poor to show sedimentary structures diagnostic of any particular regime, but overall the mineralogical assemblage is similar to that of lacustrine deposits from Sua Pan. These contain, in addition to sands, considerable amounts of clay rich in amorphous siliceous gel material and some chalcedonic bands. The detrital components (quartz, feldspar, mica, amphibole) represent aeolian additions subsequently reworked, or were introduced during fluvial incursions into the subsiding lake basin.

Brine deposits, which are currently being exploited commercially, occur in Sua Pan (Figure 19). Investigations since 1961, summarised by Stansfield (1974), have shown that large quantities of concentrated brines mainly occur at depth within a system of sandstone aquifers separated by clayey layers. Principal salts found at Sua Pan are common salt, soda ash, salt cake, muriate of potash, bromine and lithium chloride: brine analysis shows a preponderance of alkalies over calcium and magnesium (Gould, 1986). The brines are thought to have originated through the evaporative concentration of groundwaters that had previously dissolved considerable amounts of salt during their passage through the Karoo strata.

The Okavango Delta lies in an alluvium-filled graben some 400 km long within Botswana, extending for at least a further 450 km northeastwards into the Kafue Flats of Zambia (Mallick et al, 1981). The faults bounding the Okavango Delta have been delineated by both geophysical and photogeological methods (Hutchins et. al. 1986). They form well-defined scarps and have probably been active throughout the Pleistocene. Concentrations of seismic foci within the Okavango graben show that it is still actively subsiding (Reeves, 1972). A seismic survey by Scholz (1975) showed that during 1974 the Maun-Toteng area was the most active, although a number of events also occurred in the Mababe and Chobe areas. Focal mechanisms for the Maun-Toteng area were consistent with processes of normal faulting along a nodal plane dipping at 60° to the northwest.

Fault movements in the Okavango graben have apparently been responsible for the drying out of Lake Makgadikgadi. Disruption of drainage was brought about in two ways. The initiation of the Thamalakane and Kunyere faults, throwing down to the west, had the effect of damming the southeasterly flow of the rivers feeding the Makgadikgadi. Faulting along the line of the Linyanti and Chobe Rivers then deflected the Kwando and Upper Zambezi northeaswards where they were captured by the Middle Zambezi in the deepening Gwembe (Kariba) Trough. The increased flow combined with a rapid lowering of base levels caused the formation of the Victoria Falls nick-point (Bong, 1975).

Seismic refraction studies by Greenwood and Carruthers (1973) suggested a downthrow of about 115 m along the Thamalakane Fault and a thickness of about 300 m for the Okavango Delta sediments northwest of the Kunyere Fault (Figure 20). This estimate has been confirmed by drilling, which proved beds extending to 307 m in boreholes CKP-10A and 218 m in CKP-10, situated respectively to the east and west of the Linyanti Fault (Figure 20). Brief descriptions of these sequences by Meixner and Peart (1984) are summarised here, together with further observations.

In borehole CKP-10A the Okavango deposits overlie Karoo strata and commence with several metres of unconsolidated quartz sands which are dark brown and highly calcareous at the base. Overlying these is a sequence in which green calcareous mudstones or marls, some with sand and silt, form beds several meters thick. They alternate with pale grey quartz-rich calcareous sandstones. Calcrete appears near the top of the hole, although core recovery was too poor to determine its relationship to other sedimentary types. In CKP-10, poorly-consolidated quartz sands contain several black to grey-green marl or sandy marl intercalations.

Marl layers from 192 m in CKP-10 and from 250-277 m in CKP-10A were investigated by XRD techniques as part of the study. The results showed that the main component is amorphous material, which may be siliceous. There are also significant quantities of sepiolite, calcite and quartz, and minor amounts of tridyite, micas, amphibole, feldspar and smectite. In broad terms this mineralogy resembles that described above for pan deposits and could be compatible with the marls having originally accumulated in near surface alkaline lacustrine environments.

These analyses support the conclusion of Meixner and Peart (1984) that for much of their vertical thickness, the Okavango deposits are predominantly lacustrine, although containing fluviatile, sandy intervals, consistent with changes in depositional regime following channel migration and intermittent movement on the various bounding faults. Marls are less extensive in CKP-10, on the up thrown western side of the Linyanti Fault, than in CKP-10A, east of the fault (Figure 20).

The two boreholes also contain different basal facies. In borehole CKP-10, the Kalahari beds are underlain by basement augen gneiss which are brecciated, with calcrete cement in the upper 8 m, and which are overlain by massive calcrete. By contrast, in CKP-10A Meixner and Peart (1984) suggest that Karoo strata are sharply overlain by calcareous sands, with no duricrust development. These differences further emphasize the persistence of activity on the major faults bounding the Okavango.

4.2 ZAMBIA

Most of the information on the geology of western Zambia was taken from research conducted by N.J. Money in the Records of the Zambian Geological Survey, Volume 12 (1972). Money was responsible for mapping most of western Zambia and his study included the area lying between latitudes 13°00'S and 17°45'S and longitudes 22°00'E and 25°00'E (Figure 22). Much of the information on the lithologies within western Zambia was obtained from detailed logging of a series of oil exploration boreholes that were drilled within western Zambia. These holes are depicted in Figure 23. An idealised geological column of the Kalahari overlying the Karoo basalts showing the different formations present within these units in western Zambia as described by Money (1972) is given in Figure 24.

4.2.1 Geomorphology and vegetation

The Kalahari in western Zambia is a vast, sand-covered, semi-arid plain, which slopes gently from approximately 1200m in the north-north west to 900m above mean sea level in the southeast.

To the west and south-west, linear north-east trending dunes are present and extend into Namibia. Wide, hummocky and marshy flood plains, sand dune complexes, with river-scarp ridges and pans provide the major physiographic features in an otherwise monotonously flat landscape. These features are less than 100m above the plain level. The Zambezi River together with numerous sizeable tributaries is a well-developed, well-entrenched drainage network.

The vegetation varies considerably from north to south across this region The northern third is characterised by N-W-trending watershed plains consisting of riverine flood plains, floodplain grasslands (Liuwa Plain), and inland deltas. The remaining area consists of varying proportions of Munga woodlands, Miombo woodlands, Teak and Mopane woodlands, Kalahari woodlands and Kalahari savanna dry grassland plains. There is a dominant trending E-W Kalahari grassland plain feature made up of the Mulonga and Matabele Plains which begins at the Kwando River just inside Angola and extends across to the main Zambezi channel flood plain just south of Kalongola.

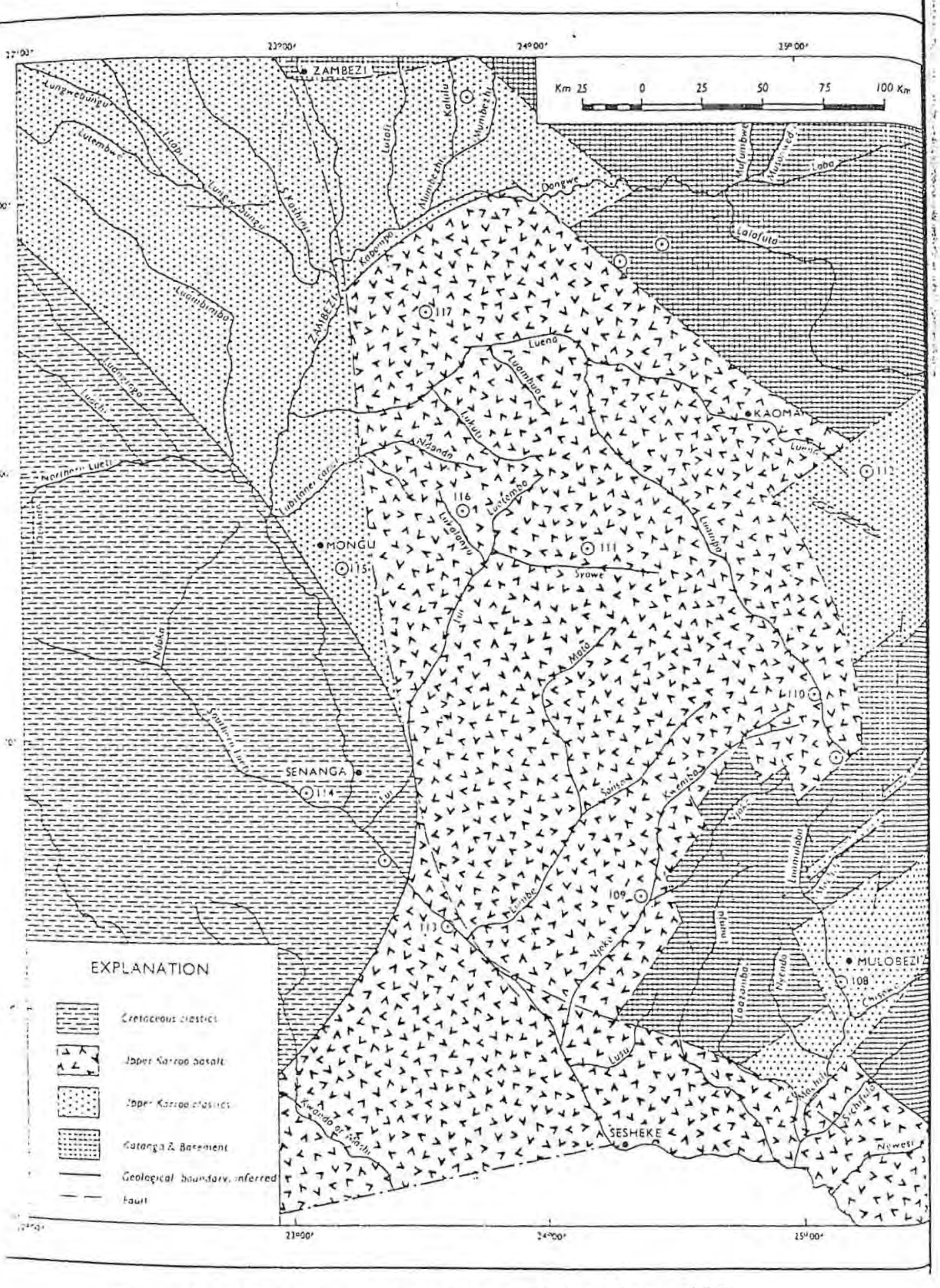


Figure 22. Geological map western Zambia (after Money, 1972)

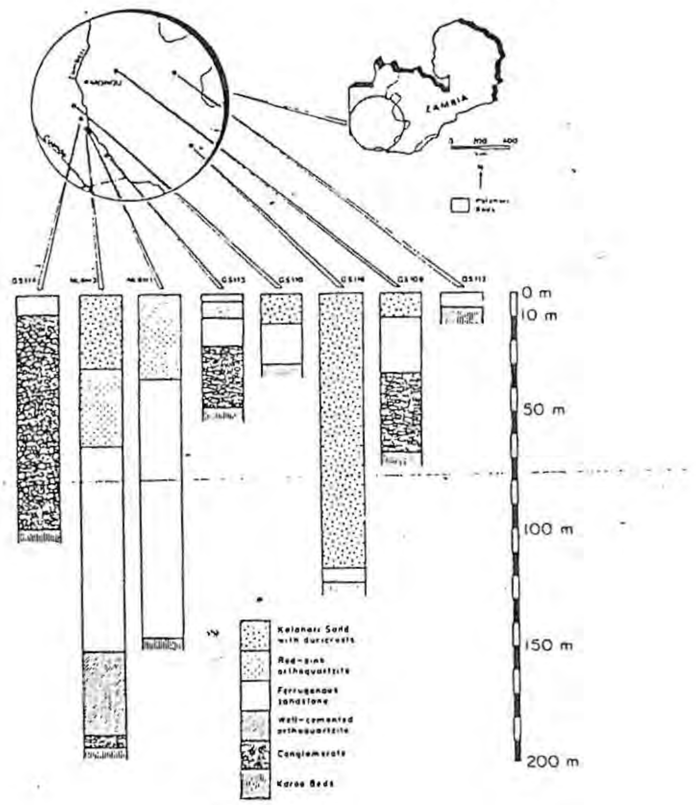


Figure 23. Variation in Kalahari bed Stratigraphy in Zambia after Money (1972)

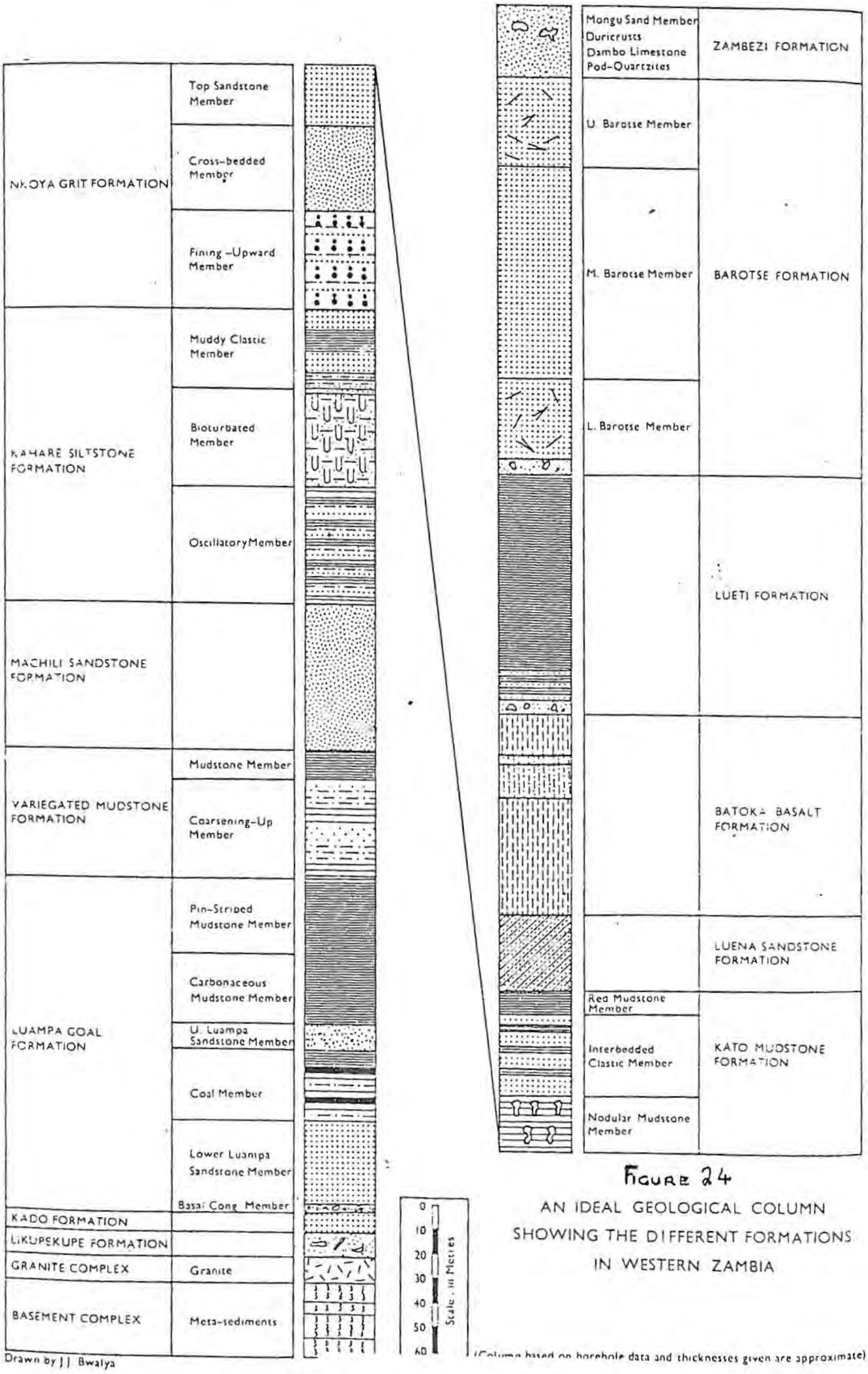


Figure 24

AN IDEAL GEOLOGICAL COLUMN SHOWING THE DIFFERENT FORMATIONS IN WESTERN ZAMBIA

4.2.2 Kalahari Geology

The Kalahari sediments of Western Zambia are sub-divided into the sandstones and quartzites of the Barotse Formation and the unconsolidated sands of the overlying Zambezi Formation (Table 1).

Information on the thickness of the Kalahari in western Zambia is limited and suggests that the thickness is highly variable. Data from boreholes discussed in Money's report and the De Beer's monthly reports were the only sources of information available. This information on the thickness of the Kalahari west of the Zambezi River is illustrated in Figure 23.

Barotse Formation

The Barotse Formation consists of a series of sandstones and quartzites and has been sub-divided into Lower, Middle and Upper Barotse Members. The type of sequence is the core from drill borehole GS113, where the succession is over 50 m thick; reference sections for different parts of the sequence crop out along the Zambezi River between Sioma and Sesheke. Similar Tertiary-Quaternary quartzite-sandstone sequences in Botswana have been described by Boocock and van Straten (1962); in Zimbabwe by Frost (1972) and Swift (1961); in Zaire by Cahen and Lepersonne (1952) and de Ploey, Lepersonne and Stoops (1968); and from various localities in southern Africa by du Toit (1966), Haughton (1969) and Rogers (1936).

Lower Barotse Member

The Lower Barotse Member rests unconformably on rocks ranging from Precambrian to Cretaceous in age. At Lusu (16° 50' S, 23° 50' E) and in borehole GS113, the lower beds include a basal conglomerate containing basalt and agate fragments and rest unconformably on the Batoka Basalt Formation, but at Nangweshi (14° 21' S, 23° 15' E) they overlie Cretaceous sediments.

Table 1. Kalahari Stratigraphy in Western Zambia (after Money 1972)

	ZAMBEZI FORMATIONS
	Pod Quartzites
	Duricrusts
	Dambo Limestone
	Pan Clays
	Evaporites
	Flood-plain Sediments
	River Gravels
	Sheei Conglomerates
	River Alluvium
	Other Recent Deposits
	Mongu Sand Member
KALAHARI GROUP	**BAROTSE FORMATION**
Western Zambia	***Upper Barotse Member***
	Quartzites
	Sandstones
	Pseudo-Conglomerate Beds
	Middle Barotse Member
	Ferruginous Quartzites
	Ferruginous Sandstones
	Lower Barotse Member
	Bedded Chert
	Silicified Limestone
	Sandstone
	Quartzites
	Basalt/Agate Conglomerate

The Lower Barotse Member is made up predominantly of fine to medium grained sandstones ranging in colour from pale olive to yellow-green. Isolated units of well cemented orthoquartzite are present. The quartzites tend to overlie other sandstone beds and generally have a gradational lower contact. Sand grains show moderate to good rounding and a bimodal size distribution.

Some beds, e.g. at Lusu Hill, contain elongate, pipe-like, cylindrical cavities varying in diameter from a few millimetres to 5 cm which tend to coalesce towards the base. A horizon of similar lithologic character and stratigraphic position in the Livingstone area was described by Dixey (1941, 1944) who named these beds Pipe Sandstone. Similar sandstone-quartzite beds in Botswana and Zaire have been described respectively as the Botletle Beds by Passarge (1904) and as the Gres Polymorphes by Cahen and Leppersonne (1952). The quartzites and sandstones are usually cemented by opaline silica and less commonly by chert, chalcedony, calcite and zeolite minerals. A two-metre thick sequence of bedded chert was intersected at the top of the Lower Barotse Member in GS 113; elsewhere, as at Ilwendo Village (17° 08' S, 24° 02' E), silicified limestone caps the Lower Barotse Member.

Middle Barotse Member

The Middle Barotse Member, which is subdivided into the Ferruginous Sandstones overlain by the Ferrugenous Quartzites, crops out below Ngonye Falls (16° 40' S, 23° 40' E) at Sioma; similar quartzites and sandstones were intersected at in borehole GS 113. The lower unit is at a series of feebly consolidated, greenish-yellow, ferruginous, sandstone beds while the upper comprises, well cemented, rust-olive-brown, iron-rich quartzites. In both units the beds are medium-grained with well rounded sand grains and contain banded, elongated, iron concretions and small isolated pockets or 'pods' of loose friable, iron coated sand grains. Here, as in the Lower Barotse Member, the quartzite horizon overlies the sandstone unit. Both lower and upper units show well developed, steeply-dipping, cross-stratification, which is interpreted as aeolian in origin.

Upper Barotse Member

The Upper Barotse Member consists predominantly of fine-to-medium-grained pink quartzites with occasional sandstones. The massive quartzites are cemented by opaline and chalcedonic silica; they are feebly calcareous and exhibit 'welded bedding'. Feebly calcareous, pseudo-conglomeratic horizons characterise the upper part where silica seems to be replacing the calcareous cement. Wood and plant fragments, presumably representing fossil soil horizons, also occur towards the top. The valley sides of the major rivers in western Zambia are flanked by Barotse Formation rocks and, as it is not always easy to determine the exact stratigraphic position, they have therefore been mapped as undifferentiated Barotse Formation. However, the quartzite units nearly always overlie sandstone layers.

The genesis of the quartzites in the Barotse Formation is considered to be related to fluctuations in the water-table and climate. In most cases the sand was aeolian in origin as is clearly seen in the Middle Barotse Member where sand dunes have been cemented by silica to give ferrugeous sandstones and quartzites. Elsewhere both aeolian and fluvatile features are evident. It is suggested that the high alkalinity of the inland arid region resulted in leaching out of the silica which was subsequently precipitated as cement on reaching areas of flowing water where pH would have been generally below 7, to give orthoquartzites of the Barotse Formation. This process is thought to have been continuing since Cretaceous times. The sub-division of the Barotse Formation and the quartzite –sandstone relationship may be related to the climatic conditions, which prevailed since Tertiary times.

Zambezi Formation

The Zambezi Formation consists primarily of loose sand, called the Mongu Sand Member which thickens towards the west and south west. The formation also includes more recent deposits such as duricrusts, pod-quartzites, pan limestones, claystones, saline evaporites, river gravels, flood-plain sediments, sheet conglomerates and alluvium. The recent deposits are incorporated within the Zambezi Formation for convenience even though much of the Mongu Sand Member, the major unit, is probably mid-Tertiary to early Quaternary in age.

The Mongu Sand Member is essentially fine-grained and is texturally uniform vertically and laterally. The sand is dominated by quartz but contains small amounts of heavy minerals. It is moderately well sorted in upland areas and poorly sorted in valleys and depressions. Though primarily aeolian in origin, it has in places been reworked by water to produce polygenetic or mixed facies. Savory (1963) postulates a complex and multi-genetic origin for the surface sands, and suggests that the sands in the Sesheke District are largely products of late Tertiary erosion of the Upper Karoo sediments deposited in shallow seasonal basins and later reworked by wind action during arid periods. A similar complex history has been postulated for the surface sands of Zimbabwe (Swift 1961), the Kalahari region and Botswana (Poldervaart, 1955).

The duricrusts include ferricrete, calcrete and locally silcrete. Calcrete is generally found in basaltic and limestone-dolomite terrains, ferricrete in regions floored by sandstone and basic rocks, and silcrete over acidic rock types. Duricrusts are extensively developed in southern Africa and their geology has been described by King (1967). Duricrusts develop because of the seasonal nature of the rainfall and the intense capillary action which occurs during the long dry period. This results in surface enrichment in insoluble salts, which cement the surface and near-surface residual clastic material. The type of duricrust depends on the cementing medium and duricrusts are nearly always formed in-situ. Numerous, irregular-shaped, smooth-surfaced, pods of quartzite occur at or near the surface within the Mongu Sand Member, especially along water courses. They are isolated units and resemble flint nodules within chalk. The genesis of these pod-quartzites is probably related to local fluctuations of the pH and the hydrological conditions.

Thin units (generally < 2 m thick) of recent gastropod-bearing limestone crop out along some dambos and depressions. The beds are due to precipitation of inorganic and organic carbonate in shallow water bodies. They are most extensively developed in areas floored by basalt or carbonate rocks.

Poorly laminated, white to pale greenish-yellow clay-stone rich in organic (plant) matter and sand grains was intersected above the Karoo rocks and below the loose sand and gravel in GS 111A and GS112. This probably represents the winnowed-out, finer sediment that floors the pans and depressions. The presence of such fine-grained clay layers may account for the imperviousness of some pans and dambos. A well lithified, bedded chert horizon was recorded in the Mongu Canal and is considered to represent a silicified podzol soil layer.

In a borehole (GS 117) drilled 40km east of Lukulu, a feebly consolidated, pale orange, mottled clayey sandstone containing pebbles was encountered between the surface Mongu Sand Member and the underlying Barotse Formation beds. This unit is over 40 m thick and is thought to represent a fossil flood-plain deposit of recent times, probably related to the proto-Luena river.

Sheet conglomerates and river gravels have been recorded at a number of low-lying areas and along ancient river terraces. They vary from poorly cemented to well lithified units, the chief cementing materials being calcareous and ferruginous minerals. An artefact-bearing sheet conglomerate crops out by the falls at Sioma.

Saline evaporite horizons have been recorded from a number of pans west of the Zambezi, but are generally less that 2 cm thick and represent soluble salts precipitated by evaporation.

Thus deposits of continental origin form a veneer over the greater part of the solid geology of western Zambia. According to Money (1972) the deposits extend back to the Tertiary.or even to the Cretaceous. Dating of these beds is difficult due to the lack of fossils and also because such fossils as do occur are closely related to present day forms. Nearly all the major rivers in the region flow in wide valleys underlain by alluvium and often show two to three well-developed terraces. These and some of the associated gravels and some evaporates are Recent and are probably related to the present climatic regime. Table 2 is a summary of the stratigraphy for western Zambia modified after Money and includes possible ages for the Lueti Formation.

Table 2. Stratigraphic Sequence for Western Zambia, (modified after Money 1972)

65MA	Tertiary	Kalahari-System	Kalahari Group		Zambezi Formation
					Barotse Formation
140MA	Cretaceous				Lueti Formation
?					
210MA	Upper Jurassic	Karoo System	Upper	Stormberg series?	Batoka Basalt Formation
					Luena Sandstone Formation
					Kato Mudstone Formation
	Middle Jurassic	?		Beaufort Series?	Nkoya Grit Formation
					Kahare Siltstone Formation
					Machili Sandstone Formation
	Lower Jurassic		Lower	Ecca Series	Variegated Mudstone Formation
					Luampa Coal Formation
	Permian			Dwyka Series?	Kado Sandstone Formation
	Paleozoic – Precambrian	Basement System			Basement Complex

4.3 ZIMBABWE

The Kalahari in western Zimbabwe is bounded by the 17°45' and the 20°00S latitudes and the 25°30' and the 28°30E longitudes (Figure 25). The physiography of Zimbabawe can be summarised as consisting mainly of a central plateau, generally considered to be an extensive uplifted peneplain that averages an altitude of 1200 m and forms a watershed of the areas to the north, to the west and to the southeast. The Kalahari is lower than the central plateau and averages an altitude of 1100 m in the east, pitching gradually to the west to an altitude of 960 m. The general relief is gentle.

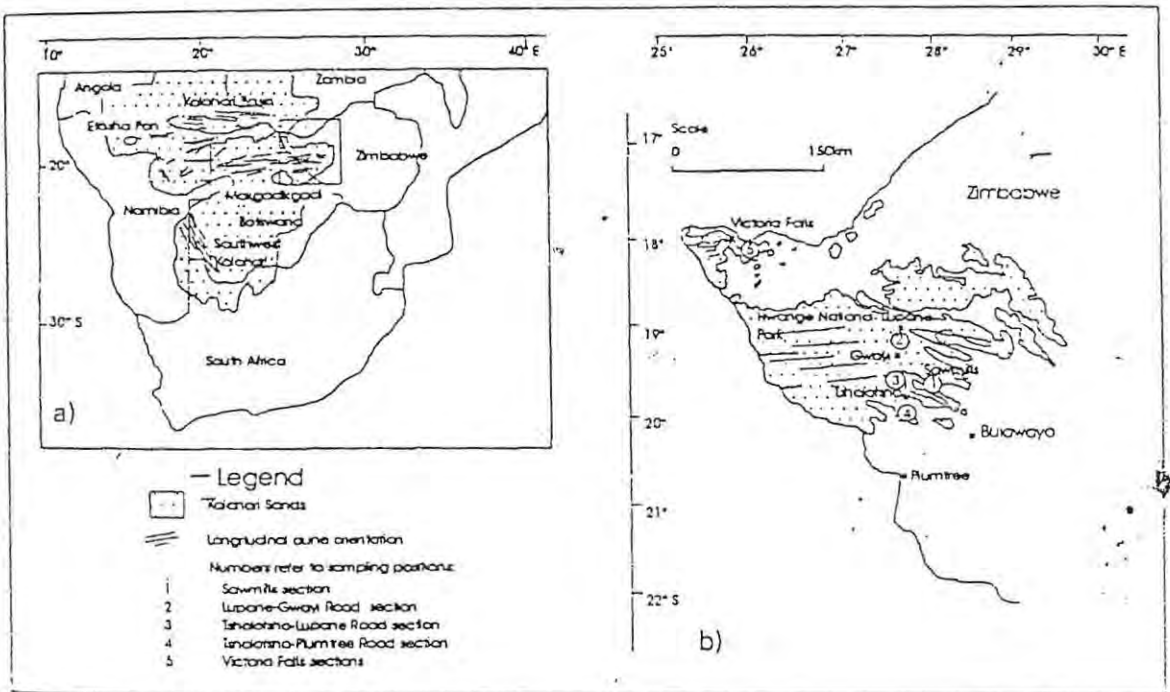


Figure 25. (a) Extent of the Kalahari sands in Southern Afrca. (b) extent of the sands in western Zimbabwe.

The Kalahari Beds in western Zimbabwe overly Karoo (Permian-Lower Jurassic) sandstones and basalts. In this area, the beds essentially comprise five major lithostratigraphical units (Lister, 1987). These are, from the bottom to the top: a silcrete unit, a sandstone unit (Pipe Sandstone), a unit of loose sands, ferricrete horizon and finally a surface unit of unindurated aeolian dune sands (the Kalahari Sands).

4.3.1 Kalahari Sands

In western Zimbabwe the sands cover an area over 40,000 km^2 (Stagman, 1978). They vary in thickness from a thin veneer at the margins to tens of meters in the central parts. Generally the sediments are loose, unconsolidated, homogenous sands with little compaction and no sedimentary structures of any form. The average grain size (190 μm) varies little throughout the sands, and the grain textures are generally consistent with those commonly associated with aeolian dune sediments. Mineralogically the sands are essentially comprised of quartz (> 95%). Feldspars and some heavy minerals constitute a minor component (Munyikwa, 1998). The sands range in colour from red to pale brown. Sands in basaltic terrain are deep red (e..g Victoria Falls and southern Hwange National Park) whilst sands overlying granitic basement (northern Hwange National Park) are pale brown. No palaesols have been documented and, apart from modern root penetration, there are no evident signs of bioturbation or reworking.

Surface expression is in the form of sand sheets and longitudinal dunes that in places exceed lengths of 100 km. Analysis of the dune alignment on LANDSAT imagery shows that the aeolian forms in western Zimbabwe are part of a much larger system of dunes that straddles western Zimbabwe, south-western Zambia, Botswana and eastern Namibia in an anti-clockwise whirl opening to the southeast. This orientation suggests that the dunes may have formed under the influence of winds blowing around the southern African anticyclone (Lancaster, 1981). Currently there are no active aeolian processes and the dunes support considerable vegetation, with the exception of the south-western Kalahari where some dune crests are being reworked.

In western Zimbabwe, Thomas (1983) classified the dunes in Hwange National Park as belonging to a different group from those in the Victoria Falls area. He also suggested that the dunes in Hwange National Park were younger than the dunes in the Victoria Falls area, basing his conclusion largely on the colour of the sands.

4.3.2 Ferricrete and Loose Red Sands

In some areas of western Zimbabwe, Kalahari Sands are underlain by a ferricrete horizon. This horizon varies in thickness from 20 cm to over 2 m. Its texture is largely nodular, but pisolitic varieties are not rare. The contact between the ferricrete and the Kalahari Sands is sharp and undulating. In places gravels, largely comprising ferruginous pebbles, occur instead of intact ferricretes, testifying to reworked ferricrete horizons. The origin of the ferricrete is not known with certainty, but Munyikwa (2000) has proposed two possible hypothesis. The first and oldest hypothesis postulates a post-depositional formation by leaching of Fe-oxides from the overlying sands to deposit them on top of a less permeable layer such as the Pipe Sandstone (Stagman, 1972). The second hypothesis suggests that the ferricrete developed as a tropical lateritic profile prior to the deposition of the overlying Kalahari Sands. After formation of the lateritic profile, an erosional phase would have ensued and would have removed the overlying material to end up with the resistive ferricrete or ferruginous gravel at the surface. This latter argument suggests a sharp unconformity within the Kalahari as well as a distinct change in climate from a sub-topical climate conducive to lateritic development to a semi-arid climate suitable for deposition of the Kalahari sands. There is little support elsewhere within the Mega-Kalahari to lithologies to support this.

Red unindurated sands (red loose sands) occur below the ferricrete. These sands do not occur in all places and where they do not occur the weathered Pipe Sandstone is directly overlain by the ferricrete or the Kalahari Sands. The sands lack any form of sedimentary structure and contain no fossils. At Victoria Falls, the sands have some pebbles of quartz and chalcedony scattered sparsely throughout and this distinguishes them from the overlying Kalahari Sands. The origin of the pebbles is difficult to ascertain, but earlier studies (Maufe, 1983) suggested that they originate from weathering of the underlying silcrete unit and were concentrated in the red loose sands by aeolian deflation. The grain size, grain shape and surface textures of the red loose sands are consistent with an aeolian origin (Munyikwa, 1998).

4.4 NAMIBIA

Documented information on the Kalahari sediments within Namibia region is gained from limited outcrop and scant borehole data from within the north east Kavango and Kaudom regions The following descriptions are based primarily on available literature, in the form of various published and unpublished papers and reports and some field observations.

Overall stratigraphic sequence in the NW Namibia (Table 3) is described as consisting of basal rocks of either Damaran or Archean age, followed by Karoo sequence sediments which are overlain and intruded by volcanics of Karoo age, consisting of plateau basalts & dolerite dykes and sills. The Karoo rocks are in turn overlain by late cretaceous, Kalahari group sediments.

Table 3. Sub-division of the Kalahari Beds, NE-Namibia after Albat

Aeolian sands	Panneveld calcrete, fossil dunes and redistributed windblown sand
Omatako formation	Ferricrete and ferruginous sandstone
	Unconformity
Eiseb Formation	Basal silcrete-alternating silcrete and grey carbonate units
	Unconformity
Tsumkwe Formation	Basal conglomerate, conglomeritic sandstone and sandstone with typical reddish brown colour
	Unconformity
Basement	Karoo, Damara, Nosib basement members

(Modified after Albat 1981)

4.4.1 Kalahari Stratigraphy

Kalahari group sediments are said to infill a broad basin which deepens towards the NW (Namibia Water Affairs report no. 2500/5/G2) with Kalahari isopachs showing a deep basin axis trending NW from Bushmanland to SW Kavango. The exact Kalahari thickness is however poorly constrained and Kalahari depth isopachs are mostly inferred from regional geophysics datasets and limited borehole data.

Albat (unpublished De Beers report) has observed that the composition and nature of the Kalahari sediments in a specific area are dependent on the rocks which comprise the local "paleo highs" with the results that the nature of the Kalahari sediments vary depending on the nature of the particular source rock. Albat has also reported thick clay development in paleo lows in pre-Kalahari surfaces.

The Kalahari sediments can be divided into 4 lithological units, namely the Tsumkwe, Eiseib and Omatako formations and overlying unconsolidated aeolian sands. Each unit is briefly described below.

Tsumkwe formation (Lower Kalahari)

The Tsumkwe formation is the basal member of the Kalahari beds. It is a sequence of red clays and mudstones that lie directly on the pre-Kalahari surface. A basal gravel or grit was identified, by Albat (unpublished De Beers report), in outcrop to the SW of Kavango area where the Tsumkwe formation rests unconformably on the older Karoo rocks. The clasts in this conglomerate are said to be poorly sorted and generally very angular containing a sandy matrix that is cemented with calcium carbonate. It has also been noted that clasts can be irregularly scattered throughout the Tsumkwe rocks or completely absent in places with the rock consisting only of cemented sand-size material. No bedding features are evident in the lower units although inclusions of cemented conglomerate and presence of irregularly scattered clasts have been noted by Albat. Finer grained sediments occur in the higher startigraphic levels and faint thin bedding is displayed. Upper levels are also reported as being more loosely cemented.

Albat (unpubl. De Beers report) has postulated that these sediments were deposited on land in very arid conditions with the clastic part of the basal Tsumkwe formation resulting from the rapid infill of fault-bounded depressions, partly as scree along slopes of Karoo topographic highs and partly transported over short distances by water during odd sheet flows and within minor drainages. The sand material is predominately windblown as many grains are frosted. The lack of bedding features and the presence of inclusions of cemented conglomerate suggest uplift while deposition was in progress. Carbonate cement owes its presence to arid conditions and to the availability of carbonate in the source region.

This depositional model proposed by Albat is based on the Tsumkwe formation sediments that outcrop in the SW of the Kavango region.

Eiseb formation

This rock unit follows unconformably (Albat, unpubl.) on the Tsumkwe formation. The cementing material is no longer of a reddish colour, typical of the Tsumkwe formation, but rather of a whitish-grey colour. The basal unit is comprised of densely packed sand-sized quartz grains, cemented together by clear slightly greenish silica. The upper-most Tsumkwe unit is described as a fine-grained Fe-rich, clay/mud-stone. No gravel or high energy zone was observed at the contact between the two formations.

Field observations of sandstone ridges (Eisieb) outcropping in the Omatako Omuramba indicate that on weathering the rock displays a brown colour due to silica replacement by Fe-oxide / hydroxide minerals. With the silica replaced, the rock becomes soft and friable. In outcrop the Eiseb unit often displays tube-like structures possibly related to bioturbation.

Overall the Eisieb sediments are partially calcretized, silcretized or occasionally ferruginized consisting mostly of alternating bands of thin, soft non-cemented sands and thicker, hard calcretized or silcretized sandstones. This bedding is more evident in the upper layers with the presence of white or cream coloured, calcareous sandstones (calcium carbonate cement) alternating with light reddish brown silcretes.

In summary the Eiseb formation consists of alternating silcretes and carbonate rocks clearly visible as light and dark reddish-brown banding during drilling.

Most sand-size material is inferred to be windblown although well rounded quartz pebbles (noted by Albat and identified in thin-section) would have been transported over fair distances within drainage channels. Albat postulates that inland seasonal lakes received abundant carbonate mud, which resulted in the deposition of cream-coloured limestones. A fluctuating climate in which periods of higher rainfall are followed by more arid climates has resulted in the observed calcrete – silcrete layering.

Petrography has identified a bimodal grain-size distribution, characterised by the presence of rounded to well rounded quartz grains as well as a population of angular, fine-grained material. A suggested geological setting for the deposition of these sediments would possibly be that of a lacustrine environment with some sediment input from an aeolian source.

Omatako formation

The Omatako formation rests unconformably (Albat, unpubl.) on the Eiseb formation. These rocks are mainly ferricrete and feruginous sandstone. The unconformity is marked by inclusions of silcrete within the ferricrete. It is of variable thickness and is completely absent in some boreholes. This ferricrete horizon outcrops within the Omatako Omarumba.

Field observations indicate that this horizon may have been produced by in-situ chemical alteration as opposed to representing a separate depositional sequence.

The above Kalahari sequence is largely similar to those described for the Kalahari formations observed in Botswana which describes the lower Kalahari as consisting of a series of units which indicate a transition from debris flow deposits through fluvio-lacustrine deposits to playa lake-type deposits. With the final stages of lower Kalahari deposition characterized by dehydration and evaporation resulting in calcrete and calcretized sandstone. This sequence is then overlain by cherts interbedded with fine-grained red silt stones and mudstones.

Aeolian Sands

Wind-blown sand overlies the above three formations. Large longitudinal fossil dunes are found through out the area and at present are covered by a thick vegetation of trees and shrubs. They are orientated WNW and are clearly evident on air and satellite photos. They appear to have a marked effect on current-day drainage patterns with tributaries of the Omatako River, pans and marshlands located along paleo dune troughs. The aeolian sands are interspersed with loamy, calcareous soil and crusts of ferricrete and calcrete.

4.4.2 Thickness and structure of the Kalahari in Namibia

The Kalahari beds in the region are flat-lying with regional dips generally less than 5 degrees (Albat, unpubl.) The maximum thickness of the sediments in the deepest part of the basin does not exceed 250-300m. The thickness of the aeolian sand does not exceed 50m below the fossil dunes (Water Affairs Report). The thicknesses of the individual members of the various formations are extremely variable from one exposure to the next. Albat infers some fault control during deposition with the floor of the Kalahari basin consisting of uplifted blocks and associated basins.

The dominant structural trend is NW, related to tension faults caused from East African Rifting and the failed arm of a short lived Gondwana spreading axis.

5. GEOMORPHIC EVOLUTION OF SOUTHERN AFRICA

Southern Africa is, in common with large tracts of Australia and South America, an area of old landscapes. The high fault-controlled margins of Africa, which formed the centrepiece of the Gondwanaland mosaic prior to continental separation, were rapidly attacked by weathering and erosion under the humid, tropical conditions of the Cretaceous. By the end of that period, up to three kilometres of material from the interior had been deposited on the continental shelf, leaving a vast, undulating plain, punctuated by occasional upland massifs (Partridge, 2000). The only continuous interruption was the step of the Great Escarpment, inherited from the time of continental rifting, which separates the interior plateau from the coastal margins. As climates deteriorated at the end of the Cretaceous, so this surface became armoured by duricrusts. It is only against this backdrop of high plains, forbidding escarpments, and rugged mountain massifs that the evolution and ultimate formation of inhospitable dune fields of the Mega Kalahari, can be properly considered and possibly understood.

Spectacular events played out since the Mesozoic. The tropical lushness of the Cretaceous, which saw dinosaurs roaming the length and breadth of the subcontinent as kilometres were swept from its surface by powerful rivers into the surrounding oceans, ended abruptly as comet impacts and massive volcanism brought to an end forever, the halcyon days of the Mesozoic (Partridge, 1998). Towards the end, explosive eruptions threw up clusters of volcanic cones; diamonds, which were carried to the surface in these kimberlites, were distributed through the network of Cretaceous rivers, some ultimately finding their way to the oceans. As climates dried and cooled during the Cainozoic and ocean circulations became established in relation to newly formed Antarctic landforms, so the first blanket of desert sand began to move inland from the west coast. The process of desertification was aided by the rise of the eastern parts of the country in two upheavals, the second mightier than the first. As the Tertiary drew to a close, changes in the mosaics of environments became even more profound. Woodlands opened up and grasslands spread over the high plains; new antelope species, adapted to open habitats, replaced earlier bush-loving lineages, and mans earliest ancestors took their first faltering steps across the veld (Partridge, 1998).

This history is unique to Africa, and is inextricably linked to the turbulent geological and climatic events which brought to a climax an almost continuous record of crustal evolution rooted deep within the Archean, 3.8 billion years ago.

This account begins with an overview of the key events in southern Africa's later geological history which exerted a major influence on the macro-scale geomorphology of the subcontinent.

5.1 PRE-CAINOZOIC EVENTS

Overall, current understanding suggests that, at a general level the Mega Kalahari is best interpreted through an understanding of the pre-and-post Gondwanaland evolution of the interior basin rather that from a sedimentological-stratigraphic perspective.

The geological evolution of southern Africa may be viewed as a series of accretionary events following the stabilisation of the Kaapvaal Craton (which forms the structural basement of the subcontinent) around 2600 Ma (de Wit, 1992) (Figure 26). Accretion occurred during a number of extensional and compressional periods; the most important, between 2000 and 1000 Ma, led to the addition of the Namaqua-Natal mobile belt and culminated in the stabilisation of the Kalahari Craton around 1000 Ma. On to this amalgam was imposed a series of orogenic belts, creating the "swells" between intervening cratonic "basins" in the course of the Pan-African tectonic cycle, which ended about 600 Ma years ago. As Burke (1996) has pointed out, this basin-and-swell structure is unique to the African continent; the swells were repeatedly rejuvenated and uplifted during the Phanerozoic in relation to the thick and mechanically strong intervening cratons. These recurrent movements have persisted into the Neogene, when uplift spread from the inter-cratonic mobile belts to affect large areas of the cratons themselves.

The gross geomorphology of southern and eastern Africa is dominated by the African Superswell (Nyblade and Robinson, 1994). This is a region of relatively elevated terrain which manifests itself both on the continent and on the ocean floor to the southwest (Figure 27). It commences in the Afar region of the Red Sea and extends southward incorporating the eastern and western branches of the East African Rift. South of Lake Victoria, the Superswell widens to include almost the entire sub-continent, where it is clearly delineated by the 1000 m contour. This broad southern section is asymmetrical with the eastern flank being of higher elevation. The terrain within the Superswell lies at an average elevation of at least 500 m higher than the equivalent crustal rocks elsewhere on the world's continents. The origin of the Superswell is linked to dynamic topography, caused by the presence of hot, low-density mantle near the core-mantle boundary beneath southern Africa, and to somewhat shallower hot mantle beneath the East African Rift.

Burke (1996) suggested that the Superswell began to develop about 30 Ma ago, when the African Plate ceased movement relative to the hot-spot reference frame. In contrast, Partridge (1998) has proposed a multi-stage uplift history for the Superswell.

In the early Paleozoic a passive margin developed along the southern edge of the Kalahari Craton. This formed the locus for the deposition of shelf sediments of the Cape Supergroup (de Wit, 1992). As shorelines migrated northwards, an active margin developed in the south, initiating the rise of the Cape Fold Mountains around the end of the Permian (a locality plan for this section is provided in Figure 28). The inter-continental Karoo Basin between the Cape Mountains and the Kalahari Craton extended well beyond the present margins of the subcontinent into adjoining areas of Gondwanaland; subsidiary depositories (such as the Botswana basin) developed around its northern and eastern margins. Karoo sedimentation, totalling 7000 m in places, culminated in extensive outpourings of basalt some 183 Ma ago as incipient rifting began to trough the later continents and India. When Karoo volcanism ceased, most of southern Africa south of 15° S was covered by Karoo strata. Only in the west along the developing coastline of Namibia, did widespread fissure volcanism occur later, with the extrusion of the 130 Ma Etendeka Basalt along the incipient Atlantic rift margin.

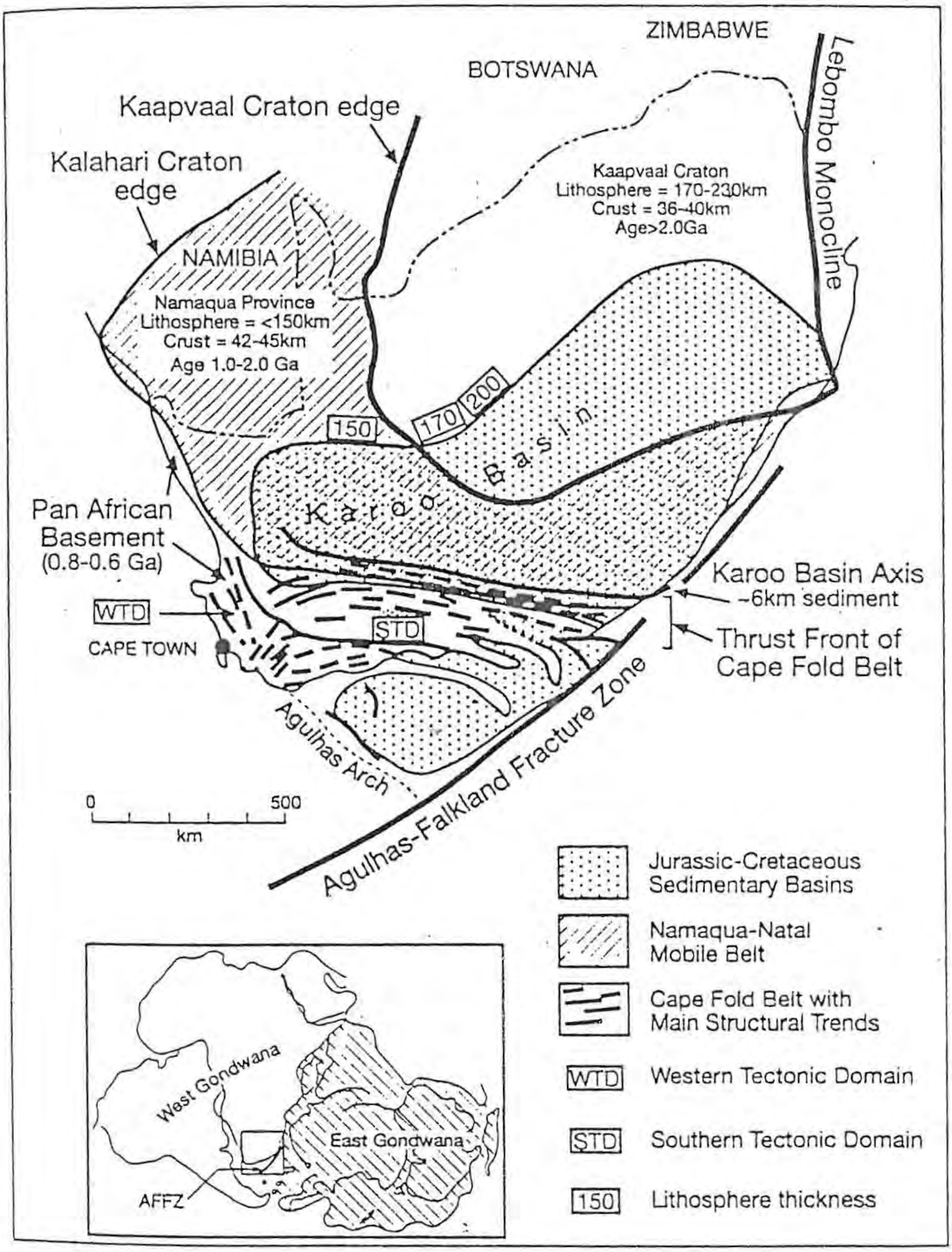


Figure 26. General tectonic framework of Southern Africa (after De Wit, 1992).

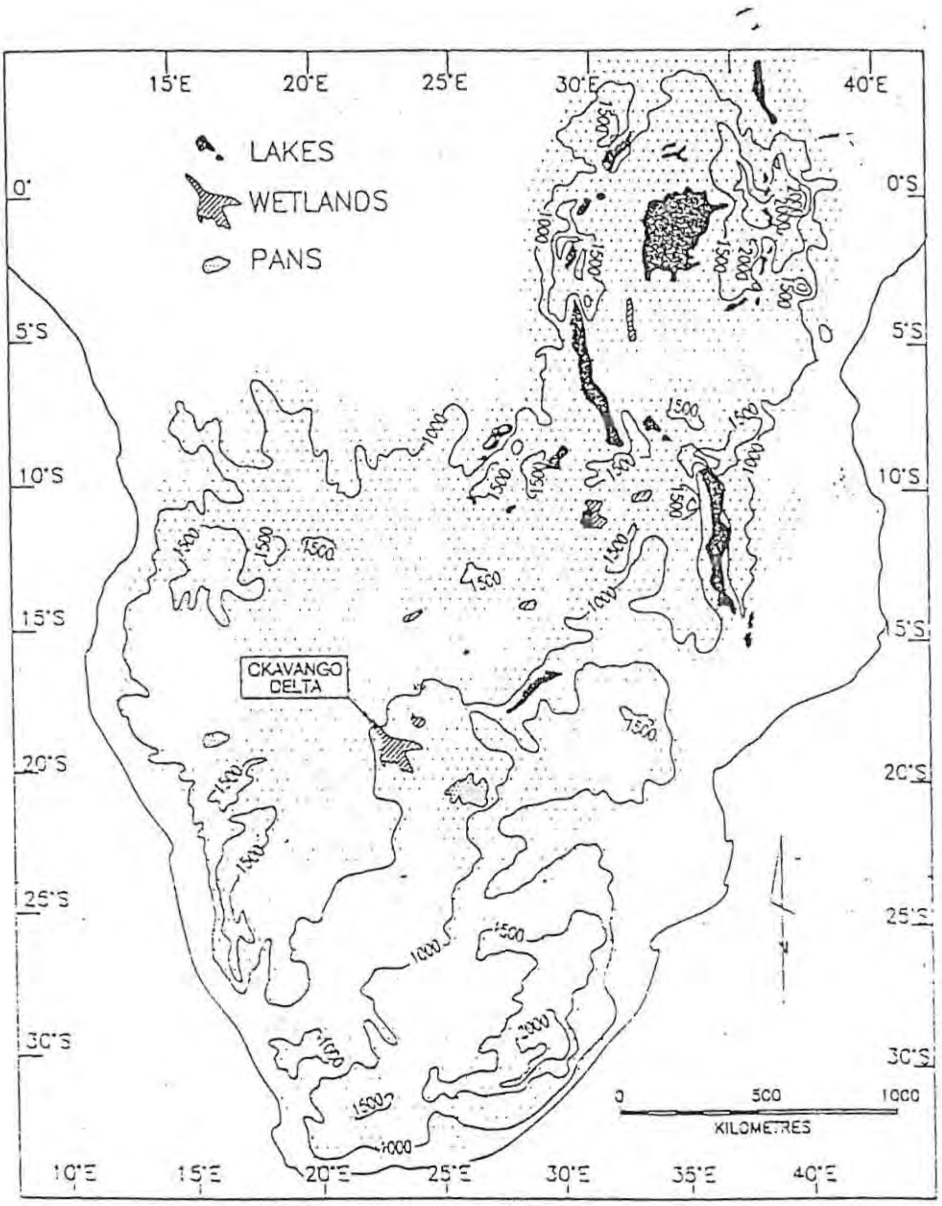


Figure 27. The distribution of the African Superswell on the continent (after Gumbricht and McCarthy, 2001)

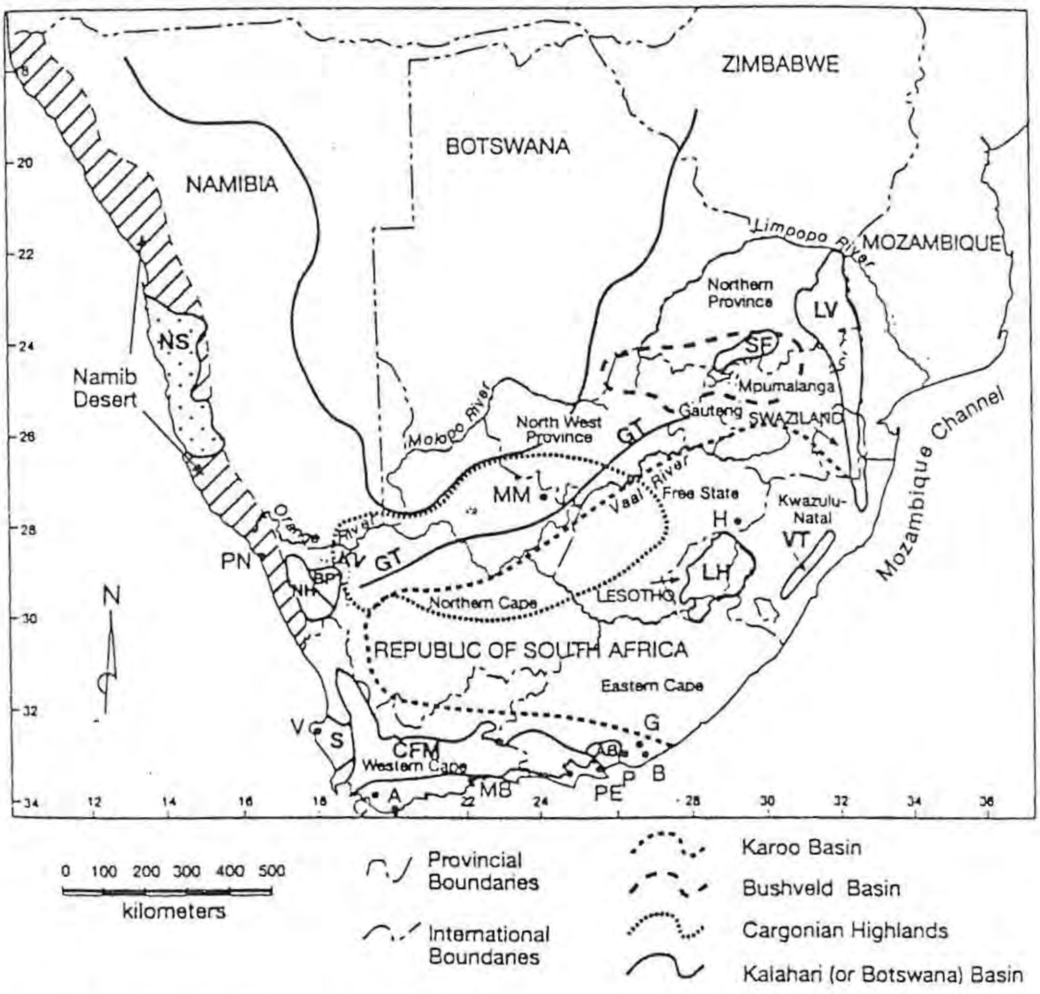


Fig. 28 Areas and localities referred to in this chapter. A-Agulhas; AB-Algoa Basin; B-Bathurst; C-Caledon; CFM-Cape Fold Mountains; G-Grahamstown; GT-Griqualand-Transvaal Axis; H-Harrismith; KV-Koa Valley; LH-Lesotho Highlands; LV-Lowveld; MB-Mossel Bay; MM-Mahura Muthla; NH-Namaqualand Highlands; NS-Namib Sand Sea; P-Paterson; PE-Port Elizabeth; PN-Port Nolloth; S-Swartland; SF-Springbok Flats; V-Vredendal; VT-Valley of a Thousand Hills.

The rifts along which Africa separated from South America in the west, and India and Antarctica the east were almost certainly associated with pre-existing Pan-African welts (Partridge 1998). As in the case of the much younger rift system of east Africa, uplift of the flanks preceded separation; the remains of these elevated rift shoulders are preserved in the cordon of high ground which is still present, especially inland of the Great Escarpment in Lesotho and the Eastern Cape. Based on evidence from kimberlite pipes, relatively small thicknesses of material have been removed from these areas by Cretaceous and Cenozoic erosion (Hawthorne, 1975). The precise dating of separation of Africa and South America remains unresolved, but most authorities place the initiation of drift between 129 and 121 Ma. (Fouche et al., 1992). On the east coast the occurrence of late Jurassic marine sediments in Tanzania and Mozambique shows that rifting had commenced there prior to about 140 Ma, while the occurrence of Lower Cretaceous, partly marine, Makanti formation on the inland margin of the coastal plan of KwaZulu-Natal farther south shows that separation had occurred by 130 Ma. Along the southern Cape coast the lowermost marine rocks of the Utenhage Group are of a comparable age, while on the west coast the first Lower Cretaceous marine sediments have an age of about 120 Ma (Dingle et al., 1983). Along both the east and west coasts continental separation appears to have occurred as a clean break with minimal shearing movements, but along the south eastern coast, the Falklands Plateau was detached from the Mozambique ridge along a right-lateral transcurrent fault. Local fault-bounded intermontane basins formed along this shear zone and began to fill with coarse bahada deposits inland of the position of the newly forming coastline. By about 100 Ma the Falkland Plateau had probably cleared the Agulhas bank and basin infilling was replaced by widespread epeirogenic sedimentation over the subsiding continental shelf (Dingle et al., 1983). The resulting marine Cretaceous sequences are of considerable thickness (in excess of 8 km in places) and support the findings of fission track analysis onshore (Brown et al., 1994) that, by the mid-Cretaceous, between one and three kilometres of material had been removed from the post-rifting surface of the subcontinent. In the process a great deal of the covering Karoo rocks was stripped from all but the main Karoo Basin, with the resulting imposition of some structural control by underlying, older strata.

All of this evidence points to the fact that, prior to the break-up of Gondwanaland, southern Africa stood high and possessed substantial relief. How high the continent actually stood is indicated by several lines of evidence: firstly, rates of terrigenous sedimentation on the continental shelf during the immediate post-rifting period (early Cretaceous) exceeded, by an order of magnitude those which characterised the Cainozoic (Dingle et al., 1983; Martin, 1987).

Only in some areas such as the Orange basin, were there departures from this general trend (Rust and Summerfield, 1990) which are probably the result of the enlargement of onshore catchments through piracy as the drainage net evolved. The bracketing of the bulk of onshore erosion and concomitant offshore sedimentation within the Cretaceous is independently confirmed by the preservation of the crater facies of kimberlitic diatremes in a number of localities extending from Bushmanland to the Kalahari basin, which have ages ranging from about 90 to 60 Ma.(Figure 29). This strongly indicates that, over considerable areas of the continental interior, little additional landscape lowering has taken place during the Cainozoic; indeed, the main interval of erosion was over by the beginning of the Upper Cretaceous.

Such a massive shedding of Cretaceous erosional detritus to the oceans would not have been possible without the existence of a high-standing landmass. Partridge et al. (1987) suggested that pre-rifting elevations probably ranged from about 2400m in Lesotho to about 1500 m in the western interior. Subsequent estimates do not contradict these conclusions (Brown et al., 1990; Rust and Summerfield, 1990). These high elevations were due, in part, to uplift along the rift shoulders and were responsible for the existence, from the time of continental separation, of a substantial margin escarpment. This precursor of today's Great Escarpment, which forms a great horseshoe rampart between 50 and 200 km inland of the coast (Figure 28), was driven back with vigour by early Cretaceous erosion. In part, the result of high energy potential provided by the elevated margin, and in part a response to the humid tropical climates of the Cretaceous, this prolonged interval of erosion dumped two to four kilometres of sediment on large areas of the continental shelf, with up to twice those thicknesses preserved in some basins off the south Coast (Dingle et al., 1983). Rates of sedimentation were not uniform throughout this period, and a hiatus shortly after the beginning of the Upper Cretaceous may be associated with an interval of extensional tectonics, with a concomitant marine regression; an important period of kimberlite emplacement around 86 Ma was almost certainly linked to these movements (Smith et al., 1984). While this and other Cretaceous events were of undoubted importance, they had little influence on the course of the cycle of denudation set in train by continental separation and the break-up of Gondwanaland.

Figure 29 Simplified map showing the distribution of land surfaces in southern Africa. Heavy dots show localities at which constraints on the age of the surfaces are available (range in Ma).

5.2 AFRICAN SURFACE

The result of this great cycle of erosion was profound. The marginal escarpment, which defined much of the southern African coast, receded rapidly and, on the evidence of studies of basalt pebble sizes in east coast rivers and in marine cretaceous deposits (Mathews, 1978), had reached a position within about 20 km of today's Great Escarpment by the end of the Cretaceous. Erosion to the new oceanic base level cut a gently sloping bench across the coastal hinterland as the escarpment receded. In the Eastern Cape, Upper Cretaceous marine deposits relating to the Campanian-Maas-Trichtian transgression are preserved in a hollow on this surface at Need's Camp Lower Quarry (Lock, 1973). A short distance away at the Upper Quarry, later marine strata record a subsequent Eocene transgression. Similar Eocene deposits near the mouth of the Great Fish River contain rolled silcrete pebbles, which, as will be demonstrated later, were derived from the erosion of an extensive duricrust capping which formed at the land surface at the beginning of the Cainozoic. In the southern Cape, near Swellandam, this silcrete caps an alnoite pipe, which is part of a cluster dated to around 63 Ma (Moore, 1979); and in the Spehrgebied of southern Namibia, Cretaceous marine strata of Santonian age are present on the local equivalent of the same surface.

It is evident that throughout this major interval of Cretaceous erosion, land surfaces on either side of the receding escarpment were formed under the influence of two separate base levels of erosion. In the coastal hinterland, sea-level constituted the ultimate control, but inland of the Great Escarpment, the base level for erosion was provided by major river systems such as the Orange and Limpopo at their point of egress from the interior plateau through the Great Escarpment. This created the unusual situation that land surfaces of essentially the same age were cut at different levels above and below the escarpment line. As in the case of its coastal equivalent, available evidence points unambiguously to the fact that the interior plateau surface had been formed by the Upper Cretaceous. The net result of this situation was the creation of two vast erosional levels, above and below the Great Escarpment. These enormous pediplains were grouped together as the African erosion surface by Lester King in a number of seminal papers and the nomeclature has been maintained by later authors such as Partridge. Above the African surface, a number of mountain massifs were preserved (figure 29 above), including the ranges of the Cape Fold Mountains, the Namaqualand Highlands, the mountains of the Eastern Cape and Lesotho, and the ranges inland of the Namibian and Angolan escarpments. Despite the controversy which has, at times, surrounded King's interpretations, especially on questions of whether the African Surface should be regarded as a peneplain, a pediplain or an etchplain (for review see Twidale, 1988), and whether it can be regarded as a single geomorphic entity, there is no longer any doubt concerning its existence and antiquity (Partridge, 2000). Evidence from the morphology of kimberlite pipes (Hawthorne, 1975) indicates that, even on the highest of these remnants in Lesotho, some 300 m of material has been eroded subsequent to the emplacement of these pipes around 90 Ma ago, indicating that all vestiges of the original Gondwanaland surface were removed by Cretaceous erosion.

In common with most continental areas, southern Africa appears to have enjoyed warm, humid climates during most of the Cretaceous. The Upper Cretaceous, in particular, seems to have been characterised by extensive tree cover on the basis of the abundance of logs and plant material preserved in the marine sequences; a large, fine terrigenous component in these sediments argues for high rainfall and deep weathering (Dingle et al., 1983). These conclusions are supported by evidence from the crater fills of kimberlite pipes in Botswana and Northern Cape, with their richly fossiliferous paludal sediments (Scholtz, 1985, Smith, 1986). There are suggestions, however, that by the end of the Cretaceous a degree of desiccation had occurred, resulting in the establishment of somewhat drier forest and shrub land communities than had characterised earlier periods. On the basis of recent work by De Wit et al. (1992), the remains of the dinosaur Kangnasaurus Coetzeei were found east of Springbok in the early years of this century in such crater sediments. In the intervening areas of the subcontinent, erosion has almost entirely removed all but a few remnants of terrestrial deposits of Cretaceous age. A notable exception to this are the lower beds of the Kalahari Supergroup in Botswana, most of which are likely to be Createceous in age. Unfortunately no diagnostic fossils have yet been forthcoming in these units.

5.3 CRETACEOUS DRAINAGE NET

It may be expected that a warm and wet Cetaceous would have left some legacy, however fragmentally preserved, of a well integrated drainage net and deep weathering of susceptible rocks beneath the African Surface; this is indeed the case. Over larger tracts in the North-West province, extending into the Northern Cape, the remains of an ancient drainage system are preserved as a series of sinuous gravel lags, following the highest points of the topography. In many cases these provide the last remaining evidence of the former presence of the African Surface; they have in fact acted as a protection from subsequent erosion, in the process creating an unusual inversion of topography whereby intact or slightly eroded remains of the original channel deposits now appear in the landscape as ridges.

Spatial changes in the drainage net of southern Africa were controlled by a combination of factors, including: (1) landscape denudation, partially exposing more resistant pre-Karoo topography with a marked structured fabric, and (2) continental uplift along the margins, and later in the eastern hinterland, of southern Africa (Partridge, 1998). The first manifestation of a major post-Karoo drainage system is recorded in the paleo-Limpopo delta in Mozambique. At that time a vast river system is believed to have eroded large quantities of material from south-central Africa, including central Angola, northern Namibia, Botswana, western Zambia, north-western South Africa and southern Zimbabwe. This system was active after the creation of the eastern coastal margin during the Jurassic and continued until the late Cretaceous.

Rifting of west Gondwanaland during the early Cretaceous initiated two river networks, one of which drained most of South Africa south of the Cargonian highlands, the other southern Botswana and southern Namibia (Figure 30). These two river systems have been referred to as the Karoo River and the Kalahari River (De Wit, 1993). This hypothesis is based on analyses by Dingle and Hendey (1984), which indicated that major drainages entered the Atlantic both via the present-day lower Orange (or Kalahari River) and the lower Olifants (or Karoo River). Behr (1989) has also suggested that the Cretaceous drainage in the south western Kalahari fed into the lower Orange valley. The watershed between these two ancient river systems now coincides with the Cargonian highlands (Visser, 1987) and the Griqualand-Transvaal Axis (du Toit, 1933). The Karoo river had its headwaters as far inland as the Ghaap Plateau and Lichtenburg, and drained most of the North West Province, the Free State and Lesotho. Remnants of this system have been preserved in the Kimberly area, but most have been eroded, especially in areas where these sediments were deposited on soft Karoo rocks. A major tributary from the north (the Trans-Tswana River) was part of the middle-Orange River system, although it is doubtful whether this rose as far north as central Africa. Capture of the Karoo River by the upper Kalahari River at the Dorinberg Hills near Prieska led to a major reorganisation of the drainage net to approximate to the pattern of today (de Wit, 1993). Late Cretaceous crater-lake deposits in Namaqualand (Smith, 1986) not only show that the climate was humid and tropical at the end of the Cretaceous, but also highlight the fact since late Cretaceous times very little erosion has occurred over most of South Africa.

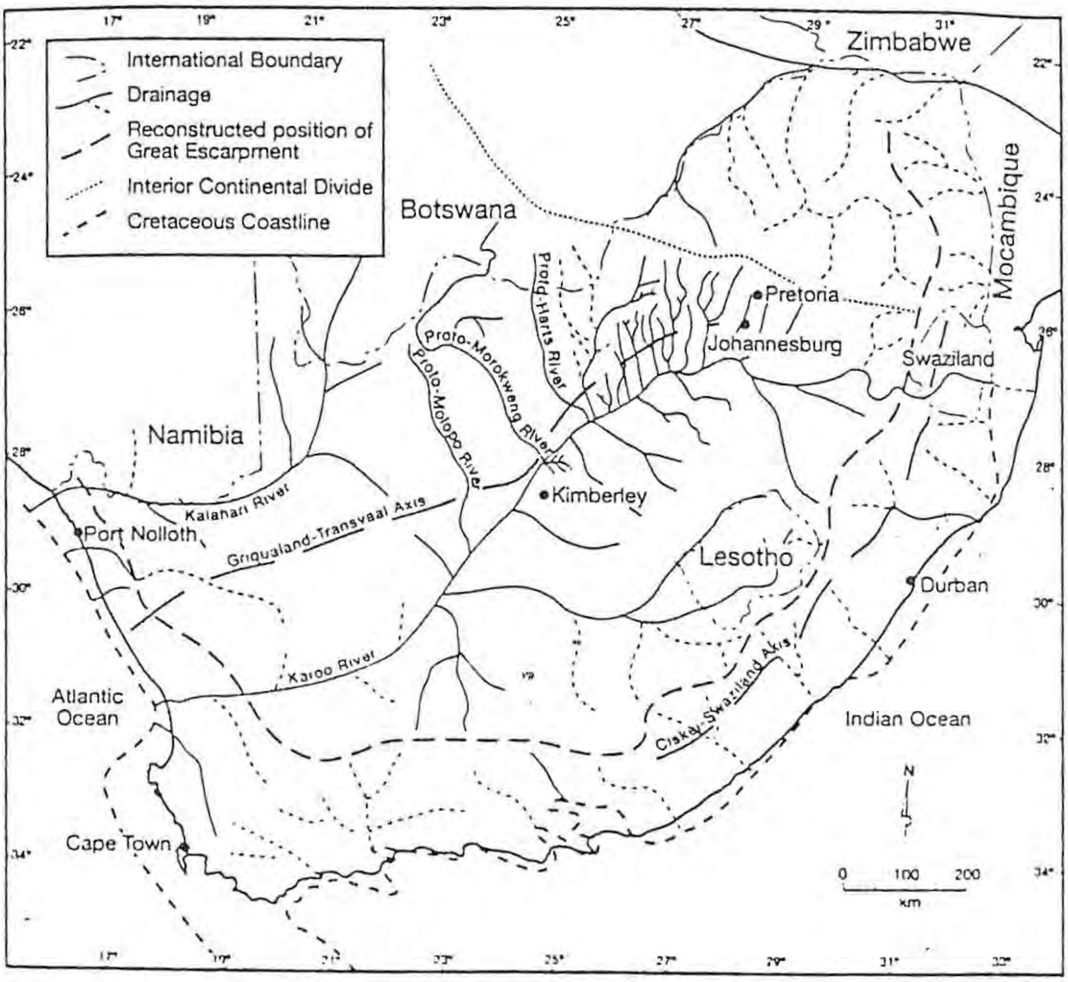


Figure 30. Reconstruction of the mid-cretaceous drainage of southern Africa after Partridge (1998)

Two major periods of river rejuvenation followed the Cretaceous: one in the Miocene, the other in the Plio-Pleistecene. With the Miocene reactivation came substantial changes in the drainage net: the middle Orange was captured by the lower Orange, causing the western reaches of the Karoo River to become defunct, and a major south bank tributary of the Orange, the Geelvloer-

Koa system, replaced those north bank counterparts that had been beheaded in the late Cretaceous by the formation of the Kalahari Basin (Figure 31). At the same time the 40-60 m proto-terraces developed along the Limpopo, Vaal and Orange rivers. Although elevation is a rough guide for correlative purposes, it must be stressed that, within each terrace, there may be more than one level, indicating that each terrace may represent gravel accumulation over a considerable period of time. In the upper reaches of the Orange/ Vaal River system remnants of primary alluvial gravels were preserved in the North-West Province, while in the Kimberly area the Holpan channel survived as part of the 60 m terrace. The Koa/Geelvloer River, which fed into the lower Orange at Henkries and had its headwaters near Sutherland (de Wit, 1993), transported diamonds from gravel remnants of the Karoo River. The environment based on paleontological evidence, was forested, and vigorous flow resulted in major river incision.

After a period of aridity during the late Miocene, which saw the development of major duricrusts, late Pliocene rivers again incised their valleys and the Younger gravel sequences were deposited. In the lower Orange, these are represented by the meso-terraces, and in the middle and upper Orange/Vaal catchment by 12-30 m terraces. The Koa/Geelvloer tributary eventually became choked and disrupted by warping, and the upper section of that river was captured by the Carnavon Leegte, which had a very restricted basin area. Concurrently the Molopo established its present course on sediments of the Kalahari Group, thereby reaching the southern rim of the Kalahari basin (Figure 32). This fluvial phase was unlike that of the Miocene, when almost tropical conditions prevailed, but was wet enough to support grasslands. Finally, oscillating climates during the Pleistocene saw the deposition of both gravels and finer fluvial sediments in cut-and-fill sequences typified by the Rietputs and Riverton Formations of the lower Vaal catchment.

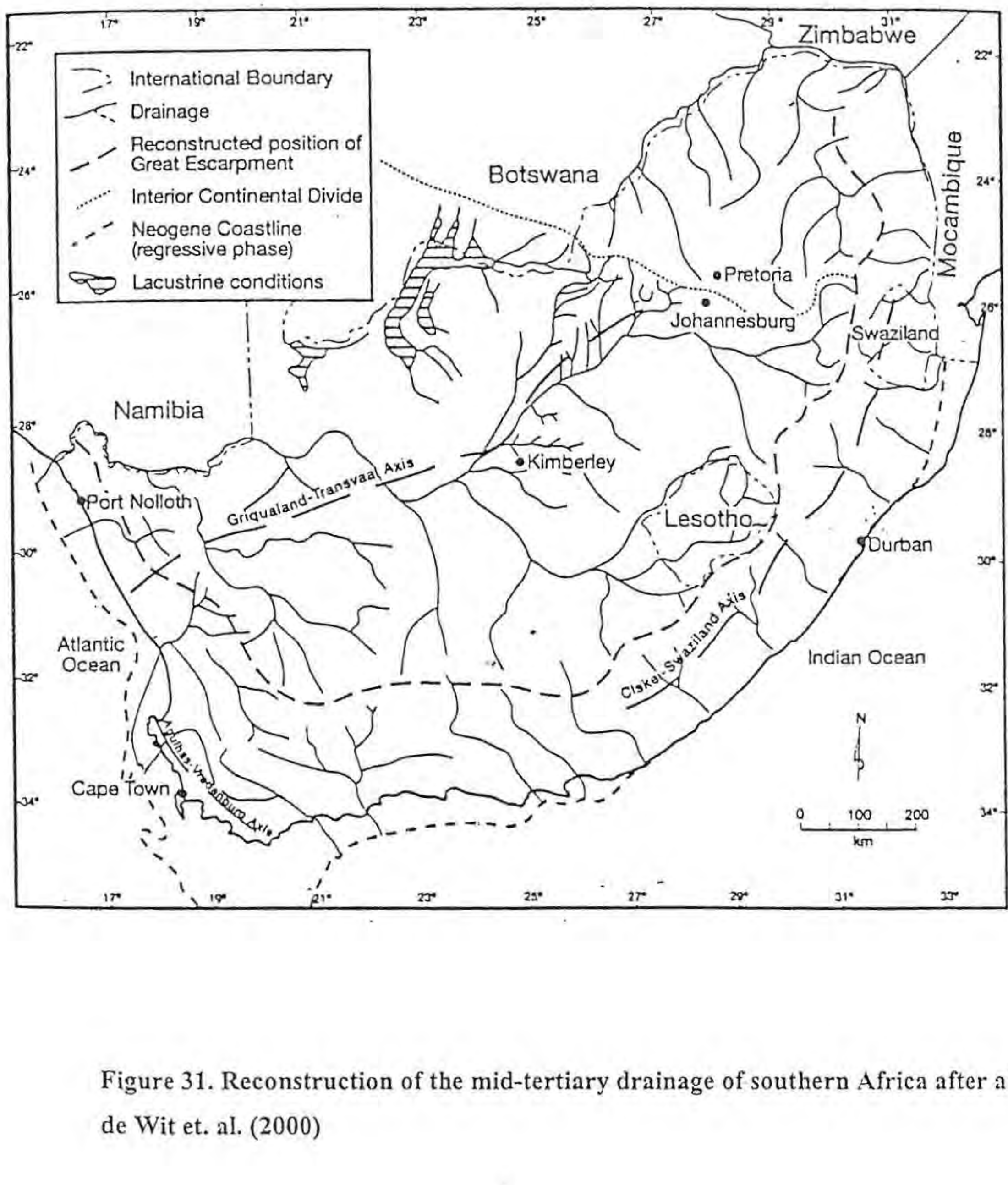


Figure 31. Reconstruction of the mid-tertiary drainage of southern Africa after after de Wit et. al. (2000)

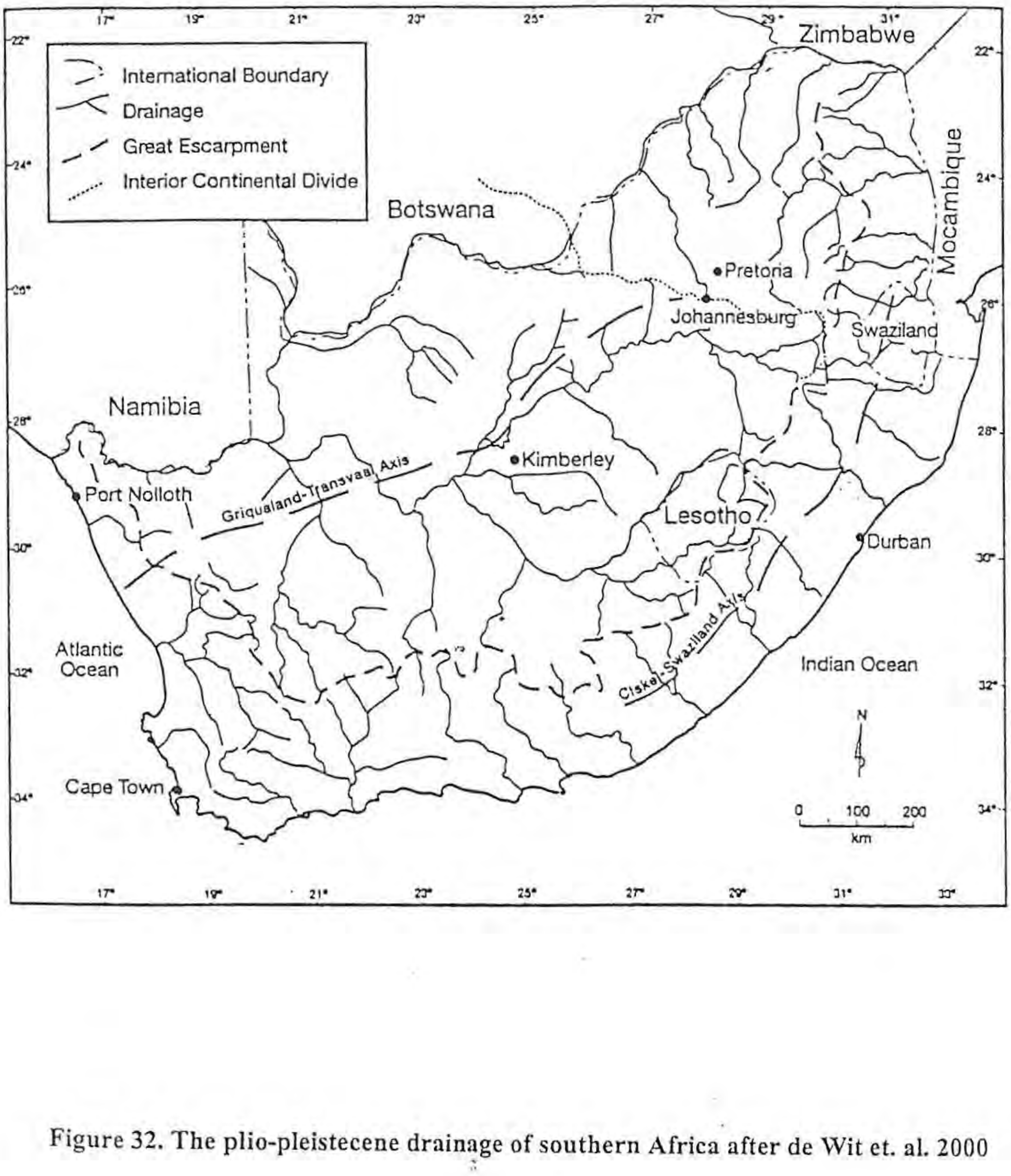


Figure 32. The plio-pleistecene drainage of southern Africa after de Wit et. al. 2000

5.4 GEOMORPHIC DEVELOPMENTS IN THE KALAHARI BASIN

5.4.1 Structural background

In a structural sense the Mega Kalahari occupies a series of contiguous Phanerozoic sedimentary basins (Figure 4); respectively the Congo basin, centred on the DRC, the Cubango/Barotse basin (Angola) and the Kalahari basin, approximately coincident with the area known as the Kalahari Desert. In the Kalahari context, these basins appear to have been sedimentologically linked and deposition enhanced or renewed by many of the events described above during the demise of Gondwanaland and the onset of Africa as a continent in its own right (Table 4), Table 1 Thomas and Shaw, 1993). Though the division of the supercontinent commenced 200 Ma ago, the initial opening of the south Atlantic did not begin until the early Cretaceous with full marine conditions achieved by 80Ma. Early division was marked by rifting (De Swardt and Bennet, 1974), particularly off-shore and in south-eastern Africa, with the later stages of break-up more likely achieved by gentler earth movements involving downwarping around the southern African coastline and in the subcontinental interior, with a zone of flexural uplift (the hinge line) in between (Figs 2 and 3 Thomas and Shaw, 1990).

Table 4. Major events in the evolution of the Kalahari after Tomas and Shaw (1993)

Period	Epoch	Ma		
	Holocene	0·01		Development of major landform suites in surface sediments
Quaternary				Continued subsidence in Okavango Rift
	Pleistocene	2		Establishment of modern course of Zambezi River
	Pliocene	5		Deposition of Kalahari sediments
Tertiary	Miocene	22		throughout Tertiary; progressive
	Oligocene	38		capture of endoreic rivers; further
	Eocene	55		uplift of Escarpment
	Palaeocene	65		Deposition of lower Kalahari Group sediments in interior basin by endoreic rivers
			80 Ma	Full marine conditions in South Atlantic
Cretaceous			130 Ma	Initial opening of the South Atlantic
		136		Major phase of rifting and initiation of the hingeline and Great Escarpment
Jurassic			180 Ma	Separation of Antarctica from eastern margin of southern Africa
		195		
Triassic			200 Ma	Beginning of Gondwanaland break-up

From a Kalahari perspective this had a 2-fold significance. Firstly, it provided a mechanism for the development of an extensive intracratonic continental basin. Secondly, it created an uplifted rim around the basin, today represented by the Great Escarpment, which acted as the source for basin-fill sediments through subsequent hingeline erosion. Hingeline uplift is thought to have continued after Gondwanaland division had occurred, due to isostatic adjustments (Summerfield, 1985), thereby maintaining the momentum for internal sedimentation, which locally continues to the present.

5.4.2 Drainage systems

One of the most important processes of Kalahari evolution that has occurred since the present structural framework was established in the Mesozoic, and one that has contributed markedly to the present character of the Kalahari Desert, has been the evolution of drainage systems (Thomas and Shaw, 1993). Very often in exploration, whatever the particular economic mineral being sought after, drainage networks in an area of interest are of primary importance in the exploration strategy. Drainages can provide sampling media representative of large catchments areas, expose bedrock geology, and reveal structural information as well as simply providing access. Understanding the drainage history in any exploration area, particularly events in which drainages have been diverted, captured, terminated or rejuvenated is of high importance. In kimberlite exploration, drainages provide a key sampling medium for kimberlitic indicator minerals in the secondary environment. Therefore greater attention is given to this section which attempts to unravel the complex drainage history within the Kalahari.

The work of Smit (1977), DeSwardt and Bennet (1974) and others has indicated that a major factor in at least the early accumulation of sediments in the interior basin was an endoreic drainage system, transferring sediment from the uplifted hingeline to the Kalahari. Overall, the tectonic framework established in southern Africa by the division of Gondwanaland led to the creation of a dual drainage system, with the hingeline acting as a watershed between a coastally orientated exoreic system and that draining into the interior.

Today the endoreic system possesses only one major river, the Okavango, which terminates in the Kalahari. Although the central section of the Mega Kalahari basin (Angola and DRC) is traversed by perennial rivers that have their headwaters in hingeline locations, all but the

Okavango of those flowing generally southwards have been captured by more erosively aggressive coastal rivers (Thomas and Shaw, 1991).

Capture of the Cunene River in Angola, Fish River in Namibia and Molopo system in South Africa have all received scientific attention (e.g. Beetz, 1933; Wellington, 1938; McCarthy, 1983), with perhaps most interest afforded to the evolution of the Zambezi system. This has recently been considered elsewhere (Thomas and Shaw, 1988, 1992, Nugent, 1990) and two points particularly relevant to the Kalahari evolution were noted. Firstly, the present course of the Zambezi is geologically young. Lister (1979) followed earlier workers to highlight that the modern Zambezi course was not established until the early-middle Pleistocene. Although one author has attempted to provide a more precise figure (125,000B.P.: Nugent, 1990), there are a number of doubts about the methods used to achieve this date (Thomas and Shaw, 1992).

Secondly it follows that the Zambezi was therefore a significant geomorphological feature of the Kalahari Desert prior to establishment of the present course. The southward course of what is now the Upper Zambezi is variously thought to have led to it being a tributary of the Orange (Lister, 1979) or Limpopo (Bond, 1963) but it is more likely to have been endoreic, terminating in the Kalahari and depositing sediment much the same way as Smit (19977) proposed for an ancestral north-flowing Orange River. It is also probable that the Upper Zambezi contributed to the development of the Makgadikgadi basin sediments and inflow (Cooke, 1980) and that its present course may in part be due to earth movements in that area, discussed below. Later Pleistocene links between Zambezi discharge and the northern Botswana lake system have also been significant (Shaw and Thomas, 1988).

5.4.3 Paleo-lakes of the Kalahari

Despite the limited number of basins for which data are available and the fact that most records are short or discontinuous, they are important in showing that, at times in the past, rainfall was sufficient to sustain large lakes in the now predominantly semi-arid Mega Kalahari (Figure 33). Today, seasonal pans (or playas) are common over large areas and these were probably present throughout the Quaternary; their ephemeral nature and susceptibility to wind deflation limits their potential as archives of past environments, but some useful records have survived in favourable settings (Partridge and Scott, 2000).

Evidence for the former existence of mega-lakes in the semi-arid Kalahari points to major changes in the Quaternary environments (Thomas and Shaw, 1991). Interpretation of the paleoclimatic signal is, however, complicated by the fact that all of these water bodies were fed by large rivers originating in the high rainfall areas of eastern Angola.

The most important paleo-lakes of Botswana were marginal to the present day wetland system of the Okavango Delta which is fed by the Okavango River. These lakes almost certainly had past connection with the Zambezi River via the Kwando-Linyanti-Chobe depression, which links the

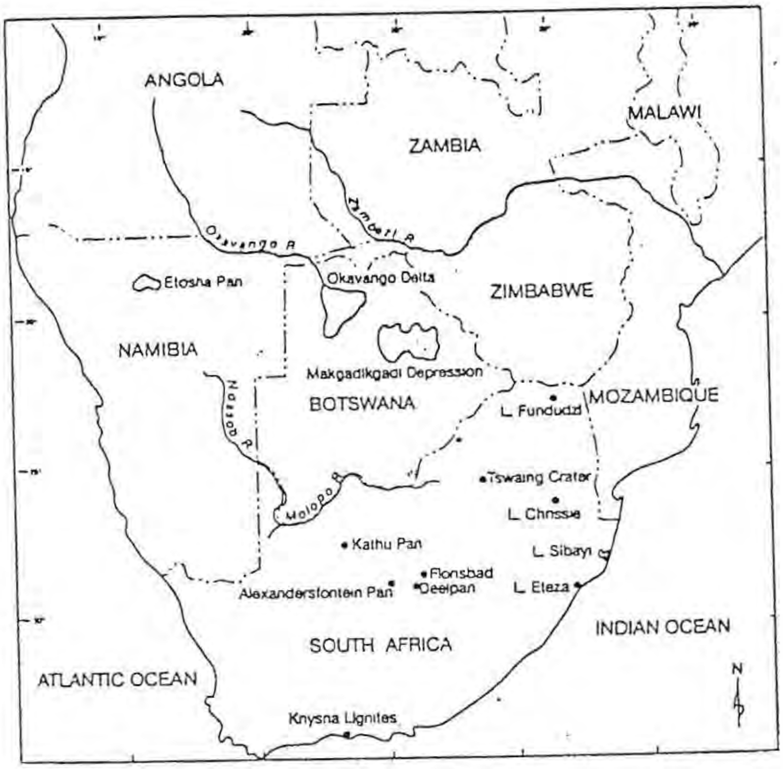


Figure 33. Locality map showing southern African lakes and pans discussed in the text after Partridge and Scott (2000)

eastern part of the Delta to the Zambezi trough above the Mambova Falls (Figure 34). Both the termination of the delta and the link to the Zambezi are controlled by Quaternary faults which continue to show seismic activity; neotectonic influences may therefore have played an important role in providing conduits for inflows from both river systems and in creating topographic thresholds for different phases of lake formation. The accordance of paleo-shorelines throughout the major systems is nonetheless remarkable and points to the dominance of climatic factors in their genesis.

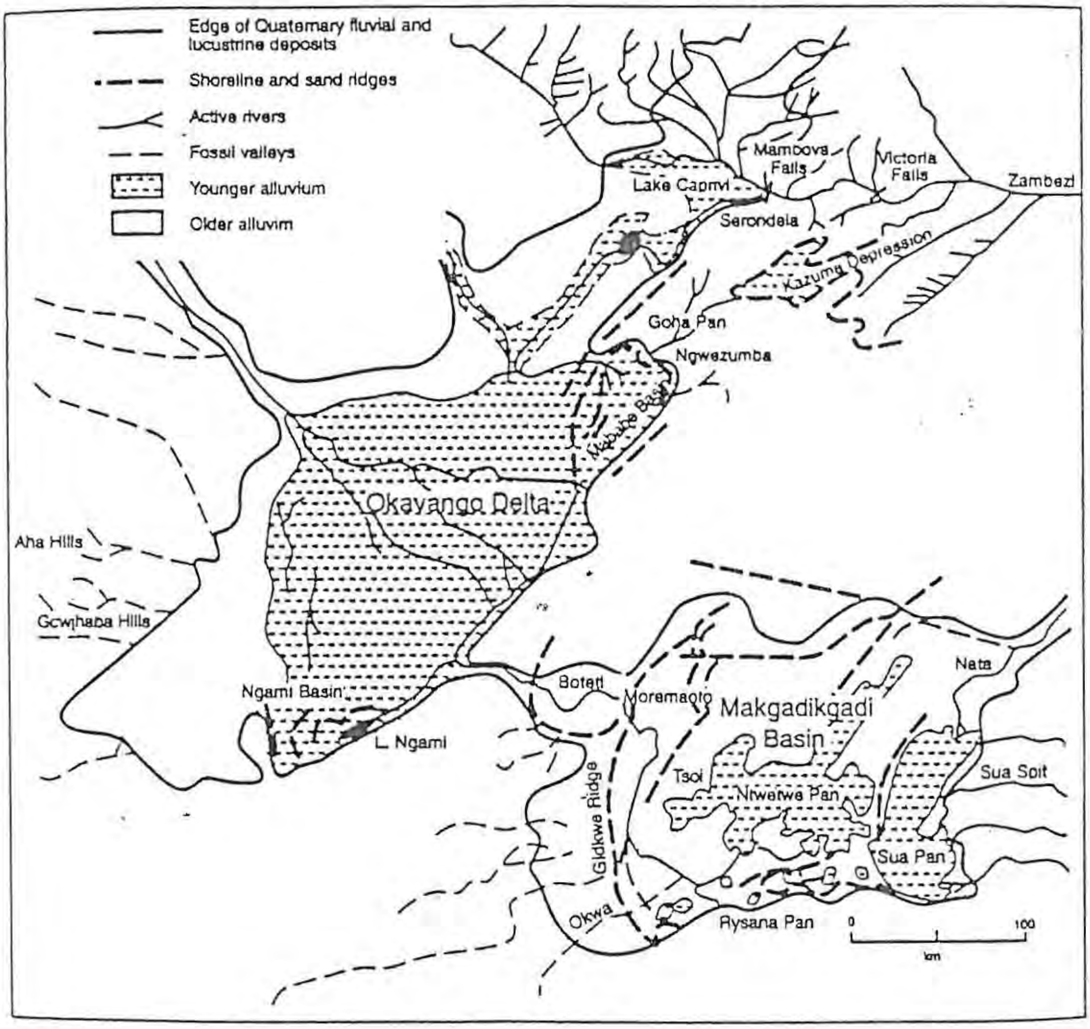


Figure 34. Paleo lakes of the central Kalahari after Shaw and Cooke (1996)

The major paleo-basins linked to the distal portions of the Okavango Delta are Ngami, the Mabebe and the Makgadikgadi (Figure 34). Lake Caprivi, which includes the ephemeral marsh of the present Lake Liambezi, occupied an area on either side of the Zambezi Formation. Thomas and Shaw's (1991) comprehensive account of these features indicates that the paleo-shorelines of Ngami Basin enclose an area of about 1800 km^2 which is bounded to the west by the Dautsa Ridge; like its counterparts in the other basins, this was probably an offshore barrier bar (Partridge and Scott, 2000). Shoreline elevations range from 940-945 m amsl to 923,5 m which is the maximum level of the present-day Lake Ngami. The Mababe basin of some 4000 km^2 is fed erratically from the Okavango Delta via the Thamalakane and Savuti channels, probably as a result of neotectonic adjustments; the western margin is formed by the Magikwe River, which, like the Dautsa Ridge, has crest elevations of 940-945 m. Other shorelines present are 936 m, 930 m and 927 m. Lake Caprivi had an area of about 2000 km^2 ; remnants of its offshore bars to the south of Lake Liambezi have crest elevations of 936 m.

The paleo-lake which occupied the Makgadikgadi Basin was much larger, covering some 37 000 km^2; it lies a little more than 100 km to the south east of the Okavango Delta, to which it is joined by the Boteti River. The long axis of the basin is controlled by recent faults, and it is bounded to the north and west by the Gidikwe Ridge. The crest elevation of this feature is again 940-945 m, indicating the unity of the Ngami-Mababe-Makgadikgadi system at the time of its maximum extent. The basin is presently occupied by the Sua and Ntwetwe, between their surfaces and the basin margin shorelines are preserved at the 920 m and 912 m levels (Partridge and Scott, 2000).

The entire system has been named Lake Paleo-Makgadikgadi by Grey and Cooke (1977). At its maximum it covered an area in excess of 80 000km^2, including the lower part of the Okavango Delta and much of the upper Zambezi trough. Its extent therefore exceeded that of today's largest African lake (Lake Victoria in Uganda). The 936 m level was evidently controlled by a single hydrological threshold in the Ngami, Mababe and Caprivi basins and represents a lake with an area of some 7000 km^2 in the Delta area. Referred to as Lake Thamalakane by Shaw (1988), this water body probably overflowed via the Boteti River into the Makgadikgadi Pan to feed the 920 m lake, which it enclosed at that time (Partridge and Scott, 2000). Diatomaceous earths at Moremaoto on the lower Boteti River have been linked to ponding at this level and have yielded OSL dates of 32 –37 ka (Shaw et al., 1997). These dates are considered more reliable than radiocarbon ages, obtained mostly from calcretes associated with the various shoreline features. Such calcretes are polygenetic and may postdate the deposits in which they formed, often by a substantial margin. They have given a wide range of apparent ages for features at the same level. All that may be concluded from the present data is that Lake Paleo-Makgadikgadi probably formed some time during Isotope Stage 3; water budget analyses indicate that an inflow of some 50 km^3/yr would have been required to sustain such a lake, which implies a major input from the Zambezi River (Thomas and Shaw, 1991). The Lake Thamalakane stage was characterised by multiple transgressions, suggesting that the 936 m threshold was reached repeatedly over a considerable period of time, possibly both before and after the Last Glacial Maximum, the dates of 32-37 ka cited above may define the earlier part of this history. Shaw (1988) has estimated that an increase in rainfall of about a 100 per cent would have been required to sustain this level and allow overflow sufficient to maintain a lake at the 912 m level in the Makgadikgadi Basin.

Any account of the paleo-lakes of the Kalahari would be incomplete without reference to the Etosha Pan of northern Namibia. Like the Botswana paleo-lakes, this basin was fed originally by the Cunene River, which has its source in the well watered highlands of Angola; the lake was subsequently drained as a result of piracy by a coastal stream extending headward from the vicinity of the Ruacana Falls (Wellington 1938). A lake defined by stromatalites at + 8m is beyond the range of radiocarbon dating; its flooded area would have exceeded 6000 km^2. The evolution of the system after it was cut off from the Cunene River is typical of that displayed by the many pans of the semi-arid western interior of the subcontinent: there is evidence of repeated active cycles of inundation, evaporation and aeoloian activity which caused lows in the floor of the basin to propagate and coalesce. Intervening periods of calcrete growth gave temporary stability to the periphery of the basin.

5.4.4 Neotectonics

Tectonic activity within the Kalahari basin has continued to the present and has contributed significantly to locally enhanced sedimentation rates and also to drainage changes. Both the Okavango Delta and the Makgadikgadi depression occupy fault controlled structures (Reeves, 1972) with the Delta lying in a north-east-south-west rift structure that is an extension of the East African rift system (Scholz, 1976). Subsidence in the Okavango system has been up to 1000 m (Hutchins et al., 1976), favouring the accumulation of a considerable thickness of fluvio-deltaic sediments. The earth movements responsible for the initial development of the Makgadikgadi basin also involved uplift on the southern side of the depression (Bond, 1963) which may have contributed to renewed endoreism in the Upper Zambezi and to its subsequent capture (Thomas and Shaw, 1991). Within the Kalahari, the links between long term valley development and structural lineaments in the sub-Kalahari basement are the subject of investigations by Nash (1992).

5.4.5 Basin Margin erosion

Although the Kalahari occupies a structural basin, it has the appearance of a plateau along parts of the western and eastern rims, notably in southern Namibia and western Zimbabwe. In Zimbabwe this is a consequence of the eastern extent of the Kalahari being reduced through back cutting and erosion by southbank tributary systems of the middle Zambezi during the Pleistocene (Thomas and Shaw, 1988), continuing up to the present day. The efficiency of the erosion achieved by the middle Zambezi system when compared to other networks of the Kalahari rim has been due to rejuvenation caused by uplift of the southern Afican hingeline (DeSwardt and Bennet, 1974), continued downwarping along the Gwembe Trough, through which the middle Zambezi flows (Dixey, 1945), and its location on the wetter eastern side of the subcontinent. In Namibia, erosion by the Fish River and its tributaries is creating a scarp near Mariental and Keetmanshoop.

In western Zimbabwe therefore, the sub-Kalahari basement has been revealed in the region stretching south-east from Victoria Falls to Bulawayo, with the Kalahari rim existing in the form of a scarp. Away from the river valleys, isolated outliers of Kalahari sediments remain on many interfluves in north-central Zimbabwe.

6. AGE AND PALEOENVIRONMENTAL HISTORY OF THE MEGA KALAHARI

Constraints on the age and paleoclimatic history of the Mega Kalahari have been a protracted and difficult affair. Although many 19th century travellers (e.g. Livingstone, 1858) realised that the Kalahari landforms had a paleoclimatic significance, it is only since the publication of Grove's (1969) paper on the region that the potential of remote sensed imagery and radiocarbon dating has been used to produce a preliminary palaeoenvironmental framework. Difficulties in accurately dating the sequence have arisen from factors such as; the paucity of fossils and datable materials, lack of sites with an uninterrupted depositional history, and the absence of agreement on stratigraphy at even a local scale. Faced with a fragmentary archaeological record, and the unsuitability of palynology as a research tool, research has focused on studies of large landforms. Further difficulties have also arisen from inadequate field investigation, and the importation of hypothesises from other parts of Africa.

Much of the dating has focused on of the upper, younger Kalahari sediments such as; the dune fields, lakes and pans, dry valleys and springs, caves and upper duricrusts. This is largely because these landforms are more readily accessible for study and sampling. An accurate age estimate of the deeper, basal Kalahari sediments has proved more difficult.

There are several broad age estimates for onset of the Kalahari based largely on the relationship of the Kalahari to the underlying sequences. Partridge and Maud estimate the start of the deposition to be late Cretaceous representing a transition to dryer climatic conditions from the predominantly wet conditions of the Jurassic and early Cretaceous. Deposition occurred onto a erosional landscape correlated with the African Surface of King (1962). From the work that Money (1972) did in the Kalahari of western Zambia his estimates suggest a late Jurassic age for the onset of this succession.

Most of the early studies that attempted to elucidate the different aeolian episodes during which the Kalahari sands were deposited, focused on inferring periods of aeolian activity from indirect proxy indicators such as lowered lake levels, hiatuses in speleothem formation etc. Invariably the chronometer used was radiocarbon from sources which include pedogenic calcretes (Cooke and Verstappen, 1984; Lancaster 1989), spring tufa (Lancaster, 1989), buried vegetal matter (Heine,

1988) and molluscs (Heine, 1981). Chronological correlation between the various paleoenvironmental reconstructions is generally poor but the results suggest climatic fluctuation between arid and humid conditions for the period ca 50-11 ka.

The emergence of luminescence dating method made the direct age determination of aeolian sediments possible as opposed to inferring phases of aridity from the absence of indicators of humid conditions. Luminescence dating studies carried out in southern Africa so far (Wintle, 1999) point to episodes of significant aeolian mobilisation of the Kalahari Sands in Western Namibia and central southern Botswana.

There has been much debate as to the paleo-climatic significance of calcretes ranging from formation in semi-arid to humid conditions. The present assumption is that calcrete formation is dependant on site, rather than climatic factors, and that calcrete dating can be valid in paleo-lake, pan and valley environments. In general calcretes provide a "youngest date" (Netterberg, 1978). Although Goudie (1973) and Watts (1977) record Quaternary calcretes in Botswana, it is possible that calcrete formation has been taking place episodically since at least the Pliocene (King,1963). Mabbut (1952) recorded calcrete development on schists which were overlain by Pliocene sediments. This led Netterberg (1969) to speculate that calcrete older than Pliocene must exist in southern Africa. He proposed four major episodes of calcrete formation (1) Pliocene times, including the Mid-Pliocene 'Kalahari Limestone'; (2) Acheulian or Middle Pleistocene times; (3) Upper Pleistocene time (4) Late Pleistocene-Recent. The precise ages of these episodes are, however, not well known. Netterberg (1969) also suggests that calcretes are forming at present within the Kalahari.

Thomas and Shaw (1996) established a chronological framework for the past 40,000 years within a large study area of Botswana, based on a corpus of 239 radiocarbon dates from various landforms within the Kalahari. The sites yielding these dates are shown in Figure 35.

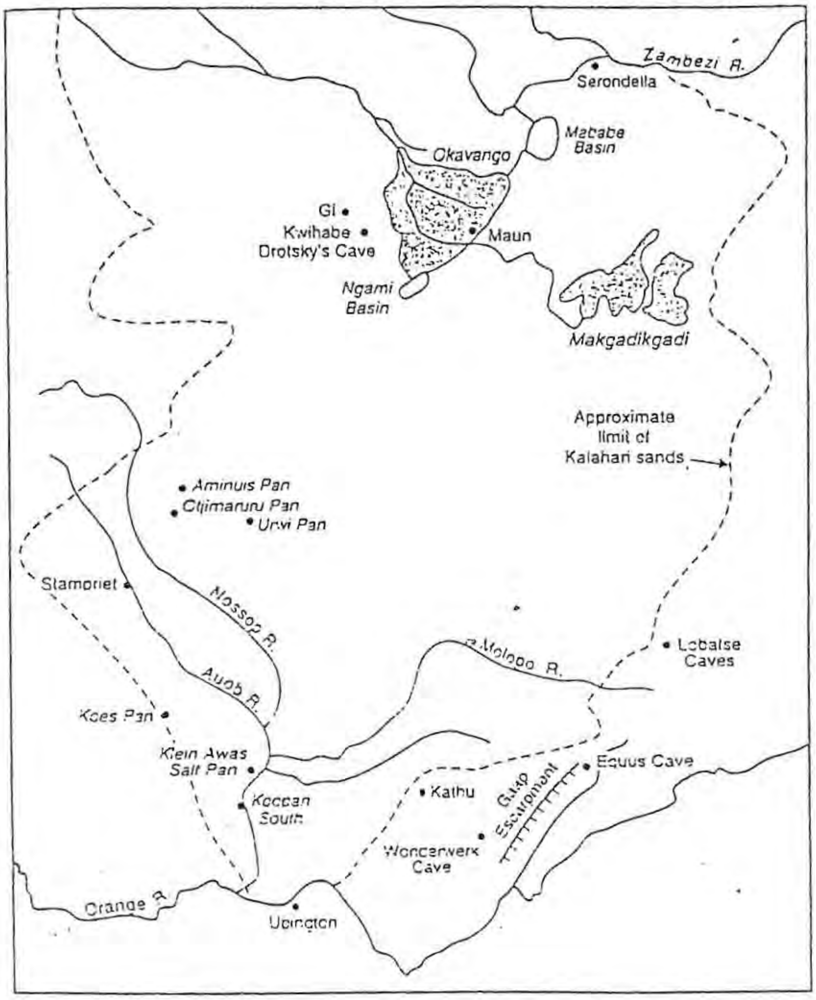


Figure 35. The location of sites yielding important paleoenvironmental data for the Kalahari (after Deacon and Lancaster, 1988; Thomas and Shaw, 1991)

er half the samples dated by Thomas and Shaw (1996) used calcrete as the sampling medium l require careful interpretation. Future reinterpretation of the radiocarbon dates will become essary in the Kalahari as improvements in the calibration of radiocarbon errors are made. The eoenvironmental information for all the major Kalahari sites sampled by Thomas and Shaw

(1996) are shown in Figure 36. The data indicate precipitation changes translated through the proxy of landforms. As the Kalahari in Botswana occupies a transitional position between the tropical summer rainfall and temperate winter rainfall belts, some intra-regional variation should be expected.

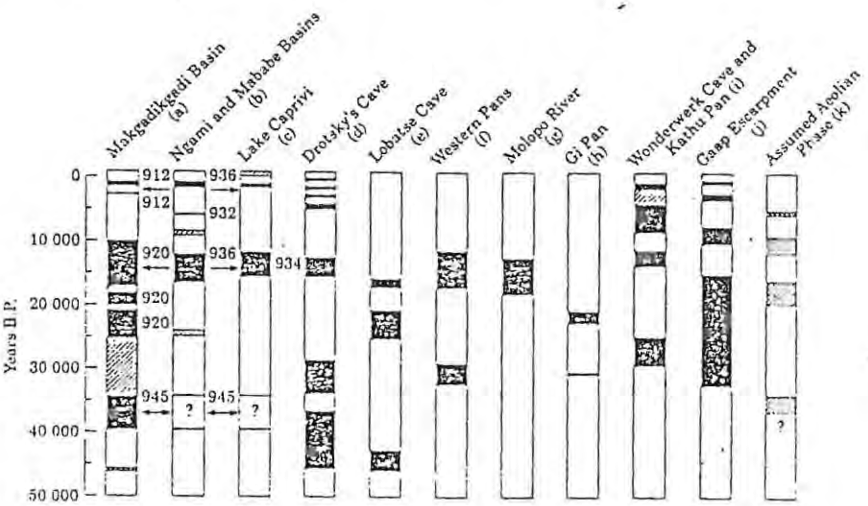


Figure 36A summary of palaeoclimatic data for the Kalahari. Numbers indicate identified high stands of Lake Palaeo-Makagadikgadi in metres a.s.l. (After Thomas & Shaw, 1991*a*. Sources: (a) Cooke & Verstappen, 1984; Cooke, 1984; (b) Shaw, 1985; Shaw & Cooke, 1986; (c) Shaw & Thomas, 1988; (d) Cooke & Verhagen, 1977; Cooke, 1984; Shaw & Cooke, 1986; (e) Shaw & Cooke, 1986; (f) Deacon & Lancaster, 1988; (g) Heine, 1982; (h) Helgren & Brooks, 1983; (i) Beaumont *et al.*, 1984; Butzer, 1984*a*; (j) Butzer *et al.*, 1978; (k) Deacon & Lancaster, 1988.) Key: = 1. High levels (lakes and pans); 2. Perennial flow (rivers); 3. Sinter growth (caves). = 1. Low lake levels. = 1. Assumed aeolian phases.

The record for 50-20,000 years B.P. show the most discrepancies, with some conflict between the Paleo-Makgadikgadi record, and adjacent cave and pan sites in north-west Botswana. Humid conditions seem to have prevailed at different times, including 35-28,000 years B.P. at Drotsky's Cave, Gi Pan and the south-west, and 24-22,000 years B.P. in the Makgadikgadi, Gi Pan and Lobatse Cave. The Okavango linked dates of Ngami and Mabebe dried out c. 25,000 years B.P., suggesting conditions at least as dry as present.

From c. 20,000 B.P. through to the Last Glacial Maximum at 19-18,000 B.P. cold and dry conditions prevailed, possible accompanied by aolian activity and limited peat formation. This was followed by an episode of greater moisture availability at 17-12,000 B.P., with a core of 16-

13,000 B.P. This episode had great spatial extent, occurring throughout the Kalahari and southwards into the northern Cape and Orange Freestate, but not eastwards into the Transvaal. It is also important to note that this wetter episode is out of phase with Africa north of the equator, and with other major sub-tropical deserts.

By 12,000 B.P. drier conditions, with falling water tables and diminishing lakes, set in. The Holocene has been characterised by a series of unphased fluctuations of lower amplitude and duration, and has included humid episodes in the Middle Kalahari at 6,000 B.P., 4,200-3,600 B.P. and 2,500-2,000 B.P., the latter indicated by both cave and lake studies. In the southern Kalahari these episodes were not synchronous, with the last major humid episode at 3,200-2,000 B.P. Smaller fluctuations have been suggested in the archaeological record (Thomas and Shaw, 1991).

Munyikwa and others (2000) applied thermolumonescence dating to 10 samples from the aeolian Kalahari Sands of western Zimbabawe using the total bleach (additive dose) and regeneration techniques on coarse-grain quartz separates. Their results suggest that the main phase of sand accumulation occurred between 10 and 96 ka. The oldest sands were found in the Victoria Falls area and were dated between 96±8 and 160±23. According to Munyikwa et al. (2000), the dune building activity in western Zimbabwe occurred in recurrent cycles of aridity interspersed with periods during which aeolian activity was limited or non-existent.

Luminescence dating and stratigraphic dating by Stokes et al. (1997) show that the last major period of linear dune development in the southern Kalahari occurred between 17 and 10 ka and probably involved reworking of sediment deposited in a previous period of dune building from 28-23 ka that may represent the initial deposition of sand in this region. Holocene dune activity in the region was localised in extent and occurred at 6-5 and 1-2 ka (Thomas et al., 1997), although final stabilisation of linear dunes may have occurred as recently as 9-8 ka in drier western areas of the region (Blumel et al., 1998).

Parallel studies of dune chronology in the northern and eastern dunes in Botswana indicate that multiple periods of dune building, each spanning 5-20 ka, occurred during the late Pleistocene (Stokes et al., 1997). Ages for these periods are 95-115, 41-46 and 20-26 ka. Short lived periods of aridity were separated by much longer periods of humid conditions (Stokes et al., 1998). Late Glacial and Holocene aolian activity was restricted to reworking of the crestal areas of the linear dunes in the area of Hwange National Park, Zimbabwe during the period 10-16 ka, and to western

Zambia, where linear dune construction from local sediment sources occurred 32-27, 26-13, 10-8, and 5-4 ka (O'Connor and Thomas, 1999).

7. KIMBERLITES BENEATH THE MEGA KALAHARI

Southern Africa contains the highest known concentration of kimberlites in the world, in excess of 850 occurrences. It is a well established rule of diamond exploration that the most prospective areas for potentially economic diamondiferous kimberlites are regions underlain by Archean cratons (Clifford, 1966; Boyd and Gurney 1986; Helmstaedt & Gurney, 1995). Cratons are extensive, stable continental areas comprising a shield, or exposed core of a craton, and overlying platform sequences consisting mostly of sediments and, in places associated volcanics. Cratons are characterised by a thick lithospere and low geothermal gradients and provide the conditions necessary for diamond formation within the mantle (Boyd and Gurney, 1986). The craton areas should be relatively unaffected by major deformational and metamorphic events to allow preservation of early formed diamonds.

All known diamondiferous kimberlites in southern Africa occur within the boundaries of the Zimbabwe/Kaapvaal Craton. This cratonic block is bounded in the north and northeast by the Pan African (-600 Ma) Zambezi and Mozambique belts respectively, in the west, by the early-Proterozoic (1800 Ma) Kheis belt, in the south by the mid-Proterozoic (1100 Ma) Namaqua belt, and in the east by the Jurassic (200 Ma) Lebombo monocline (Skinner et al., 1992). The craton has an ancient root, formed earlier than 3 billion years ago, composed mainly of peridotites that are strongly depleted in basaltic components (Cf. Herzberg, 1993). The asthenosphere boundary shelves from depths of 170-190 km beneath the craton to approximately 140 km beneath the mobile belts (Boyd and Gurney, 1986).

The Zimbabwe Craton collided with the Kaapvaal Craton between approximately 2.7 Ga and 1.95 Ga. The combined craton, called also Kalahari Craton, underwent a complex late and post-Archean history, detailed earlier on.

Kimberlites in southern Africa have been divided into Group I and Group II types (Skinner, et al 1989). Six provinces are recognized for the Group I and one or two for the Group II. The kimberlite magmatism ranges in time from the Proterozoic to the late Mesozoic. An earlier, Archean, diamondiferous event, is also probable, indicated by the occurrence of detrital diamonds in the Witwatersrand Basin. Several emplacement periods are recognized: Lower Proterozoic (Kuruman), Middle Proterozoic (Premier), Early Paleozoic(Venetia, Martins Drift, River Ranch, Triassic (Syn Karoo Jwaneng), Middle Cretaceous and Upper Cretaceous. The mid-Cretaceous event is the most widespread by far.

Defined in figure 37 are the Kalahari and Central African cratons, which form the cornerstones of Southern African geology. The Kalahari craton underlies most of Zimbabwe, the central and northern part of South Africa, Lesotho, Swaziland and eastern Botswana. The Central African craton underlies most of Angola and the southern parts of Gabon, Cameroon and the Democratic Republic of the Congo. Also shown are the portions of these cratons which are covered by sediments of the Mega Kalahari.

The location and boundaries of cratons are defined by a variety of factors. These include; seismic tomography, geological mapping and distribution of kimberlite intrusives for which the mineral chemistry is known and indicates the depth of underlying mantle root. However, many parts of the craton boundaries, particularly those under cover of the Mega Kalahari, are largely inferred. As further kimberlite pipes are discovered beneath this succession, the craton edges may be adjusted and better constrained. From figure 37 it can be seen that several hundred kimberlites have been discovered beneath the Mega Kalahari, both on and off craton.

Kimberlites and lamproites have certainly intruded the Earth for a very long period. Known deposits record ages between 1600 Ma and 20 Ma. Most known kimberlites are younger than 600 Ma but deposits hidden by depositional cover or removed by erosion may obscure this distribution. The age of intrusion for southern African pipes varies from the mid-Proterozoic to the mid-Cretaceous spike at 80-120 Ma. As the Kalahari is not considered to be older than late Cretaceous, and more probably early Tertiary (Paleocene), this succession should post-date and overlie all kimberlite occurrences with only one known, possible exception. This is an unsubstantiated De Beers report of several small kimberlites of the Sikereti cluster in north-western Namibia which may be syn-Kalahari in age. The possibility of syn-Kalahari pipes is not excluded, merely considered unlikely.

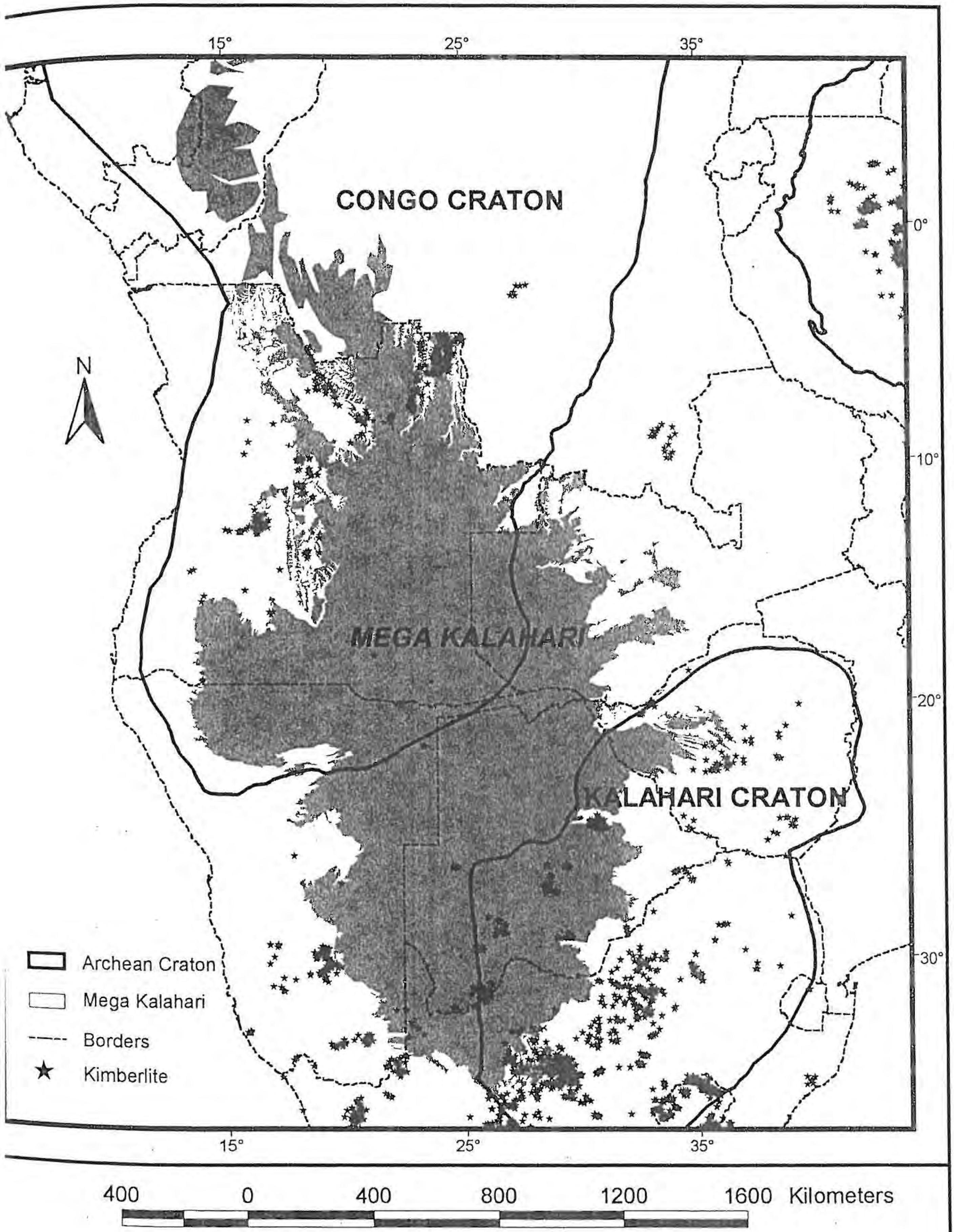


Fig 37. Central and Southern Afrca, showing the extent of the Mega Kalahari Sediments overlying known craton and kimberlite occurences

A brief account follows of known kimberlite distribution beneath the Mega Kalahari by country.

Botswana

Eastern Botswana occurs on the Kalahari Craton and is overlain by Phanerozoic Karoo Supergroup platform rocks, in turn overlain by Tertiary and Quaternary Kalahari sediments.

Janse (1995) reported 140 kimberlites, grouped into 11 clusters. In total, 56 are diamondiferous. The large, highly economic Orapa pipe (106 Ha), which partially outcrops in the Kalahari, was discovered by De Beers in 1967. Follow-up work identified an additional 29 pipes including another 2 economic pipes, Lethlakane 1 and 2. The Jwaneng pipe (45 Ha), probably the most spectacular economic pipe in the world, was discovered in 1975 by De Beers under 55 m of Kalahari sand cover. Currently, Orapa, Lethlakane, and Jwaneng collectively produce about 15 million carats annually.

Prospecting by Falconbridge in the 1970s discovered an additional 60 buried pipes in several fields. Only one pipe, Gope 25, found under 75 m of Kalahari cover, appears to have some economic potential.

Namibia

Numerous kimberlites occur in at least 6 fields in soutwestern, central, east-central and northwestern Namibia, but, with the possible exception of Sikereti all of them are located off craton and none have been proved to be even remotely economic. There is some uncertainty as to the presence and extent of craton underlying northeast Namibia and much of the craton lies beneath the Kalahari and is therefore largely inferred. Central and western Namibia lie on a triangular western extension of the South African Craton underlian by Proterozoic basement which is bordered in northern and western Namibia by the Late Proterozoic Damara mobile belt and in southern Namibia by the Late Proterozoic Namaqualand Gneiss Complex.

Angola

The first kimberlite pipe, Camafuca, was discovered in 1952 in the Chicapa River drainage area in Lunda Norte, northeastern Angola, beneath Kalahari sediments. Many other kimberlites have been discovered since then and the total number may be as high as 638, of which at least 300 are stated to contain diamonds (Janse, 1995) and probably 5-10 pipes may be economic (Jourdan, 1990).

The known Angolan kimberlites intruded an early Proterozoic-Archean craton. Host rocks vary from Archean gneisses to Triassic sandstone and shale and Late Cretaceous sediments of the Calonda and Kwango Formations. Northeast, and eastern Angola is overlain by extensive Kalahari sediments (up to 150 m in thickness) that have received little or no modern exploration for kimberlites during the last 30 years.

DRC

Diamondiferous kimberlites, which appear to be the source of the eluvial and alluvial diamonds were found in 1946 in the Bakwanga-Mbuji Maye area, also below shallow Kalahari sediments. Fifteen occurrences represent small, shallow craters above very narrow pipes.

Zimbabwe

About 30 kimberlites, seven of which were diamondiferous, were known in Zimbabwe in 1995. Several non-economic pipes were discovered in western Zimbabwe within shallow Kalahari sediments in 1996 (Mining Weekly, 1996). Most of Zimbabwe is underlain by the Kalahari Craton in which granite-greenstone rocks are near or at surface. Therefore due to erosion the likelihood of large pipes or crater facies occurring is lower than in Botswana.

8. GEOPHYSICAL APPLICATIONS WITHIN THE MEGA KALAHARI

The Mega Kalahari represents a unit of significant overburden covering bedrock-hosted kimberlites. Therefore geophysics is essential in most parts of the Mega Kalahari, particularly in defining specific targets for follow-up, drill targets and providing drill collar coordinates. It is very useful where no previous work has been done or where follow-up work is required in areas demonstrated to be prospective for kimberlites by other methods such as kimberlite indicator mineral sampling.

Modern geophysical surveys gather digital data, which enables rapid processing and imaging abilities, and can be readily integrated with geological, structural, geochemical and other datasets. Helicopter or fixed-wing aircraft (airborne) geophysical surveys are rapid and relatively low cost reconnaissance tools. The most commonly used techniques in the Mega Kalahari are magnetics, electromagnetics and gravity. Geophysics is only effective where sufficient contrast in magnetic susceptibility, conductivity, and specific gravity etc. exist between the kimberlite and its host rock.

8.1 MAGNETIC METHOD

Airborne magnetic surveys have significant advantages over ground magnetic surveys. The higher costs of airborne surveys are offset against major gains in time, increased size of survey area, reliability and sensitivity. Experience and the results of numerous programs have demonstrated the advantages of low level, high-resolution airborne surveys over regions of complex geology totally obscured by thick overburden such as the Mega Kalahari. The capability of the survey to penetrate the Mega Kalahari cover and provide a detailed regional picture of the solid structure and geology is something that, given time constraints and logistics, is often not possible with ground magnetic surveys.

Rock magnetism is a function of magnetic susceptibility, in other words the ease with which the constituent minerals may be magnetised. The highly susceptible magnetic mineral content in kimberlites can range as high as 5-10% of iron oxides, most of it in the form of magnetite and ilmenite. Magnetite is formed as a secondary mineral during the serpentinisation of kimberlite (Gobba, 1989). As the Mega Kalahari sediments are comprised mainly of quartz-rich sands, siliceous and calcareous duricrusts, they exhibit very low susceptibility and are magnetically "quiet" or uniform. The non-magnetic properties of these sediments mean that they are effectively a "magnetically transparent cover" above the bedrock. The thickness of the Mega Kalahari sediments in any area of interest must be taken into account. Areas of thick cover reduce the magnetic resolution of the underlying bedrock. Areas of thin cover (<40 m) allow better resolution and modelling of magnetic features within the bedrock.

Bromley and others (1994) conducted a multiparameter low-level airborne geophysical survey incorporating magnetic, VLF (very low frequency), and twin coil EM (electromagnetic) measurements over a portion of the Mega Kalahari in eastern Botswana. The survey demonstrated the efficacy of an airborne magnetic survey to penetrate the cover and provide a detailed regional picture of the bedrock geology. Graben, dykes, faults, areas of thin basalt, and fractures were all clearly picked out.

For the direct detection of kimberlites a flight line has to pass over or close to the pipe for an appreciable contact with the surrounding rocks. A flight height of 50-100 m and a flight line spacing of 200 m are generally considered the most technically effective and cost effective specifications for magnetic airborne exploration in the Mega Kalahari. Generally known kimberlites have a "bull's eye" appearance on contoured magnetic data. However, other geological features can also produce confined magnetic anomalies or "bulls eyes" such as;

- confined mafic/ultramafic intrusives eg. dolerite plugs (with an associated gravity high)
- confined basic intrusives eg. Granodiorite dome
- amphibolites
- alkaline diatremes
- magnetic skarns
- magnetic concentrations in granites
- surficial maghemite or magnetite pockets in the regolith
- structural intersections of magnetic dykes
- weathered patches or "holes" in basalt

Gerryts (1967) reviewed airborne magnetometer surveys conducted by De Beers over a number of known pipes at a low flight altitude of 150 m. Some of the known pipes were not detected as they were only marginally anomalous relative to their background, some were missed completely because of their small sizes, and those that returned anomalous values were only moderately more magnetic than background.

Kimberlites have highly variable magnetic susceptibilities, which often frustrates efforts to identify possible pipes during interpretation of airborne data sets. Thus kimberlites may be magnetic highs, lows or even magnetically invisible within a particular host rock. The variable response is partly a function of differential weathering of the pipe or remanent magnetism of the pipe determined by the prevailing magnetic field at the time of intrusion (Fipke et al.,1995). Atkinson (1989) suggests that it reflects the nature of the body itself, in other words its mineralogy, successive intrusive phases, weathering, size and geometry. For example the hypabyssal facies at depth is generally far more magnetic that the upper crater facies. Kimberlites are easily susceptible to weathering which often converts the magnetic minerals to non-magnetic oxides. Ground geophysics becomes useful in following up on airborne targets as a discriminator and mapping tool for outlining pipes.

The magnetic character of the bedrock below the Mega Kalahari in any particular area of interest plays a major role in the efficacy of magnetic surveys in detecting kimberlites. In areas where the Mega Kalahari is underlain by low susceptibility (non-magnetic) host rocks such as platform-sediments/meta sediments (ie. Karoo sediments) or basement granite gneisses, the magnetic response is typically uniform and "quiet". The magnetic response of kimberlites hosted in bedrock of this nature is often distinct and clearly identifiable. This is well illustrated in Figures 38 and 39 where the Okwa and Tsolo hills kimberlitite clusters are immediately visible within the magnetically uniform Karoo sediments beneath 30-70 m of Kalahari cover.

The magnetic response of areas underlain by high-susceptibility rocks such as volcanics (ie. basalt), or by high-grade metamorphic-amphibolite belts is far more variable and "noisy". Basalts in particular can present major difficulties in identifying kimberlites. These rocks present a complex geophysical magnetic response from which to identify unambiguous diatreme-related signatures. Basalts are highly magnetic and tend to swamp the potential magnetic signature of most pipes hosted in this rock type. Furthermore, basalts tend to exhibit a weathering pattern, which produces extensive circular type anomalies (both magnetic highs and magnetic lows or remanents) that camouflage the presence of pipes even further. Figure 40 shows the location of the Khutse kimberlites within a basaltic bedrock. Their presence is far less distinct within the

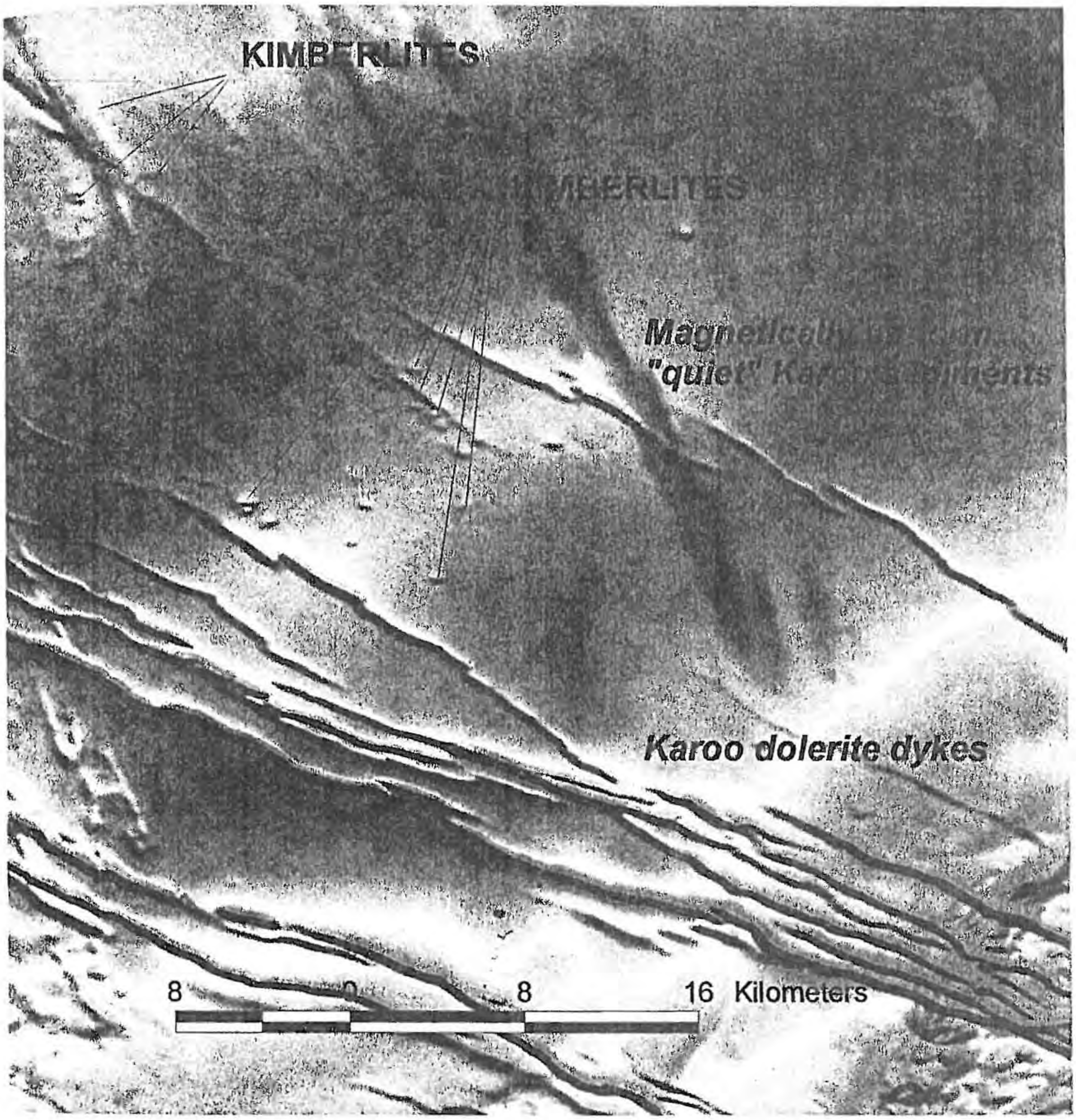
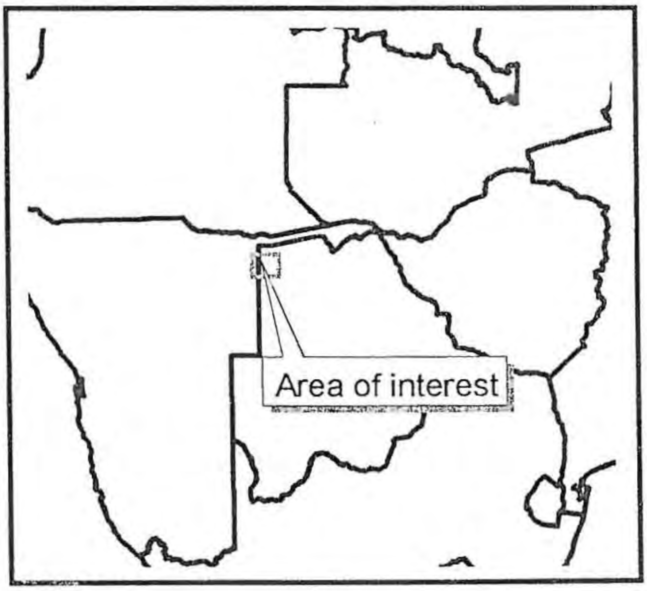


Figure 39. Total magnetic image of the Tsodilo Hills kimberlite cluster, western Botswana. Note the prominent NW-SE Karoo dykes and the magnetically uniform Karoo sediments



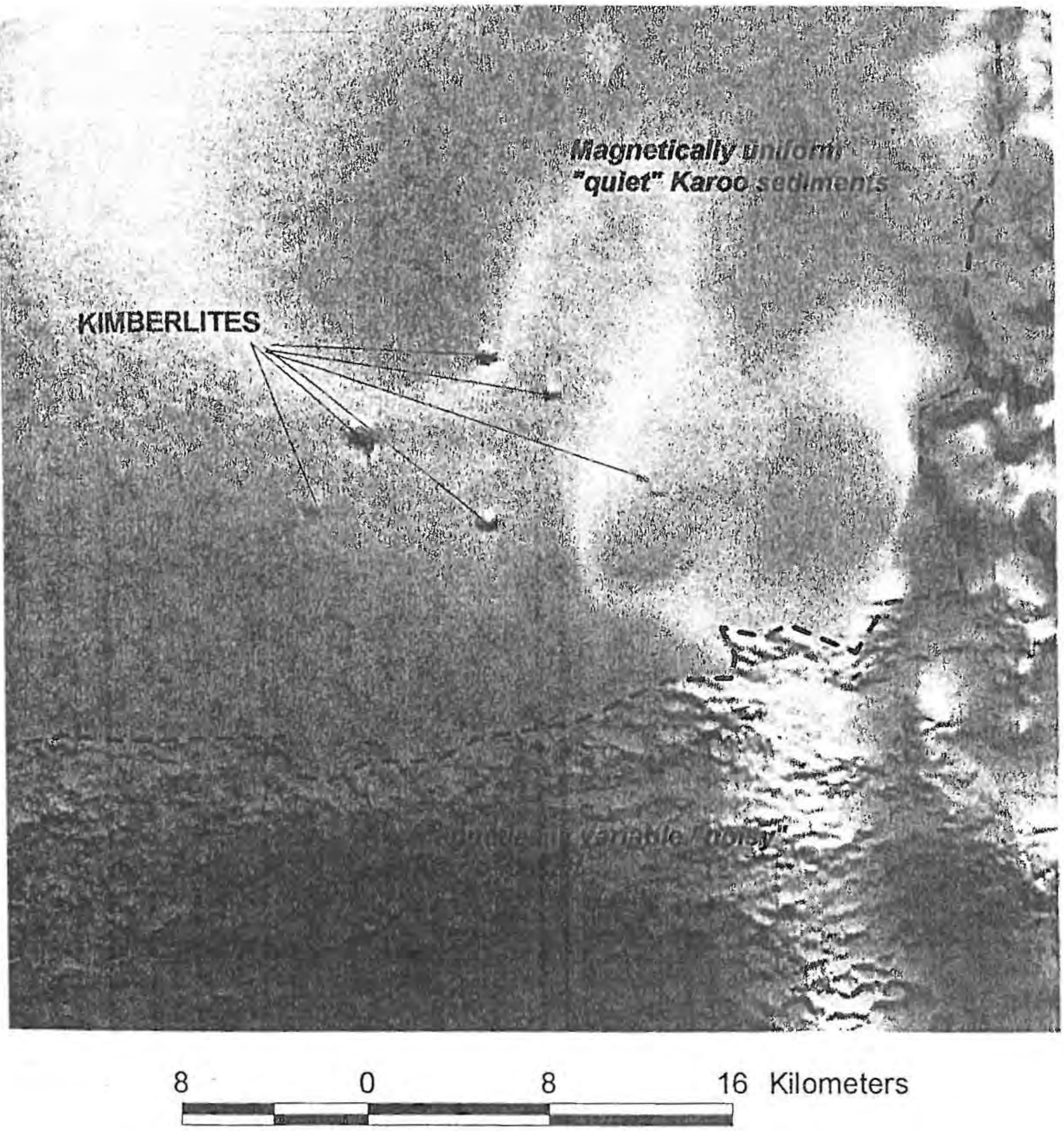
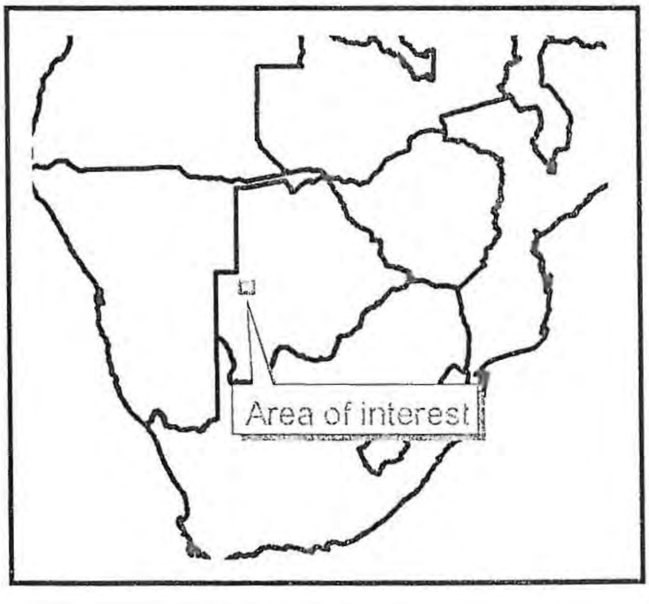


Figure 38. Total magnetic image of the Okwa kimberlite cluster, western Botswana. Note the contrast between the magnetically uniform Karoo sediments and the magnetically varable Karoo basalt



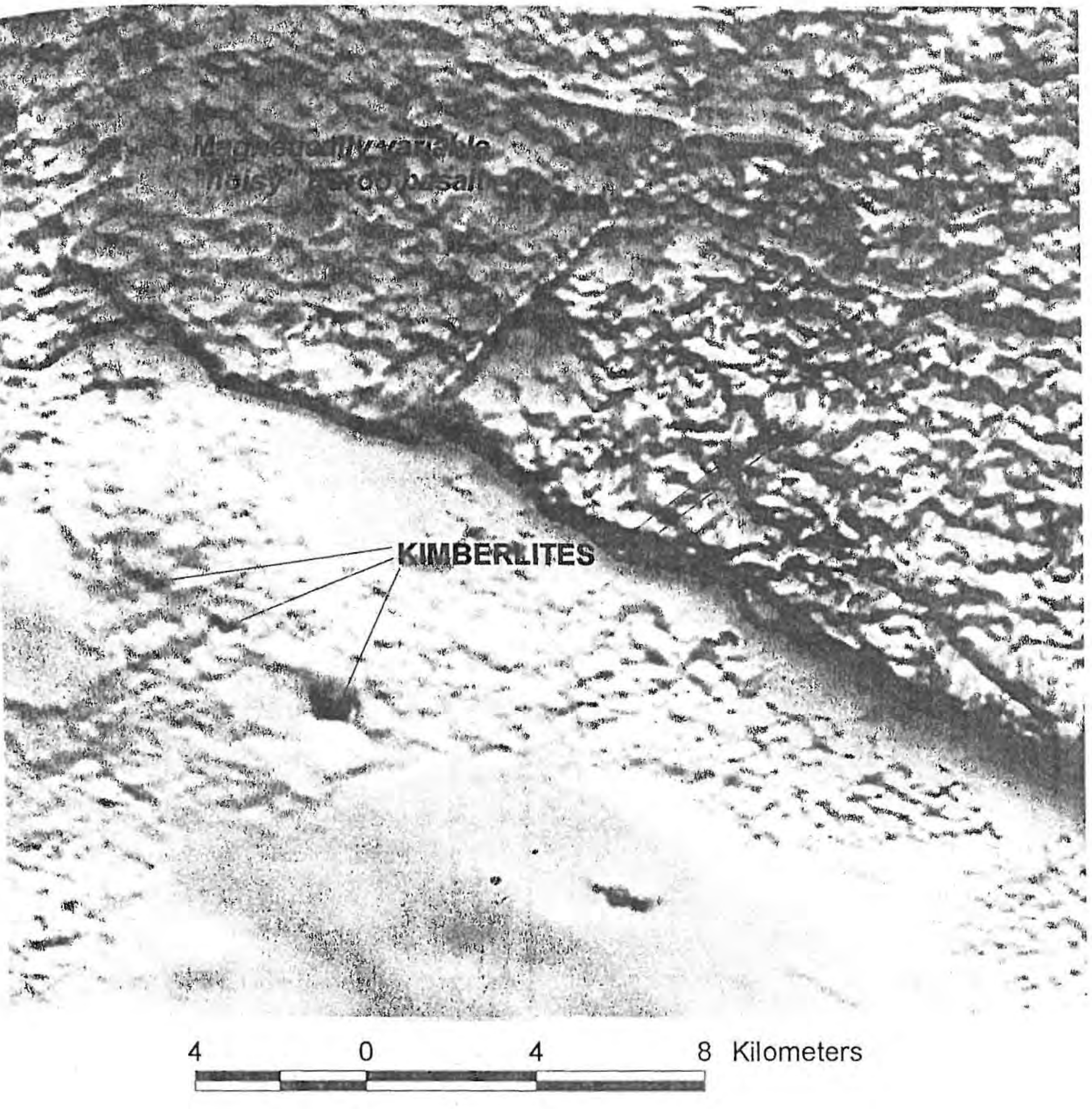
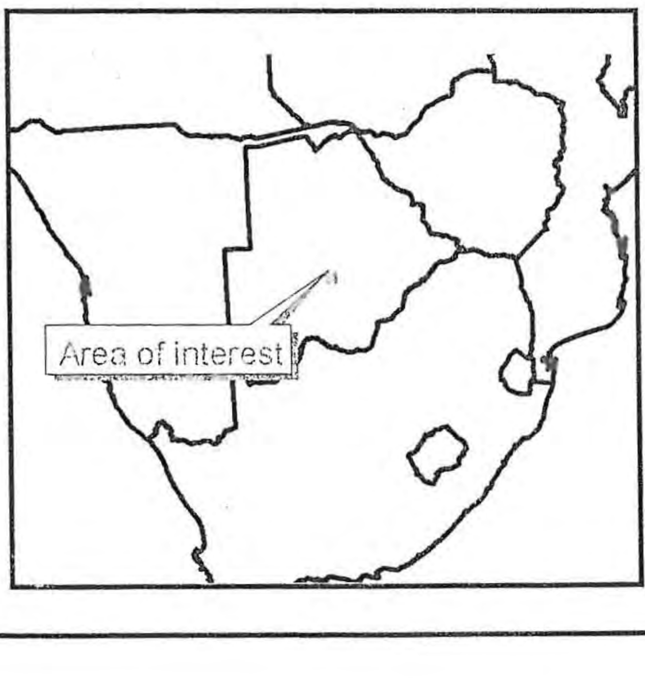


Figure 40. Total magnetic image of the Khutse kimberlites, central Botswana. Note the intense and magnetically variable signature of the Karoo basalts



8.2 GRAVIMETRICS

Density is the physical property that determines gravitational response, in particular the density of the target within its surroundings. Differences in rock density produce small changes in the gravity field of the earth which can be measured.

The practicalities of gravity surveys require that this method is used predominantly as a follow-up technique over specific high interest targets. Generally a negative gravity response is characteristic over kimberlite pipes due to rapid serpentinisation and weathering or the presence of thick crater facies sediments. The lack of response from the deeper, denser, fresh kimberlite is ascribed to the greater depth of burial and shrinking carrot shaped geometry in depth. However, there are many cases where the crater facies of a pipe has been removed by weathering. In these instances an increased gravity response or lower contrast between the hypabyssal and diatreme facies with the host rock can be expected.

In areas of complex geology and buried paleo-land surfaces, gravity is not easy to interpret and provides no easy answers. A key application of ground gravimetrics is to discriminate amongst confined magnetic anomalies produced by kimberlites and other common sources as listed above. Other igneous (non-kimberlitic) intrusives within sedimentary and metamorphic packages should generally display a positive magnetic response which would down-grade them as a potential kimberlite. However, these are general rules and exceptions do exist. A gravity survey over the Okwa kimberlites showed slight gravity highs even with crater facies intact (De Beers unpubl. Report). Gravity will also show features unrelated to diatremes such as paleochannels, small lakes and scour holes in buried paleo-topographies and variations in the subsurface weathering front (Russell, 2002).

Nevertheless, gravity represents a further vital data source in exploration in the Mega Kalahari which must be integrated with other data types. Very recently, Mt Burgess found gravity to be an effective tool for outlining diatremes in the Tsumkwe area of north-east Namibia (Russell, 2002). This is an area of the Mega Kalahari with 50-80 m cover overlying paleo-land surfaces and magnetically noisy Karoo basalts.

8.3 ELECTRICAL METHODS

Electromagnetics provides a means of mapping the conductivity of the uppermost rocks in the earth's crust. Electrical measurements only came into their own when airborne resistivity measurements became viable for primary kimberlite exploration in the mid 1970's. Before that electrical methods were mostly used in groundwork for delineating pipe geometry and the determination to depth of known kimberlite. Airborne electomagnetics (AEM) is commonly used in addition to magnetic surveys.

The target of AEM is the conductive material in the top of the pipe, either the crater facies or the porous weathered diatreme-facies kimberlite. Fixed wing AEM systems have large transmitter strengths and can energise and be sensitive to conductors within a large volume of ground, at a radius of 200 to 300 m from the plane.

Kimberlites are characterised by low resisitivity readings because of the various assemblages of clay minerals from weathered kimberlite. The rapid near-surface weathering of ultramafic kimberlitic material and high porosity of tuffaceous and brecciated parts of the original diatreme facies acts as a conduit for groundwater flow. The clays developed in the weathering process can produce very conductive cover at the top of pipes, especially where crater facies rock is present. The conductive part of the pipe will thus form at the top of pipes.

According to Macnae (1995) other common sources for confined conductive anomalies are;

- weathered amphibolites
- sulphides
- weathered local mafic and ultramafic bodies
- weathered alkaline diatremes
- some magnetic skarns
- overburden or sediment patches
- sediments contained in depressions

However, in areas of thick, generally resistive overburden such as the Mega Kalahari, AEM has limited application. In the survey by Bromley and others (1994) AEM was found to produce the least information due directly to the dominating effect of the thick, resistive Kalahari cover. Although of limited use in areas of thick Kalahari cover (>60 m), AEM should be considered where the overburden is anticipated to be thinner and will therefore have less of an impact on the response. The cost of AEM surveys is generally very high and can be a limiting factor for the exploration program. In areas of thin Kalahari cover, ground EM can provide a useful additional layer of information when following up and discriminating amongst ground targets.

9. REMOTE SENSING

The application of satellite imagery and aerial photography to the direct detection of kimberlites in the Mega Kalahari is limited due to the thick overburden. A few pipes are known to have had an associated surface expression in the form of vegetation differences or topographic anomalies (raised area due to increased bioturbation, circular depression, soil colour tonal differences). These were the Orapa kimberlite in Botswana and the Sikereti kimberlites in north-eastern Namibia. However, in these instances the Kalahari cover was less than 10m. In areas of deeper cover a surface expression of this nature is highly unlikely.

Remote sensing should be used in many aspects of kimberlite exploration in the Mega Kalahari. It can provide useful basic data on regional structure, geomorphologcal and vegetational mapping. Most critically, remote sensing can prove extremely useful in drainage analysis and reconstructing the drainage history of the area of interest. This is necessary when attempting to trace kimberlitic heavy mineral indicators back to source.

10. SUMMARY

Whilst recent investigations into the Mega Kalahri and the new data gathered in the last decade have improved the understanding of this sequence, it has further highlighted the vastness and complexity of the system. The past 20 years have seen a major shift in the interpretation of Kalahari paleoenvironments from conjectures based on evidence from outside the region, to hypothesis based on geomorphological studies and radiometric dating on selected landforms in localities south of the Zambezi. These studies have been sufficient, despite the limitations imposed by the environment, to show that the Mega Kalahari has a different climatic history from other southern hemisphere arid and semiarid regions. Recognition has been made of the complexities of controlling factors and a movement away from unhelpful overgeneralisations towards an increased awareness of the need to identify local and regional factors affecting sediment stratigraphy, geomorphic processes and Mega Kalahari development. Many of the details in the long-term evolution remain skeletal and require refinement but enough data are available for an appropriate framework in which the better documented Late Pleistocene-to-recent environmental changes can be evaluated and interpreted.

A tectonic framework was established during the break-up of Gondwana, which provided many of the controls for the evolution of the Mega Kalahari. The gross morphology of southern and eastern Africa is dominated by the African "Superswell" (Nyblade and Robinson, 1994), a region of relatively elevated terrain. The origin of the Superswell is linked to dynamic topography, caused by the presence of hot, low-density mantle near the core-mantle boundary beneath southern Africa. The plume hypothesis of Burke and Dewey (1973) introduced the term "sub swells" which they believed reflected zones of local mantle upwelling. These sub-swells were first recognised by du Toit (1910,1933) and more widely on the continent by Holmes (1944). The basins defined by the sub swells accommodate Cenozoic sedimentary material. Du Toit (1933) established that these sub-swells play an important role in delimiting drainage basins in southern Africa, indicating a youthful origin or prolonged and continuing uplift. This theory was taken further by Cox (1989) and correlated with characteristic drainage patterns observed in central and southern Africa. These patterns indicate topographic doming associated with plume activity. Cox (1989) describes a remarkable general correspondence between postulated plume locations and the occurrence of the predicted drainage patterns. The Cape-Angola topographic high is deeply eroded, as the towering Jurassic/Cretaceous plutons of Damaraland testify. Fission-track studies indicate several kilometres of erosion during the Cretaceous. As a result the drainage system into the Kalahari basin adopted a more complex pattern as it adjusts to heterogeneous basement rocks.

The Kalahari Basin is one of several large basin like depressions (or sub-swells) bounded by elongated regions of local uplift which characterise the African Plate (Burke, 1996). The Kalahari Baisn is displaced to the western side of the sub-continent by the higher terrain of the eastern margin of the Superswell. The basin is divided into northern and southern sub-basins by the Ghanzi Ridge, along which basement is exposed. Isopachs of the Mega Kalahari indicate that the southern basin has only a single depocentre, while the northern basin has three, one of which lies beneath the Okavango Delta (Gumbricht and McCarthy, 2001). On a regional scale, Figure 41 illustrates a reasonable correlation between the position of the "sub-swell" highs (areas of uplift) and the immediately adjacent thick basinal areas within the Mega Kalahari. These areas of uplift and erosion provided much of the sediment supply to the basin.

One of the most important processes to have contributed to the present character of the Mega Kalahari has been the evolution of the drainage systems. The drainage system in south-central Africa has undergone major reorganisations since the disruption of Gondwana. Isopachs of the Mega Kalahari sequence and a variety of geomorphological features can be used to pinpoint abandoned drainage lines. Continental fluvial systems of the Mesozoic-Cenozoic age reflect river systems which existed prior to and immediately following continental break-up. Overall, the tectonic framework established in southern Africa by the division of Gondwanaland led to the creation of a dual drainage system, with the hingeline acting as a watershed between a coastally-orientated exoreic system and an endoreic system draining into the interior (Thomas and Shaw, 1991). Today the endoreic system possesses only one major river, the Okavango, which terminates in the Kalahari Desert.

During the upper Jurassic to Cretaceous, the Okavango, Cuando and Zambezi-Luangwa rivers formed the headwaters of the proto-Limpopo (Moore and Larkin, 2001). End Cretaceous uplift along the Okavango-Kalahari-Zimbabawe Axis severed the links between the Limpopo and the Okavango, Cuando and Zambezi-Luangwa. This resulted in a senile endoreic system which supplied sediment to the Kalahari basin. However, the uplift rejuvenated the lower Zambezi, initiating headward erosion and progressive capture of the Luangwa, upper Zambezi and Kafue.

Predatory headward extension of the Zambezi is still active, and this river will eventually capture the Okavango.

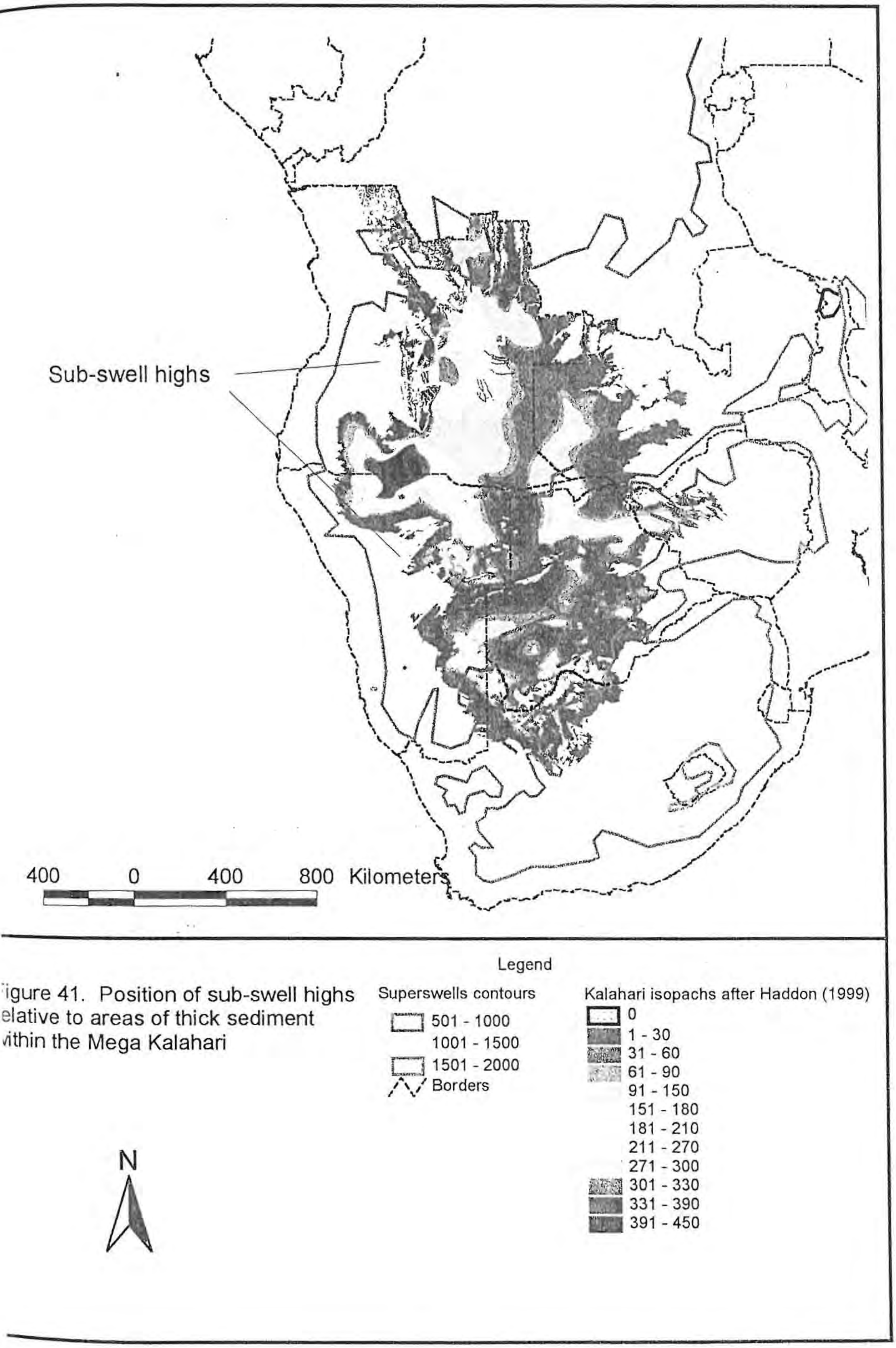
The Mega Kalahari sediments are Cainozoic sediments of the interior of central southern Africa. Analysis of borehole data has permitted an assessment of the depositional setting. Accumulation occurred in a series of contiguous basins, some of which reflect the location of longer periods of Phanerozoic sedimentation whilst others result from more recent periods of tectonic subsidence. Despite gross lithological similarities between many of the individual units from different basins, the fact that the Mega Kalahari beds extend over an area of 2.5 million km^2, affected by tropical to arid climates today, should give rise to caution when depositional sequences and chronologies are considered. This is emphasised by the areal distribution of fossil dune systems formed in the Mega Kalahari Sand (Thomas, 1984), which illustrates the spatial and temporal variability of former conditions and associated aeolian sediment-moving processes in the Mega Kalahari.

Climatic conditions during the entire Cenozoic appear to have fluctuated around a semi-arid mean, with periods of increased humidity, coupled with tectonic activity, providing the impetus for accelerated sedimentation. Thomas and Shaw (1996) cite a wetter period between 16,000-13,000 years B.P. followed by a period of lowering water tables and return to the climatic mean. Landforms such as paleo-lakes, caves, pans, dry valleys and spring deposits are all indicative of wetter episodes. Periods of geometric stability permitted pedogenic calcrete formation, while intervals of increased aridity resulted in the formation of extensive aeolian sand deposits and dune fields.

That the Mega Kalahari sediments have undergone post-depositional modifications is now clearly understood. The duricrust suite is currently being viewed from a process perspective rather than in a simple chronostratigraphic sense with investigations concentrating on local and regional considerations rather than broad descriptive generalisations. The role played by groundwater processes as opposed to climate in their evolution is an important focus of attention. The calcretes within the duricrust suites of the Mega Kalahari are amongst the thickest in the world representing pedogenic episodes in a semi-arid climate during Pliocene to Recent times. Varying degrees of calcrete maturity are related to a number of interdependent factors: time, climate, host materials, carbonate source, geomorphological position, organic influences, sedimentation (or erosion rate) and various localised conditions. The interplay of these parameters has resulted in a in a highly diverse suite of calcrete types. Consequently broad conclusions must be circumspect.

It is possible that calcrete formation has been taking place episodically since the Pliocene (King, 1963).

Extensive dune systems are present throughout the Mega Kalahari mostly in linear form. Cover sands across this sequence are similar in size and composition but differences are present and notable.



igure 41. Position of sub-swell highs elative to areas of thick sediment vithin the Mega Kalahari

10.1 IMPLICATIONS FOR KIMBERLITE EXPLORATION IN THIS MEDIUM

Kimberlite exploration in areas of deep surficial cover and transported overburden is a demanding exercise. The insights gained from reviewing the Mega Kalahari medium provide a regional-scale guide to understanding areas of interest selected for kimberlite exploration. In this type of exploration it is important to understand and prioritise certain criteria within any particular portion of the Mega Kalahari. These criteria are presented and discussed below.

10.1.1 Tectonic framework and Structural background

The Mega Kalahari occupies a series of contiguous Phanerozoic sedimentary basins. The position and controls of these basins can be broadly correlated with the mantle-plume "sub swells" described by Burke and Dewey (1973) and Cox (1989). The location of the area of interest must be considered in the context of these large, regional basins and the characteristics of these basins as well as any similarities or differences that may exist between the basins.

10.1.2 Mega Kalahari Thickness

From the isopach maps it is evident that the Mega Kalahari beds are highly variable in thickness. This cannot be attributed to surface topographical variations, since the sub-continental interior is an area of markedly low relief. Nor is it simply the result of decreasing thicknesses of sediments towards the rim of the interior basin. Much of the variation in thickness can be attributed partly to recent, relatively minor rifting in the central Mega Kalahari and the relief of the pre-Mega Kalahari surface. The thickness of the Mega Kalahri in any area of interest is critical information for the exploration program. The thickness of the cover has a direct influence on the mobility, dispersion and dilution of kimberlitic minerals as well as the sensitivity and resolution of geophysical applications and the modelling of these results. The negative impact on these factors increases with increasing thickness. Thicker cover naturally increases anticipated drilling costs.

Haddon's (1999) Kalahari isopach map provides a good regional guide as to the thickness of the Kalahari. However, from the author's experience, it is of limited use on a local scale. Experience from drilling in the Kalahari has shown major differences to Haddon's isopach map and significant variability between holes on a small scale (5-10km). These depth differences can be as much as 50 m. On a scale of 10s' of km the depth variability can increase further to as much as 100 m between holes. A terrain analysis or geomorphological model of the area of interest within he Mega Kalahari should always be developed. This can be very useful to define and explain the distribution of indicator minerals and potential kimberlite provenances. However, geomorphology becomes less effective when specific drill targets are required.

As the Kalahari is almost entirely a transported medium, geochemistry is of very limited application except in areas of very thin (<10m) cover such as the outermost edges of the Mega Kalahari.

10.1.3 Lithological Variations

Whilst gross similarities in lithological types between the structural basins can be expected caution should be exercised in assuming any major correlations and extrapolations. A well sorted, quartz rich sand of predominantly aeolian origin can be expected as the surface expression of the Mega Kalahari. Magnesium-rich ilmenite and mantle derived garnets are typically the most common kimberlitic heavy minerals targeted and recovered in the semi-arid sandveld environments of the Mega Kalahari. The sands overlie repetitive layers of duricrusts, predominantly pedogenic calcretes. The explorationist must understand calcrete geology and its effect on heavy minerals and geochemical sampling techniques. In sampling programmes it is imperative to monitor variations in proportions of the coarse-size fraction when considering the appropriate size fraction with which to capture heavy mineral samples.

Whilst drilling is typically a final tool in exploration programmes it may be a prudent step in the middle to late stages in programmes within the Mega Kalahari. This is particularly relevant where local information on Kalahari thickness and lithology types is scarce or absent. It may also reveal the presence of any secondary, fluvio-gravel horizons rich in heavy minerals.

10.1.4 Pre-Kalahari Bedrocks

Base of Mega-Kalahari bedrock type can have a major affect on geophysical applications. For reasons discussed above, areas of the Mega Kalahari underlain by rock types with high magnetic susceptibility and variability (Karoo basalts, metamorphic gneiss terranes) present considerably greater difficulties for kimberlite identification than areas underlain by magnetically "quiet" bedrock types. In magnetically uniform areas, magnetics is a highly effective tool. Gravity is a useful follow-up technique and discriminator for discrete magnetic targets. Electrical methods are best applied in relatively thinner parts (< 60 m) of the Mega Kalahari. The nature of the bedrock type is important in constraining certain physical properties of the rock for geophysical modelling of data generated by magnetic, gravity and electrical methods.

10.1.5 Drainage and Aeolian Patterns

It is imperative to understand the existing drainage patterns in an area of interest and how these relate to the evolution of the Mega Kalahari endoreic drainage systems since the disruption of Gondwana. These drainage patterns may allow reconstruction of kimberlitic heavy mineral indicator dispersion histories, which are vital to the kimberlite explorationist. The distribution histories are complicated in many cases by a combined fluvial-aeolian history. Primary processes of deposition and subsequent modifying process are difficult to distinguish. Fluvial and aeolian processes in the Kalahari, whether alone or in combination, are capable of transporting heavy minerals across hundreds on kilometres in the Mega Kalahari environment and creating anomalies attributed to non-primary sources.

Extensive paleo-dune patterns within the Mega Kalahri provide indications of paleo wind directions, which must be factored into the transport history of heavy mineral anomalies. Current prevailing wind patterns should not be ignored and may currently be influencing the dispersion of heavy minerals.

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