The development and environmental significance of the dry valley systems (*mekgacha*) in the Kalahari, central southern Africa

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VOLUME ONE

Thesis submitted for the degree of Doctor of Philosophy, September 1992

For Mum

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Abstract

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The dry valley systems in the Kalahari of southern Africa are traditionally considered to have developed as a result of past fluvial activity. However, it has also been suggested that erosion by groundwater processes (sapping and deep-weathering) had an important role in development. This thesis aims to establish the relative role of each of these process areas in *mekgacha* evolution using a combined geological and geomorphological approach.

The study area is restricted to the valley systems of Botswana, eastern Namibia and the Northern Cape Province of South Africa, which can be subdivided into exoreic and endoreic systems directed towards the Orange River and the continental interior, respectively.

Field studies, analyses of remotely-sensed imagery and a consideration of network orientation identify evidence for the role of both fluvial and groundwater processes in valley development. However, whilst both groups of processes have operated, the importance of each is suggested to have varied both spatially and temporally.

Fluvial processes are indicated by sequences of sediments, relict channels and terrace levels, and appear to have operated most recently. Sapping processes are implied in the formation of certain valley systems, primarily from morphological properties and the presence of relict spring lines. Deep-weathering processes are implicated from the close parallelism of many valleys with geological structures now buried beneath thicknesses of Kalahari Group sediments. Borehole records also indicate deep-weathering of bedrock beneath valleys developed above fracture zones, which is suggested to have operated over the longest timescales.

Thin-section studies of duricrusts from valley flanks, together with duricrust profiles and records from lithological boreholes, indicate the role of groundwater in their formation. Results suggest an intrinsic link between duricrust formation and valley development. Geochemical and thin-section analyses of duricrusts further suggest that previous considerations of the palaeoenvironmental significance of Kalahari silcretes based upon TiO_2 levels may not be wholly appropriate.

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Prologue

This thesis concerns the factors involved in the development of the dry valley networks of the Kalahari. The mythology of the G/wi San of central Botswana includes an account of the origins of the Okwa valley (Silberbauer, 1981 p.96; Main, 1987 p.183), which they believe was formed by the "evil" supernatural being G//awama.

Whilst hunting one day to the west of the Kalahari near Gobabis, G//awama was bitten in the leg by a python. The bite was bad and he became feverish and very thirsty, so he headed east in search of water, towards the Boteti River. G//awama dragged his injured leg as he walked, his trailing foot gouging out the present Okwa valley. Wild animals harassed him *en route*, taking advantage of his weakness to avenge some of the evil things he had done to them in the past. This caused his course to waver, and as a result created the twists and turns present in the valley. The fever induced by the python venom made him nauseous, and he vomited frequently. The dried vomit formed the calcrete- and diatomite-floored tributary valleys of the Okwa (which itself has a generally sandy bed). As G//awama neared the waters of the Boteti he found new strength and increased his walking speed, thus dragging his leg less heavily. He was eventually able to walk almost upright, dragging his leg only slightly; this explains why the course of the Okwa almost disappears as it nears the Boteti River and the Makgadikgadi Depression.

Part 1

Background and aims

Chapter 1
Outline of research problem

1.1 Introduction and background

Research into the geomorphology of dryland valley systems, especially valleys that are considered "fossil" or ephemeral, invariably attributes formation to erosion by fluvial activity during times of more humid climatic conditions. As such, valley systems which exist in presently arid or semi-arid areas are traditionally considered to represent large scale shifts in climate and on this basis are accorded palaeoclimatic significance. This is the case for valley systems in the tectonically stable interior of Australia (e.g. Van De Graaff *et al.*, 1977; Rust and Nanson, 1986; Pickup *et al.*, 1988; Tapley, 1988) where much of the internal drainage consists of ephemeral streams which occasionally discharge into salt lakes. The valley systems are usually interpreted as indicating considerably wetter periods during the Eocene-Oligocene, and there is some suggestion that valleys may have existed since the end of the Permian (Gale, 1992).

The same viewpoint has existed for the dry valley systems (termed *mekgacha* in SeTswana, the language of Botswana) of the Kalahari Desert in central southern Africa since the records of Andersson (1856 p.374) and Baines (1864 p.119) who independently crossed the Hanchai Valley in western Botswana. The seminal work of Grove (1969) reinforced this perception (which exists to the present day; cf. Heine, 1982) following the discovery of stone tools on the flanks of the Okwa Valley, suggesting that they "might conceivably have been left there by men who lived alongside the Okwa when it was a sizeable river" (Grove, 1969 p.202). With the exception of the hydrogeological study of the Kalahari by Boocock and Van Straten (1962), no specific study of the dry valley systems of the Kalahari had been undertaken until the work of Shaw and De Vries (1988). Reference to the form of many valleys had, however, been made by a number of authors, in geomorphological, geological and anthropological contexts (e.g. Jennings and Crockett, 1961; Crockett and Jennings, 1962, 1964, 1965; Grove, 1969; Yellen and Lee, 1976, 1984; Wright, 1978; Lee, 1979, 1984; Helgren and Brooks, 1983). In contrast to the traditional view of valley development, Shaw and De Vries (1988) suggested that, on the basis of the morphology, duricrust suite and underlying geology associated with four valleys in southeastern Botswana, *mekgacha* had formed predominantly by groundwater sapping, deep-weathering and groundwater erosion along fractures.

Palaeoenvironmental studies in the Kalahari have, until recently, lent further credence to the development of *mekgacha* during former wetter periods. Such studies have tended to be dependent upon geomorphological lines of evidence, primarily because of a lack of sites suitable for the preservation of faunal and palynological material (Thomas, 1987*b*; Deacon and Lancaster, 1988). Studies of extensive vegetated dune systems (e.g. Lancaster, 1981; Thomas, 1984*a*) have indicated past extensions of the arid zone. Fluctuating wetter and more arid periods are inferred from sediments and features of caves (e.g. Cooke and Baillieul, 1974; Cooke, 1979*c*, 1980, 1984; Brook *et al.*, 1990) and pans (e.g. Lancaster, 1978*a*; Goudie and Thomas, 1985, 1986), with shorelines in ancient lake systems (e.g. Cooke, 1980; Shaw, 1985*a*) used to infer former wetter climates. Whilst cave studies in the Kalahari provide unequivocal evidence for more humid past climates, recent work has cast doubt on the simple humid to arid shifts envisaged by other landform studies. This is primarily due to increased awareness of the range of factors which can influence landscape-forming processes. In particular, the significance of vegetated sand dunes has been questioned in light of recent studies of the role of vegetation in dune development (Thomas, 1988*c*, 1992; Thomas and Tsoar, 1990; Thomas and Shaw, 1991*b*). The data indicating higher water levels in Middle Kalahari palaeolakes cannot now be viewed solely as evidence of increased rainfall, but needs to be interpreted with regard to neotectonic activity and complex shifts in regional drainage patterns (Shaw and Cooke, 1986; Shaw and Thomas, 1988, 1992). Additionally, the identification of the role of groundwater, as opposed to simply deflation processes, within the formation of Kalahari pans suggests a need for detailed hydrological interpretation of evidence from pan sites (Lancaster, 1986*b*; Thomas *et al.*, 1993). The overall picture of environmental change emerging from the landforms of the Kalahari is one of much more subtle shifts in climatic parameters, as opposed to the simple changes in rainfall amounts previously envisaged (Klein *et al.*, 1991; Thomas and Shaw, 1992; Nash *et al.*, 1993).

Within this context, the suggestion by Shaw and De Vries (1988) that Kalahari *mekgacha* developed by groundwater erosion processes has important implications for the interpretation of climatic shifts in the region. The role of groundwater processes, particularly groundwater sapping, in valley development is well documented from other semi-arid environments, where major contemporary valley networks are developing as a result of the process (e.g. Pieri *et al.*, 1980; Laity and Malin, 1985; Howard *et al.*, 1988, Baker, 1990). Valleys developed by groundwater sapping evolve primarily by headwater erosion as a result of weathering at a site of groundwater emergence. The resultant landform has a characteristically flat valley floor, steep sidewalls and an amphitheatre head (Gomez and Mullen, 1992). Whilst less well documented, *in situ* deep-weathering processes are also implicated in the evolution of many shallow southern African valleys associated with dambos (McFarlane, 1989; Boast, 1990). Dambo formation by deep-weathering involves the progressive lowering of the valley floor and flanks due to solution and removal of valley floor materials, and is considered to have operated during periods when fluvial activity was absent (McFarlane, 1989). The environmental significance of groundwater processes is that, from contemporary observations, valley development can occur under a semi-arid climate without a need for increased rainfall, given otherwise suitable environmental conditions. Indeed, Howard *et al.* (1988) suggest that greater rainfall and groundwater outflow may inhibit the operation of sapping processes by hindering the accumulation of minerals and salts which act as important mechanical weathering agents at seepage sites.

Sapping is closely linked to geological factors controlling the height of the regional water table and zones of emergence (Higgins, 1984). Fractures and other geological structures are also considered vital in controlling the location and intensity of deep-weathering processes in dambo formation (McFarlane, 1989). In order for sapping processes to operate, significantly higher water tables would be required than exist in the Kalahari at present. However, the control of water table levels may be influenced more by tectonic or anthropogenic rather than climatic changes, and it is also possible that Kalahari aquifers may have been supplied by transfers of groundwater from beyond the Kalahari region (Farr *et al.*, 1981). The role of geology in both sapping and deep-weathering processes may be of particular significance, given the recognition by Summerfield (1985*a,b*) and Thomas and Summerfield (1987) that the large-scale tectonic setting of a region needs careful consideration in any model of long-term landscape evolution. The tectonic setting of the Kalahari has been characterised by periods of regional uplift since the Mesozoic. Should fluvial activity have been the main factor in valley development, then uplift may have had a significant effect upon drainage systems. Additionally, it is also likely to have influenced water table levels and, by implication, the potential operation of groundwater processes. However, the extent of this effect is difficult to assess due to the combined effects of the cessation of major groundwater recharge at around 12,500 years BP (De Vries, 1984) and lowering of regional watertables due to the contemporary use of groundwater resources (Thomas and Shaw, 1991*a*). What is certain is that fluvial activity is presently inactive in the majority of *mekgacha*, with the exception of comparatively rare flood events. Furthermore, on the basis of a general absence of seepage sites within valleys, the contemporary action of groundwater processes may be extremely limited, primarily due to the depth to regional water tables.

1.2 Aims and approach of thesis

In light of the preceding discussion, the main aim of this thesis is to establish and evaluate the evidence for the mode of development of Kalahari *mekgacha*, and more specifically the role of groundwater and fluvial erosion (either by ephemeral or perennial rivers) in their formation. As such, this thesis concentrates primarily upon valley as opposed to channel characteristics.

Three main approaches are possible for the investigation of this problem. Firstly, evolution could be viewed from a palaeohydrological perspective, taking sedimentary and morphological data from specific locations into account in order to assess possible flow conditions represented by these various lines of evidence (see Gregory, 1983, for a variety of examples). Secondly, past hydrology could be assessed by investigating changes in the drainage network characteristics, and attributing changes in network density, structure and composition to variations in hydrological regime (Gardiner, 1983). Thirdly, following the possible significance of the relationship between duricrusts, geological structure and *mekgacha* suggested by Shaw and De Vries (1988), a more geological approach could be adopted.

The scope for the application of the two palaeohydrological techniques in the context of the Kalahari is extremely limited, for a number of reasons. The use of sedimentary data to suggest past hydraulic properties of a channel within a valley necessarily requires either exposures of sediment or access to excavating or drilling equipment. In the majority of *mekgacha*, exposures of sediment are extremely limited, if present at all, primarily because of the lack of a channel or due to burial by an extensive cover of Kalahari Sand on the valley floor. Where suitable exposures do occur, they are invariably in peripheral settings (see chapter 5) and the data derived from them may be of only limited applicability. However, there are a number of major disadvantages to a sediment-based palaeohydrological approach. Firstly, the hydraulic properties which may be derived from sediments are for specific former flows, irrespective of environment (Gardiner, 1983). Secondly, such an approach cannot identify the influence of groundwater in valley development, and most importantly, it identifies changes specific to channel and not valley properties and represents only depositional environments, not erosive ones.

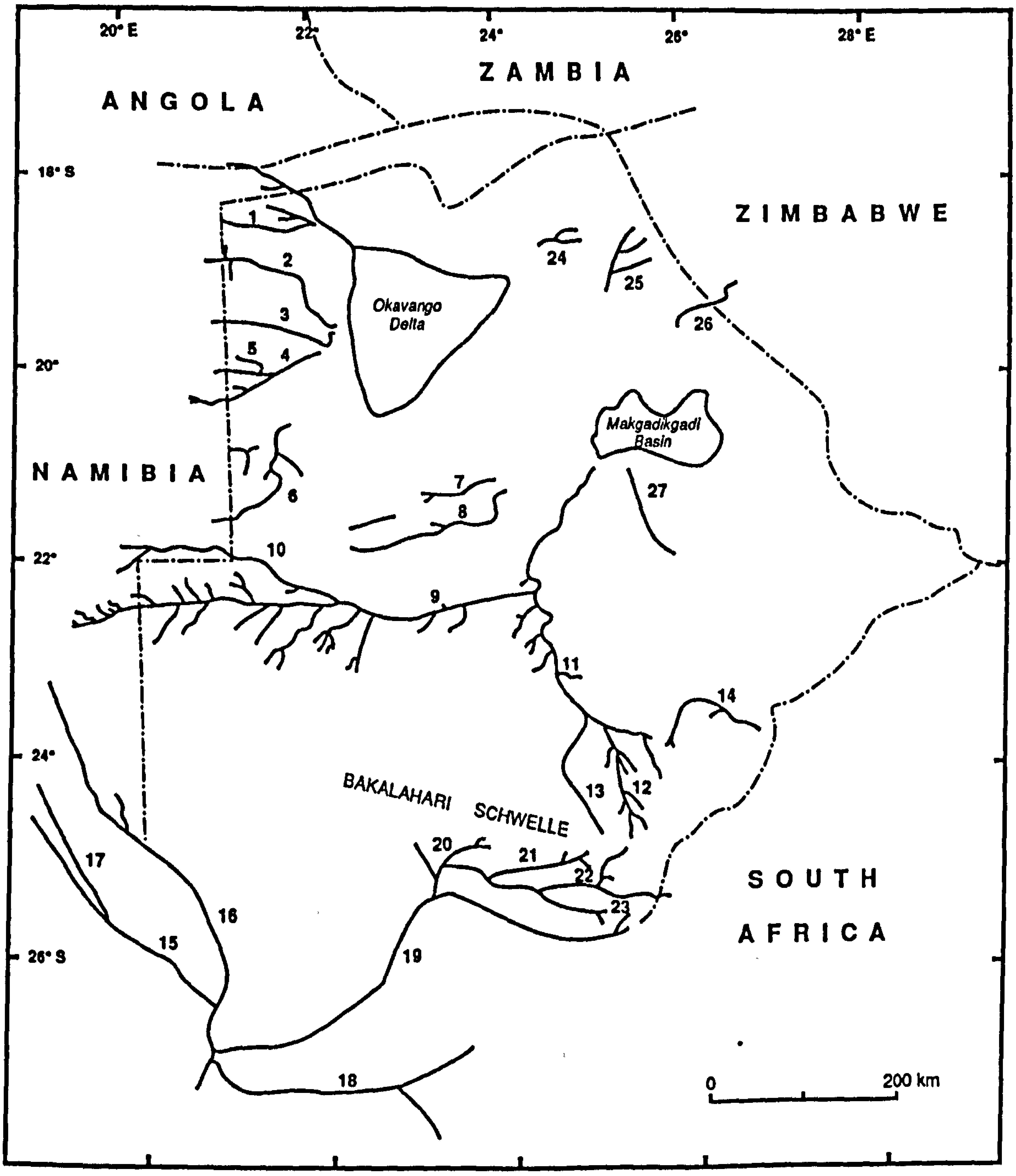
The presence of a Kalahari Sand cover also precludes consideration of changes in the total drainage net because many valley courses are either partially or completely buried by aeolian sediments. Such an approach would require the adoption of remote-sensing or seismological techniques to identify the extent of buried valleys prior to any assessment of network characteristics. However, an approach involving a consideration of network structure, specifically considering the valley network and not requiring details of drainage density, would be appropriate and applicable in evaluating *mekgacha* formation.

As a result of the limited applicability of palaeohydrological techniques in the Kalahari, the approach adopted within this study is primarily, but not exclusively, a geological one. This choice is based upon the available evidence attainable from *mekgacha* which is predominantly in the form of exposures of duricrusts, with few valleys containing channels. Thus, three main approaches to the evaluation of modes of valley development are used. Firstly, variations in valley morphology are considered from field investigations and studies of remotely sensed imagery, including the use of sediment-based palaeohydrological techniques where channel deposits are present. Secondly, the morphological, petrological and geochemical properties of duricrusts exposed within and occurring beneath *mekgacha* are considered. This approach is based upon the recognition of the potential relationship between valleys and duricrusts identified by Shaw and De Vries (1988), and particularly aims to identify the relative timing of the formation of duricrust suites in relation to valleys. Finally, the general structural characteristics of *mekgacha* networks are considered, with specific regard to valley orientation. This technique is used to identify evidence for control of valley location and orientation by structures within Precambrian and Palaeozoic-Mesozoic bedrock now buried by Kalahari Group sediments. The three main approaches include a wide range of scales, from the microscopic analysis of duricrust samples to the megascale analysis of network characteristics.

1.3 Area of investigation

The drainage of the Kalahari can be considered to consist of three components (Shaw, 1989). These are the perennial systems of northern Botswana, including the rivers and swamps of the Okavango, Zambezi and Kwando-Linyanti-Chobe, the ephemeral "sand rivers" of the eastern Kalahari periphery (Wikner, 1980; Nord, 1985) and extensive *mekgacha* networks.

The main dry valley systems of the Kalahari are indicated on figure 1.1, with studies covering most of the systems within Botswana in addition to South African and Namibian valleys. The valleys can be subdivided into "Southern Kalahari" systems directed exoreically and ultimately connecting with the Atlantic via the Molopo Valley and Orange River, and endoreic "Middle Kalahari" systems directed towards the Makgadikgadi Depression or Okavango Delta. Field investigations concentrated primarily upon the endoreic systems.



MEKGACHA NETWORKS

1 NCAMASERE	8 DECEPTION	15 AUOB	22 MOSELEBE
2 XAUDUM	9 OKWA	16 NOSSOP	23 SEKHUTANE
3 QANGWADUM	10 HANEHAI	17 ELEPHANTS	24 GHAUTAMBI
4 EISEB	11 MMONE/QUOXO	18 KURUMAN	25 NUNGA
5 GCWIHABEDUM	12 LETLHAKENG	19 MOLOPO	26 LEMEMBA
6 GROOT LAAGTE	13 NALEDI	20 MABUASEHUBE	27 LETLHAKANE
7 ROOIBRAK/PASSARGE	14 SEROROME	21 UKWI	

Figure 1.1: The dry valley systems of the Kalahari.

Field studies were undertaken in 1989 and 1990 and included detailed investigations of the Okwa, Mmone/Quoxo, Auob, Kuruman and Serorome valleys, whilst the Ncamasere, Xaudum, Groot Laagte, Hanehai, Nossop, Molopo and Rooibrak/Passarge systems were studied at a reconnaissance-level. Valley systems not studied in the field include the Ghautambi, Nunga and Lememba near the Zimbabwe-Botswana border, the Deception and Letlhakane valleys in central Botswana, the Gcwihabedum and Qangwadum in northwestern Botswana and the many valleys in northeastern Namibia which connect with the Okavango River in Namibia and Angola (not shown on figure 1.1). The specific locations for detailed studies of duricrusts, and the systems included within the analysis of network structure and orientation, are given in chapters 6 and 7 respectively.

A duplicate of figure 1.1 is also included as a folded insert bound into the back of this thesis to assist in the location of individual valley systems during the course of reading the text.

1.4 Organisation of thesis

This thesis is sectioned into three parts. The first contains background information in two chapters; a review of the literature concerning valley development by fluvial and groundwater processes (chapter 2), and a summary of the past and present environment of the Kalahari (chapter 3). The chapter concerning valley development includes a summary of the characteristics of arid zone fluvial systems, details of interior drainage networks outside the Kalahari and information regarding groundwater processes. Chapter 3 places the study in its geological, tectonic and geomorphological context, with an emphasis upon the nature of environmental changes in central southern Africa.

The majority of the thesis is contained in Part 2; hypotheses, methods and analysis. The four chapters include a summary of the possible modes of *mekgacha* development (chapter 4), followed by full accounts of each of the three main methods (field studies, duricrust and network orientation analysis) used in their evaluation. These chapters (5 to 7) include reviews of background information, methodology and full details of results, together with discussion where appropriate. Although chapters 5 to 7 include summary conclusions, Part 3 contains the overall discussion of results and evaluates their significance in determining the mode of *mekgacha* development (chapter 8). The thesis concludes with a summary of the main points resulting from this study.

1.5 Valley nomenclature

It should be noted that there are various regional words meaning "valley" within the languages of southern Africa. These include *mokgacha* (plural *mekgacha*) in SeTswana, *laagte* in Afrikaans, and *omuramba* (plural *omiramba*) or *dum* in the various languages of northwestern Botswana and northeastern Namibia. The terms "valley" or *mokgacha* (and appropriate plurals), as opposed to other regional terms, are used throughout the text of this thesis when discussing valley systems in general.

When referring to specific valleys, the nomenclature becomes more complex, particularly for valleys to the west of the Okavango Delta in northwestern Botswana and northeastern Namibia. It would be tautologous to refer to the "Groot Laagte Valley", "Qangwadum Valley", "Gcwihabedum Valley" or "Xaudum Valley", since the words *dum* and *laagte* mean valley, and these systems are referred to simply as the Groot Laagte, Qangwadum, Gcwihabedum or Xaudum. The valleys in northeastern Namibia are indicated on topographic maps with the prefix "*omuramba*" (hence the Omuramba Eiseb or Omuramba Telabashe). These are referred to simply as the Eiseb Valley or Telabashe Valley for consistency with the valleys of central and southern Botswana which have no word meaning "valley" commonly attached to them.

This problem is further compounded since many valley systems have different names and/or spellings in different parts of their course (Stigand, 1923). This is particularly the case where valleys are continuous across international boundaries e.g. the Okwa Valley in Botswana is known as Chapman's River in Namibia. The various regional names for individual systems are referred to in chapter 5, but it is generally the SeTswana name which is used. For overall consistency, the nomenclature used throughout this thesis is that of Thomas and Shaw (1991*a*), as shown in figure 1.1 and in the foldout at the back of this volume.

Chapter 2
Methods of dryland valley development

2.1 Introduction

The study of the development of valleys within dryland areas, and the nature of the channels they contain, has received comparatively little attention in the geomorphological and hydrological literature. The lack of information becomes particularly apparent when compared with the vast accumulation of data for temperate latitudes. The reasons are numerous, but two factors appear to dominate (Reid and Frostick, 1989). Firstly, studies in arid and semi-arid areas have tended to concentrate upon landforming processes peculiar to those environments, namely aeolian processes. Secondly, due to the unpredictable spatial and temporal distribution of rainfall and rainfall events in such environments (Graf, 1988), the collection of data on fluvial processes is very costly both in terms of finance and time. As a result, most fluvial process studies have tended to centre upon temperate environments.

Such studies that do exist, however, highlight two major process areas of importance to the formation and evolution of dryland valley systems. These can be broadly grouped into fluvial and groundwater processes, which can be considered to act at the surface and sub-surface respectively (Dunne, 1980; Gomez and Mullen, 1992). The two suites of processes are closely interlinked, with groundwater supplying perennial streams by base flow and potentially influencing channel bedforms by positive seepage, and fluvial activity providing recharge to groundwater supplies (Keller and Kondolf, 1990). As such, runoff and groundwater erosion form end-members of a process spectrum, with valley development potentially influenced by both processes acting alone or in combination at different spatial and temporal scales.

This chapter considers valley development by both fluvial (section 2.2) and groundwater processes (2.3) in dryland environments. Section 2.2 contains a brief review of the main features of dryland fluvial systems together with examples of long-term palaeodrainage studies from northeastern Africa and Australia. Section 2.3 includes accounts of both groundwater sapping and *in situ* deep-weathering processes, together with brief consideration of the role of deep-weathering and water table fluctuations in pan (playa) development.

2.2 Fluvial processes and valley development

Recent studies in fluvial geomorphology have tended to address the relationship between channel processes and form, in contrast to early studies which concentrated more upon landscape evolution than the processes which generate change (e.g. Davis, 1899). Few authors consider valley evolution *per se* (Kelsey, 1988 being a recent exception), primarily because the timescales over which such evolution occurs are beyond the scope of process studies or involve hazardous extrapolation of erosion rates.

Clearly, the most important influences upon valley form and development are the factors which control erosion and deposition, generating scour and fill within a valley. The primary controls are rainfall and its conversion into runoff interlinked with vegetation and soil characteristics, slope angles and slope lengths, all of which vary both in time and space and need to be viewed within a potentially changing environmental setting (Graf, 1988). Whilst recognising their importance, to provide a full review of the complex relationships between these factors and their effect upon dryland fluvial systems is beyond the scope of this thesis (see Cooke and Warren, 1973; Graf, 1988; Reid and Frostick, 1989 for reviews). As such, this section contains a brief overview of fluvial processes operating within drylands, concentrating particularly upon contrasts between humid and semi-arid fluvial systems and the response of systems to environmental change. Due to the predominantly low relative relief of the Kalahari, discussion centres upon environments away from montaine areas, and thus omits much of the literature stemming from badland, alluvial fan and pediment environments (reviewed by Campbell, 1989 and Harvey, 1989).

2.2.1 Fluvial processes in dryland environments

(a) Rainfall and flood events

The physical mechanisms of entrainment and sediment transport are common to all fluvial systems, regardless of environment, with the extent to which particles are eroded or deposited depending primarily upon the availability and capacity of stream discharge above a critical threshold to move material (Graf, 1983). There are two main differences between dryland and temperate fluvial systems; the frequency with which such discharges occurs, and the production and propagation of the flood wave. The magnitude and frequency of floods are of great importance since flood events perform many of the changes in channel morphology in such streams (Stengel, 1964, 1966; Stear, 1985; Hereford, 1986). This is of particular significance in semi-arid areas where flow may occupy a river channel for less than 1% of the time (Reid and Frostick, 1989).

The differences between floods and flood propagation in temperate and dryland fluvial systems can be largely attributed to differences in the nature of rainfall events. Precipitation in drylands is dominated by four main processes; frontal activity, convection, tropical storms and orographic effects. Of these, convectional and frontal rainfall are the most important for runoff generation (Graf, 1988). Convective storms are characteristically of short duration (i.e. < 1 hour), high intensity and limited spatial extent (Sharon, 1972), and as such, affect only small areas of drainage basins. If runoff does occur, it is possible for different areas of a basin to be affected by different storms. Frontal rainfall is comparatively rare in drylands since most dryland areas are located beneath high-pressure systems (Graf, 1988), but when it occurs it may lead to the production of longer periods of rainfall of greater significance in runoff generation.

The major effect of the limited spatial extent of rainfall in drylands is upon flood characteristics. Graf (1988) notes that peak discharges in drainage basins in dryland areas are larger than for similar sized basins in temperate areas given the same magnitude of rainfall. This is, of course, in part due to differences in vegetation cover and soil characteristics (see below). However, it is possible that different parts of a basin will contribute water during different rainfall events, making the prediction of flood propagation problematic (Reid and Frostick, 1989).

Essentially, four non-mutually exclusive types of flood can be distinguished in dryland areas, dependent upon rainfall and basin characteristics and the location of source areas; these are flash floods, single- and multiple-peak events and seasonal floods (Graf, 1988). Seasonal floods occur primarily in rivers which flow through dryland areas, having originated elsewhere. The other three types are dependent upon the nature and duration of rainfall. Flash floods incorporate a steep rising flood limb, with a minimum time of approximately 10 minutes between the onset of flow and the peak discharge (Schick, 1970), a rapid flood recession and a generally short duration, commonly in the order of a few hours (Reid and Frostick, 1989). Single-peak floods are of longer duration, and usually the product of frontal precipitation, whilst multiple-peak events are characteristic of storms which have stalled over an area or where different parts of a basin contribute water at different times due to migration of the storm front (Graf, 1988).

A further characteristic of dryland fluvial systems is the extent to which water is lost by seepage into the channel bed and banks and by evaporation from the channel. Transmission losses by seepage are primarily due to the extreme dryness of channel bed and bank materials which results in water being drawn downwards into the bed. Losses to evaporation are of less significance due to the relatively short duration of most flood events, but can be important in major throughflowing rivers originating beyond dryland areas. The main effect of transmission losses is that flood discharge actually decreases downstream unless input is provided by tributary inflow. As a consequence, both sediment transport and changes in channel geometry are influenced (Reid and Frostick, 1989). In particular, approaches to hydraulic geometry which utilise the continuity equation become largely redundant.

(b) Sediment erosion and transport

Valley development is dependent upon erosive processes operating both within the channel and on adjacent hillslopes. In simple terms, the erosion of slopes is controlled by rainfall, vegetation, soil and slope characteristics which influence the type and quantity of material available for transport within the fluvial system (Kirkby and Morgan, 1980). In many dryland areas soils are typically of low cohesivity with limited vegetation cover, and thus can potentially supply large quantities of sediment to the fluvial system, depending upon the occurrence of erosive rainfall events.

With regard to the fluvial system, Reid and Frostick (1989) note that ephemeral streams are capable of moving vast quantities of sediment. Erosion during flood events is typified by removal of bed material during channel scour, with an equal amount of fill occurring during flood recession. Thus, sediment moves in pulses according to the duration and extent of each flood event (Graf, 1988). Sediment is transported both as suspension and bedload, but neither has been extensively measured due to the nature of dryland floods. Suspended sediment concentrations which have been recorded from ephemeral channels are anywhere from 6 to 4500 times higher than for perennial counterparts (Reid and Frostick, 1989). Bedload is potentially even more important in ephemeral channels, which are typically burdened with large quantities of sediment. Studies of dynamic bedload transport are limited, primarily due to the instability of stream beds during floods. As such, most studies have concentrated upon post-flood static bed forms to assess the erosive power of events. For example, Schick *et al.* (1987) have traced tagged clasts in gravel-bed rivers and have discovered evidence for scour and fill, and interchanges between buried and exposed clasts related to the depth of scour during each event.

(c) Process-form relationships

(i) Ephemeral river channel geometry and sediments

The form of ephemeral channels is described by a number of authors as being atypically wide for its basin size, with a subdued, almost planar, bed topography (e.g. Frostick and Reid, 1979; Graf, 1983). Wolman and Gerson (1978) note that for small- to medium-sized basins, channel widths are wider than for similar sized perennial basins, whilst once basin size exceeds 50 km^2 a maximum width of between 100 and 200 m is attained. This apparently universal maximum is attributed by Reid and Frostick (1989) to two possible factors. Firstly, where convective rainfall is the main contributor to a flood peak, the limited size of convective rain cells may restrict the total possible precipitation input and hence the discharge of a stream, regardless of its size. Equally, transmission losses to the wetted perimeter may counteract any additional discharge provided by runoff or tributary inputs once the channel width has crossed a critical threshold.

The planar bedform typical of most dryland channels is probably related to the channel width, which allows the spread of flow and hence maintains relatively shallow depths of water within the channel. This shallow flow discourages major bar formation and tends to lead to infilling of temporary scour hollows and planing of bar forms (Reid and Frostick, 1989). Flat beds are common in most relatively straight ephemeral channels, although bar forms do develop at bends in the channel and in braided channels (Rust and Nanson, 1986).

On the basis of studies of ancient and contemporary fluvial deposits, Reid and Frostick (1989) suggest three main attributes of sediments deposited by ephemeral streams. Firstly, deposits generally show horizontal lamination, with alternating layers of coarse and fine deposits. This is probably a result of deposition associated with the plane beds typical of ephemeral channels. Secondly, from studies of desert river deposits preserved in the geological record, sediments are usually finely bedded. Finally, clay layers may occur, either associated with the settling of suspended sediment or as clay-curls which have been entrained and deposited as intraclasts.

(ii) Changes in channel form

Channels may change their planimetric and morphometric properties at a range of timescales in response to individual floods or due to extraneous factors such as human, climatic or tectonic influences. At an almost instantaneous timescale, bedform changes are the most common response to fluctuations in discharge and sediment load, whilst changes to channel pattern coupled with either bed erosion or aggradation may occur in response to longer term variations in environmental conditions. It is these longer term changes that are more appropriate at the timescale of development represented by Kalahari *mekgacha*.

Changes in channel form are regarded by Knighton (1984) as involving two main stages. The first is the effect that environmental change may have on runoff and sediment yield, and the second is the channel response to changes in discharge and sediment load. Table 2.1 (after Knighton, 1984) gives a simplified indication of how climatic change influences runoff and sediment yield, based upon calculations by Schumm (1968). The initial climate is an important factor in determining the nature of any change, particularly with regard to its influence upon vegetation cover and runoff. Table 2.1 suggests, for example, that a shift from a sub-humid to a warmer, drier regime would reduce runoff and cause no change in sediment yield. In the light of the preceding discussion regarding the nature of floods in semi-arid fluvial systems this may be oversimplified owing to the use of mean values for both runoff and sediment yield. A shift to more arid conditions may reduce runoff but still cause channel widening due to higher magnitude flooding.

The possible ways in which a channel may respond to changes in discharge and sediment yield are shown in table 2.2, based upon relationships determined by Schumm (1969) from semi-arid and sub-humid rivers in Australia and the USA. It should be noted that the table only shows possible directions of change and is based upon relationships which are by no means universal (Knighton, 1984). Indeed, from studies of the Salt River in Arizona, Graf (1983) notes very little response of channel parameters such as width to changes in discharge and sediment characteristics. Four combinations of change in discharge and sediment yield are possible, with table 2.2 indicating the predicted response of each element of channel morphology to such change. Relating this to the changes indicated in table 2.1, the onset of an arid climate in a previously temperate area would lead to aggradation due to increased sediment yield and decreased runoff. There would, however, be increased magnitude floods, able to transport coarser sediments and potentially forming broad low sinuosity channels. Clearly these relationships are highly generalised and other factors such as the effect of biomorphic response times following environmental changes upon potential hillslope erosion need to be considered (Knox, 1972). Furthermore, channel aggradation or degradation in response to environmental change is considerably affected by antecedent conditions, with the response at different locations often varying with position within the drainage network (Knox, 1972).

The long-term effect of channel erosion or aggradation is to adjust the channel gradient. As such, a change in gradient may be regarded purely as a response mechanism to variations in discharge and sediment yield. However, if valleys and channels have existed for long time periods, then the comparatively short timescales considered by many fluvial process studies may not be appropriate. Gradient changes may be imposed upon the fluvial system due to alterations in the environmental setting of a catchment, particularly as a result of tectonically-induced changes in base-level.

Table 2.1: Possible effects of climatic change on runoff and sediment yield (after Knighton, 1984). $^{+}$ denotes increase, $^{-}$ a decrease and ^{0} no change.

Original climate	New climate			
	Cooler (T-5°C) Wetter (P+250mm)	Warmer (T+2.5°C) Wetter (P+250mm)	Cooler (T-5°C) Drier (P-125mm)	Warmer (T+2.5°C) Drier (P-125mm)
Temperate: T = 10°C P = 750 mm	R^{+} S^{-}	R^{+} S^{-} or S^{0}	R^{0} S^{0}	R^{-} S^{+}
Sub-humid: T = 12.5°C P = 500 mm	R^{+} S^{-}	R^{+} S^{-}	R^{0} S^{0}	R^{-} S^{0}
Semi-arid: T = 15°C P = 350 mm	R^{+} S^{+}	R^{+} S^{+}	R^{0} S^{0}	R^{-} S^{-}

Where: T; mean annual temperature, P; mean annual precipitation, R; mean annual runoff, S; mean annual sediment yield.

Table 2.2: The effects of changing discharge and sediment load upon channel morphology (information from Knighton, 1984). + indicates increase, - a decrease and ± either increase or decrease.

Change in;		Change in channel morphology element					
Discharge	Sediment load	Mean Width	Mean Depth	Width/ Depth	Meander Wavelength	Channel Sinuosity	Channel Gradient
Increase	Increase	+	±	+	+	–	±
Decrease	Decrease	–	±	–	–	+	±
Increase	Decrease	±	+	±	±	+	–
Decrease	Increase	±	–	±	±	–	+

The long-term effect of a lowering of base-level is channel incision due to local steepening, with an abrupt break of slope migrating headwards (Pickup, 1977). The rate and extent of degradation depends upon the rate and extent of base-level lowering, and generally decreases away from the channel outlet. Degradation may also be temporarily impeded by inputs to the channel caused by increased erosion, and the rate of nickpoint recession may be slowed by the presence of bed armouring or resistant lithologies beneath the channel bed. The effect of a rising base-level is less significant, generally resulting in the development of a "depositional wedge" of sediment as a result of decreased competence (Richards, 1982; Knighton, 1984).

In the context of southern African fluvial studies, Partridge (1969) notes the importance of localised increases in channel gradient due to crustal warping, and also changes due to river capture. Tectonically initiated nickpoint recession is also suggested by Partridge as being the most common cause of terrace abandonment, although clearly, degradation may be the result of a number of different environmental factors operating in combination.

Studies of southern African drainage systems developed upon alluvial plains within shield and platform deserts have identified differences in long-profiles between exoreic and endoreic systems (Stengel, 1964). In extremely arid areas such as the Namib Desert, endoreic basins often exhibit convex (as opposed to concave) long-profiles. This has been attributed to declining discharge downstream (Goudie, 1972*b*), which results in the deposition of heavy sediment loads during sporadic flood events and localised channel steepening. Small convexities may also be superimposed onto a general long-profile convexity, particularly in areas where flow is restricted within narrow sections of valleys (Wilkinson, 1988).

In summary, whether a channel aggrades or degrades is influenced by gross environmental changes as well as upstream and down-stream controls. Aggradation may be caused by increased sediment input in headwater regions, by a rise in the base level, or by environmental changes leading to a basin-wide increase in sediment yield. Conversely, degradation is influenced by entrapment of upstream sediment sources, falling base-level and an overall decrease in sediment yield (Richards, 1982). However, although these relationships assume constant discharge rates and may not be universally applicable to dryland fluvial systems (particularly when placed within a context of climatic change and the complex effects of such change upon vegetation cover, runoff, sediment yield and discharge) they may be broadly appropriate. Perhaps the most important contrast between dryland and temperate fluvial systems is the extent to which dryland systems are event responsive. As such they are highly complex, both in space and time, and very difficult to predict and model. As a result of this complexity, explanations of channel behaviour based upon concepts of equilibrium may be extremely limited (Graf, 1983). As Baker et al. (1983) note, this inherent variability makes the palaeoenvironmental significance of sedimentary deposits within dryland fluvial systems very difficult to assess, particularly in areas such as the Kalahari where long-term hydrological records are lacking.

2.2.2 Fluvial landscapes and palaeodrainages in dryland areas beyond the Kalahari

This chapter has, thus far, concentrated upon concepts, processes and process-form relationships arising from studies of arid and semi-arid fluvial systems. Before considering the role of groundwater processes in valley development, it is appropriate to consider some examples of palaeodrainage networks from other low relative relief dryland areas. As intimated in chapter 1, it is important to view valley development both in terms of changing environments but also within a tectonic and regional setting. The importance of a regional understanding is shown in the following accounts of the palaeodrainage systems of central Australia and the buried valleys of the Eastern Sahara, both of considerable antiquity and both considered to have developed by fluvial activity under former wetter climatic conditions.

(a) The interior drainage of central Australia

Possibly the most important factors determining the present geomorphological features of the Australian interior are the length of time over which the landscape has been exposed to subaerial processes and the low rate of denudation which the continent has experienced since the Mesozoic (Gale, 1992). Whilst still part of the supercontinent of Gondwanaland, much of Australia was affected by the Permo-Carboniferous glaciation, but since that time has not been exposed to significant glacial activity (Gale, 1992). As a result, landscape evolution in areas covered by ice during this period can be considered to date from the termination of glaciation i.e. possibly from the late Permian (Van De Graaff *et al.*, 1977). There is evidence to suggest that the Yilgarn Block of Western Australia had a subdued relief even during the Permian (Mabbutt, 1988), which would result in the relatively slow operation of water driven processes (Gale, 1992). Areas free of ice cover may have evolved over even longer periods, with some landforms suggested to date back to the Cambrian (Stewart *et al.*, 1986). In addition to long exposure to subaerial processes, the continent has generally been tectonically stable since the last major orogenic event which occurred in the Permo-Triassic. Due to the combination of relative crustal stability and long-term exposure, many landscapes have developed over similar timescales to the tectonic evolution of the Australian continent (Gale, 1992).

Within these ancient landscapes are large numbers of complex, extensive palaeodrainage networks, which have evolved over varying timescales (inset to figure 2.1). The oldest networks are developed on the Ashburton Surface in central Northern Territory, where valley terraces are overlain by and merge into (and hence are older than or contemporaneous with) Middle Cambrian marine sediments (Stewart *et al.*, 1986). In southwestern Western Australia some drainage lines follow Permo-Carboniferous valleys and are not covered by later deposits. This suggests that they may have been active since the end of the Permian (Van De Graaff *et al.*, 1977; Gale, 1992). The majority of palaeodrainages, however, appear to date back to the Middle to Late Cretaceous (Van De Graaff *et al.*, 1977).

In Western Australia, chains of salt lakes now occupy the former positions of many valleys, some supplied by ephemeral streams. Valleys are also commonly associated with the presence of alluvium and valley-calcrete which aids identification in areas now devoid of drainage (Van De Graaff *et al.*, 1977;

Mann and Horwitz, 1979; Glassford and Semeniuk, 1991). The distribution of palaeodrainage networks (figure 2.1) has been reconstructed by a variety of methods, including the use of geological maps, vegetation mapping, altimetric and geophysical data and the interpretation of aerial photography (e.g. Bunting *et al.*, 1974; Van De Graaff *et al.*, 1977). Detailed mapping of the Western Australian Canning and Officer Basins using NOAA-AVHRR imagery has identified major palaeodrainages and tributary networks with no surface expression, in areas now covered by the Great Sandy, Great Victoria and Gibson Deserts (Tapley, 1988).

Van De Graaff *et al.* (1977) identify seven major palaeodrainage "provinces" (figure 2.1), including one group of internally draining valleys in the northeastern Gibson Desert and southeastern Great Sandy Desert (Province 2 on figure 2.1). All other valley systems slope fairly continuously towards the coast, and connect with the Indian Ocean (either directly or via active rivers), the Bonaparte Gulf or link with the Finke River and drain into Lake Eyre (Wopfner and Twidale, 1967; Pickup *et al.*, 1988).

Tectonic activity, in the form of gentle crustal warping and tilting, has played a major role in the present form of palaeodrainage networks. The one group of internally directed valleys are suggested to have once drained southwards, but became endoreic as a result of drainage diversion following Cenozoic crustal uplift (Van De Graaff *et al.*, 1977). It is also suggested that palaeo-valleys which are presently directed northwards and westwards towards the southern Great Sandy Desert once drained in a southerly direction but have been disrupted by uplift. This is postulated on the basis that distinct depressions containing valley-calcretes cross continental drainage divides in many places (Van De Graaff *et al.*, 1977) and hence predate the uplift of these divides.

A common feature of many palaeodrainages is some form of structural control of orientation and location. Many valleys in the Great Sandy Desert are aligned parallel to major structural elements, whilst the Throssell and Baker Palaeorivers of the Gibson and Great Victoria Deserts have structurally controlled tributaries (Van De Graaff *et al.*, 1977; Tapley, 1988). There is also evidence that reactivation and movement along major lineaments has occurred during the development of valley networks in the Canning Basin and has disrupted the direction of flow (Tapley, 1988). Structural and crustal movements need to be viewed within the context of the overall evolution of Australia. This is perhaps most clearly illustrated by many of the palaeorivers in southwestern Western Australia, which are either directed northwards and westwards towards the Indian Ocean, or north and east to the present-day Bunda Plateau. Van De Graaff *et al.* (1977) infer from the lack of southward drainage that the valleys were initially instigated as consequent streams developed on a crustal dome prior to the rifting and separation of Australia from Antarctica during the Late Palaeocene to Early Eocene.

Superimposed upon the tectonic changes which have affected palaeodrainage networks are changes in climate. The general implication of palaeodrainage studies is that valleys developed under wetter climates but have subsequently been disrupted under increasingly arid conditions (Van De Graaff *et al.*, 1977).

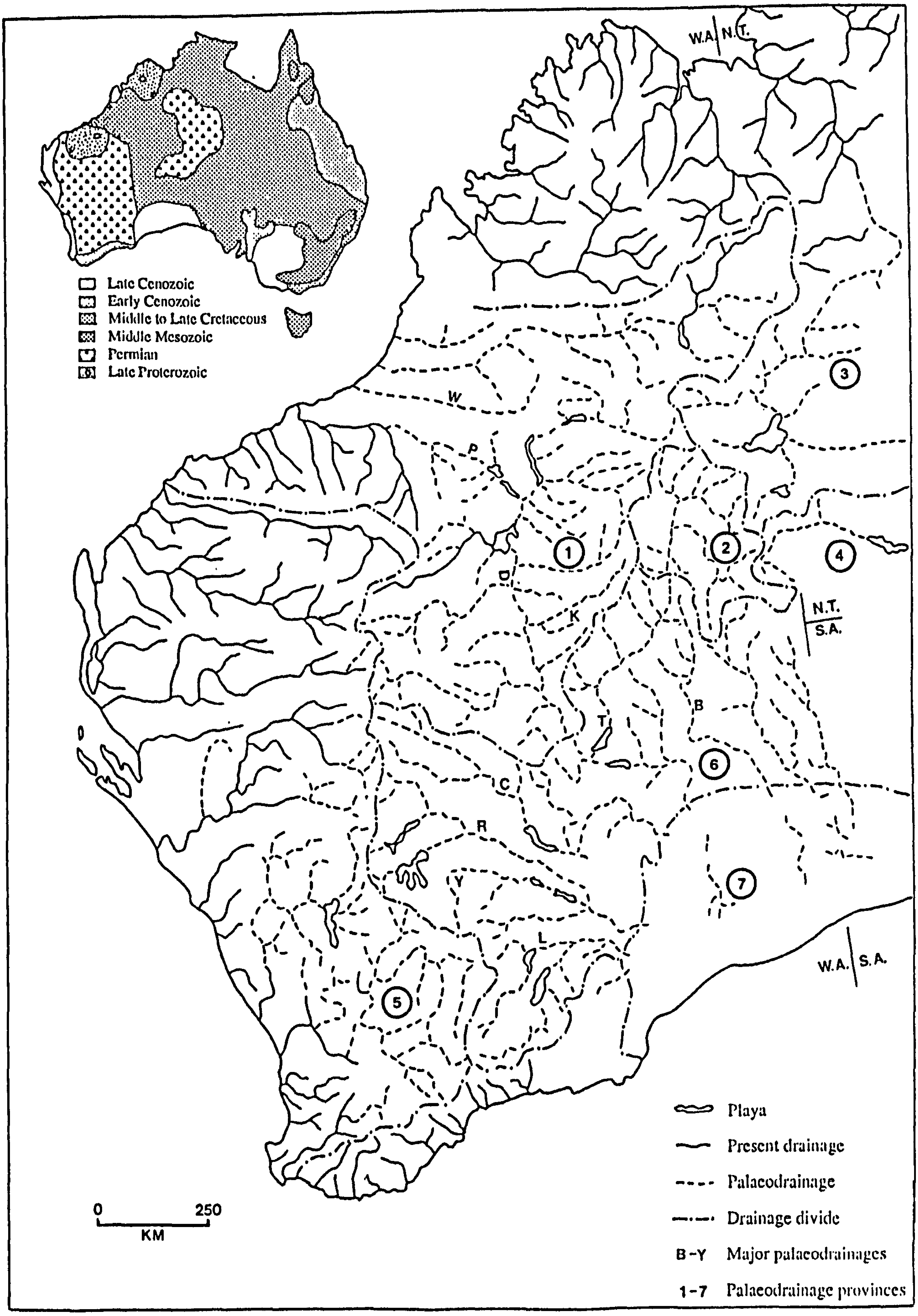


Figure 2.1: The palaeodrainage systems of Western Australia (after Van De Graaff *et al.*, 1977). Inset: The ages of drainage inception in Australia (after Wilford, 1991).

The pattern suggested for much of southern Australia is a gradual shift from subtropical conditions during the Miocene to semi-arid conditions in the Pliocene with aridity in the Pleistocene (Pickup *et al.*, 1988), possibly connected with the separation of Australia and Antarctica. The combined results of tectonic and climatic changes has led to the choking of many drainages by alluvial or aeolian sedimentary infill, with many present-day drainages now marked by playas (Jacobson *et al.*, 1988; Arakel, 1991). Contemporary flow may occur during extreme flood events (Van De Graaff *et al.*, 1977) but otherwise the main importance of palaeodrainages is in the transmission of groundwater (Arakel *et al.*, 1989).

(b) The buried channels of the Eastern Sahara

Equally extensive, although not as ancient, palaeodrainages have also been recognised in northern Africa. In 1981, the first Space Shuttle "Shuttle Imaging Radar" system (SIR-A) revealed a series of extensive buried valleys containing river systems (termed the "Radar Rivers") in the Arbain Desert, part of the Eastern Sahara straddling the Egypt/Sudan border to the west of the Nile Valley (McCauley *et al.*, 1982). These valleys, some of which have floodplains wider than that of the Nile, are presently covered by the aeolian Selima Sandsheet and have little or no surface expression except for the presence of many dry playa lakes along the line of their former courses (McCauley *et al.*, 1982). However, shallow pits dug in the centres of features identified as valleys from SIR-A imagery revealed fluvially deposited alluvium comprising coarse sand to fine pebbles at depths of as little as 1 m below the sand sheet surface (Schaber *et al.*, 1986).

The discovery of these valley systems had a major impact upon the previously held view that fluvial activity had little role in the development of landforms in the Eastern Sahara. Fluvial denudation had previously been suspected, on the basis of the presence of relict drainage divides, but not on the scale identified by SIR-A imagery. Three types of channel can be identified; broad alluvial valleys or basins, braided or anastomosing channels sometimes within wide valleys, and long narrow isolated channels incised into bedrock (McCauley *et al.*, 1986*a*). The broad alluvial valleys are between 10 and 30 km wide and up to hundreds of kilometres long with stubby tributaries. The areas of braided channel range between 0.5 and 2 km wide, with calcified islands of alluvium and fine gravel between the channels. The third type of channel is also identifiable on Landsat imagery, although clearly defined alluvial fans are identifiable from SIR-B data where the bedrock channels once debouched into larger alluvial valleys (McCauley *et al.*, 1986*a*).

The main significance of the radar rivers is in their implication for regional drainage in northeastern Africa. The valleys trend in a generally southwesterly direction, and predate the development of the various predecessors to the Nile Valley (McCauley *et al.*, 1986*a*). Development was associated with the 2000 m uplift of northeastern Africa as a result of crustal doming during the formation of the Red Sea Rift approximately 30 to 40 million years BP (McCauley *et al.*, 1986*b*). Following this uplift, the Red Sea Hills to the east of the present course of the Nile would have formed the highest mountains in northeast and central Africa and the source for a number of major consequent rivers. McCauley *et al.* (1986*a*,*b*) suggest that at its greatest extent, the "Trans-African" drainage system formed by the coalescing of these rivers crossed Africa from the Red Sea Hills via the Bodélé-Chad Basin to the Gulf of Guinea (figure 2.2). Drainage orientation was controlled by the regional slope, with the connection to the Atlantic controlled by low ground in the Benue and Niger troughs or at the site of the Cretaceous "Trans-Sahara seaway" which once linked the Atlantic to the former Tethys Ocean (Kogbe, 1980). Disruption of this once major drainage came as a result of Miocene crustal doming (figure 2.2) and later intracontinental vulcanism, together with headwater capture following the incision of the Nile Valley due to the drying up of the Mediterranean during the late Miocene (McCauley *et al.*, 1986*b*). The isolated segments of the drainage system were subsequently reoccupied by younger regional drainages during the Quaternary (Burke and Wells, 1989).

The concept of the "Trans African" drainage system, has, however, been criticised on a number of counts by Burke and Wells (1989). Firstly, absolute dating of bedrock in the Red Sea Hills and Uweinat indicates that the timing of uplift events does not correspond to the chronology of development proposed by McCauley *et al.* (1986*a*,*b*). Fission-track dates for the Red Sea Hills suggest that uplift may have occurred much later than the proposed 30 to 40 million years BP, and as such these hills could not have formed a source area. Uplift at Uweinat probably occurred during the middle Eocene prior to the development of southwestward drainage, and would have formed an obstacle to the flowpath proposed by McCauley *et al.* (1986*a*,*b*). Secondly, there is evidence from ancient sediments that the course of the ancestral Nile may have been largely integrated by the middle Miocene, which would have cut the course of the "Trans African" drainage system. Thirdly, the timing of most rapid sedimentation in the Niger Delta does not coincide with the maximum extent of the "Trans African" system. Sedimentation proceeded at the comparatively slow rate of 2 km/m.y. when the drainage system would have operated, but reached a maximum rate of 8 km/m.y. during the Miocene when the system is suggested to have been dismembered. As an alternative to the southwestward draining system proposed by McCauley *et al.* (1986*a*,*b*), Burke and Wells (1989) suggest that the "Radar Rivers" once formed tributaries to the ancestral Nile draining into the Sudan basin after early Miocene uplift (inset to figure 2.2). However, neither explanation may necessarily be completely correct. The "Trans African" drainage system and the alternative hypothesis are equally viable interpretations of the available data, although the idea that the Radar Rivers once formed part of the Nile better explains some features.

(c) The significance of palaeodrainage studies

The examples of both the interior drainage of Australia and the Radar Rivers demonstrate the need for a long-term view of landscape evolution, and the importance of placing drainage evolution in its tectonic framework. The tectonic setting of Australia has remained remarkably stable since the Permo-Triassic orogen (Gale, 1992) with the development of continental drainage systems dating back at least as far as the Permian. In contrast, the drainage of northeast Africa has been significantly affected by regional uplift and crustal doming associated with continental rifting during the Tertiary (McCauley *et al.*, 1986*a*). The tectonic development and drainage evolution of central southern Africa will be discussed fully in chapter 3.

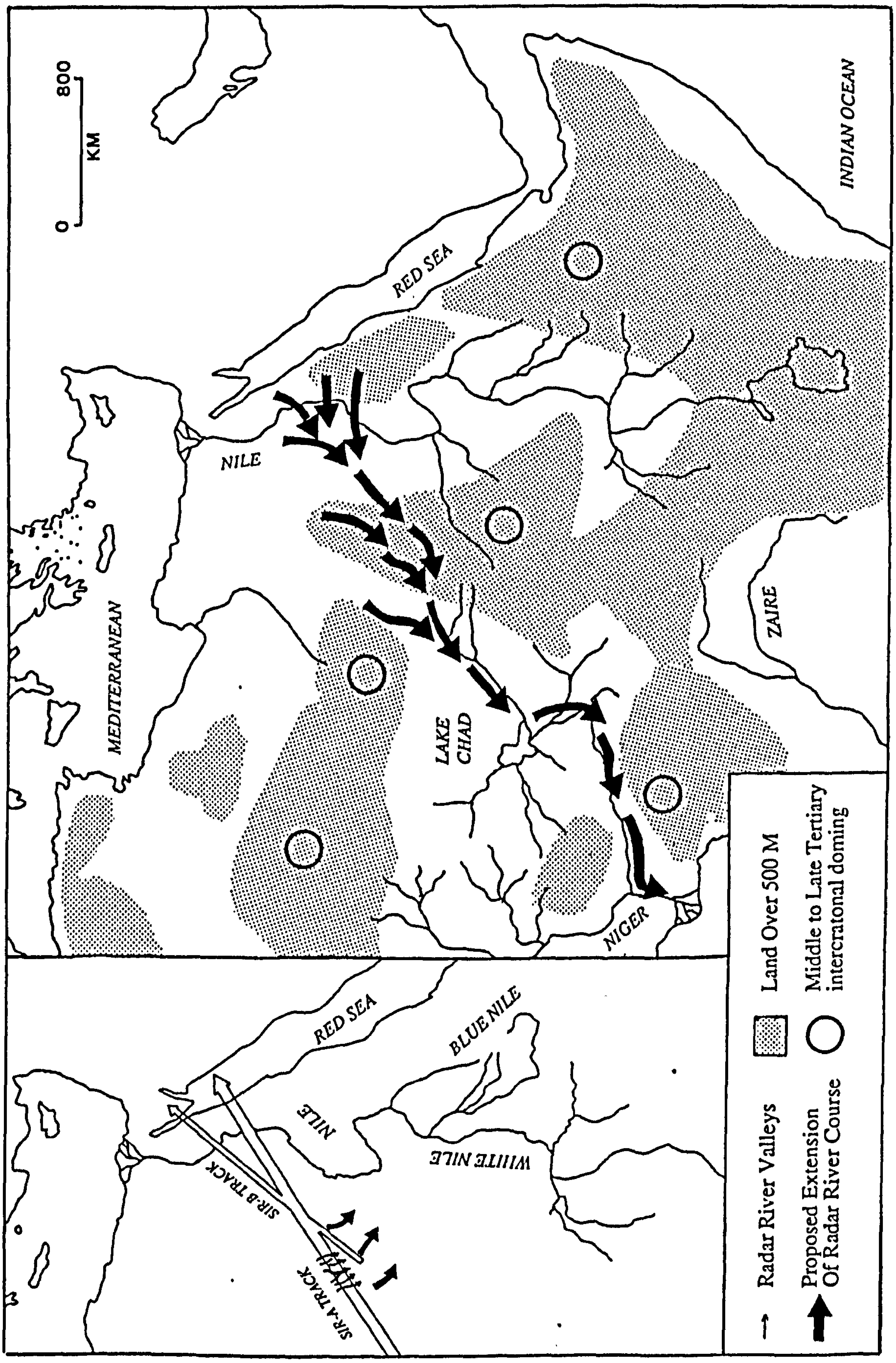


Figure 2.2: The hypothesised course of the "Trans African" drainage system (after McCauley *et al.*, 1986*b*) and the proposed connection of the Eastern Sahara "Radar Rivers" to the Nile (inset; after Burke and Wells, 1989).

2.3 Groundwater processes in geomorphology

The role of groundwater in the geomorphology of humid environments has been much studied, particularly in terms of its role in the process of solution of carbonate rocks and cave development. However, whilst an extensive literature exists on carbonate solution, the general importance of groundwater is little understood, to the extent that many authors regard cave formation as the most important, if not the only, product of groundwater circulation (La Fleur, 1984).

Groundwater is also important in landscapes where carbonate rocks are not present and in less humid climates, although owing to the increased timescale over which many processes operate in these non-carbonate, arid to semi-arid landscapes, the temporal role of groundwater is much less easy to observe. It has been suggested that groundwater may be especially significant in the early development of landforms (Higgins, 1984), although in many environments its influence may be obscured and overwritten by subsequent subaerial processes (La Fleur, 1984). Only where processes have somehow been prevented from operating and evidence is preserved can the role of groundwater in a landscape be appreciated.

The potential role of groundwater as a geomorphic agent in valley development has been recognised since the time of Peel (1941), and yet has been described by Higgins (1984) as one of the least understood factors in landform genesis, particularly in semi-arid environments. Thus the remainder of this chapter is devoted to a consideration of the role of groundwater erosion as a factor in landscape development, in the form of a summary of studies from a number of environments. The scope of this section is necessarily large owing to the fact that the potential geomorphologic importance of groundwater in the Kalahari has only recently been postulated (Shaw and De Vries, 1988). Consequently the landforms of the Kalahari are only briefly considered and most of the discussion centres around other examples. The role of groundwater in valley development from a number of environments, both terrestrial and extra-terrestrial, is considered, with an emphasis on areas with semi-arid climates (although some examples from more humid climatic regimes are included). The role of groundwater in pan or playa development is also briefly discussed.

2.3.1 Groundwater and valley formation

(a) Groundwater processes

The groundwater processes considered important in drainage development are *piping* - erosion by shallow throughflow of soilwater - and *sapping* - erosion by outflow of groundwater at the base of a cliff or river bank (Higgins, 1984). Comparatively little quantitative work has been undertaken on either process, in terms of mechanisms, rates or resulting landform morphology, although piping has been more extensively studied. Piping is, however, of greater importance in channel initiation (Dunne, 1980) whilst sapping is more dominant in valley network development. As such, sapping is given more detailed consideration.

(i) Piping processes

Piping was originally a civil engineering term used to describe the flushing of sediment from the base of dams by seepage, but has also been used by soil scientists to describe the flow of soil water underground where pipes or tunnels are created. In drylands, piping occurs when runoff drains into surface cracks or burrows in friable but relatively coherent material and is able to move laterally due to the presence of a porous layer or lateral opening, or is forced to do so by reaching an impermeable layer, the bottom of a vertical opening or a perched water table (Higgins, 1984). Erosion occurs along the subsurface route of lateral movement to form a pipe or tunnel. Should this tunnel or pipe collapse, the gully may extend upslope and consequently modify soil water flow towards it, thus propagating network growth by a process of positive feedback (Dunne, 1980). In areas where the water table is in close proximity to the surface, soilwater through-flow effectively merges with shallow groundwater flow, so piping becomes sapping and the processes overlap according to the definition by Laity and Malin (1985) below.

(ii) Sapping processes

The first suggestion of the role of sapping in valley development was from observations made on the Gilf Kebir plateau of Libya by Peel (1941). The study of sapping processes became particularly popular, however, in the mid-1970's, when Mariner 9 images of Mars revealed features of apparent "fluvial" origin. The term "sapping" can be used to describe any undermining of a cliff or bank by a stream, waves, biological activity or other mechanical weathering processes, but in this discussion the use will be restricted to activity influenced by groundwater. Sapping is the main process associated with the return of subsurface flow from either shallow or deep sources (Dunne, 1980; Gomez and Mullen, 1992), and is defined by Laity and Malin (1985, p.203) as;

> "the process leading to the undermining and collapse of valley head and side walls by weakening or removal of basal support as a result of enhanced weathering and erosion by concentrated fluid flow at a site of seepage".

Conceptual models of drainage network development by sapping processes are described by a number of authors, including De Vries (1974; 1976) and Baker (1990). Actual sapping processes are, however, very difficult to observe, primarily due to the problems of accessibility at headwalls of active gullies and streams. As a result, field studies of sapping processes are limited and most quantitative assessments of the role of sapping in valley development have come from experimental work (e.g. Kochel and Piper, 1986; Howard and McLane, 1988; Gomez and Mullen, 1992). Whilst useful, differences in experimental technique, particularly in the variety of initial conditions used in stream table experiments, make it possible for only general observations to be made from such experiments.

The process of sapping involves at least some inter-granular flow (Higgins, 1984) and weathering proceeds by slow grain release at the point of groundwater emergence due to chemical weathering, hydrological action, and mechanical weathering processes such as freeze-thaw and salt weathering. Experimental studies show that the main method of drainage network development by sapping processes is headward erosion, which proceeds rapidly during the early stages of formation (Kochel *et al.*, 1985; Gomez and Mullen, 1992). As development proceeds complex feedback mechanisms occur (Baker, 1990), with changes in valley form influencing sapping processes, and *vice versa*. Tributary growth occurs as a result of permeability variations (Howard *et al.*, 1988) and disturbances in subsurface flow (Dunne, 1980). Joints and structures have been suggested to have considerable importance in trunk and tributary valley development, particularly in terms of orientation of stream sections (e.g. Laity *et al.*, 1980; Bannister, 1980; Pieri *et al.*, 1980; Laity *in* Baker, 1990). In cohesionless sediment, seepage forces at the site of emergence of sub-surface flow are most important controls on headward erosion (Howard and McLane, 1988). However, in cohesive bedrock, mechanical and chemical weathering (Laity *et al.*, 1980) are likely to be the dominant displacive processes. Higgins (1984) separates erosion at springs ("spring sapping") and laterally uniform discharges of groundwater ("seepage erosion") with these two processes forming the ends of a continuum. Spring sapping is the most important for drainage network development.

From studies in Colorado, Howard *et al.* (1988) suggest five factors necessary for sapping processes to operate;

i) A permeable aquifer of a transmissive rock type,

ii) A rechargeable groundwater system, ideally of large areal extent,

iii) A free face at which water can emerge, which may have developed by a variety of processes,

iv) Some form of structural or lithological inhomogeneity to increase local hydrologic conductivity,

v) A means of transporting material from the free face.

Additional factors affecting sapping processes occur at a variety of scales (Baker, 1990). At the largest scale, sapping is influenced by climate, water table levels, material type and structure, with a massive, resistant lithology overlying a less competent one providing the ideal conditions for scarp retreat. At a meso-scale, the microclimate of a seepage area may affect weathering processes, whilst at a micro-scale, material structure and cohesivity is significant.

The notion that sapping processes operate in valley development does presuppose some form of depression or free face at which groundwater can initially emerge, a fact which the literature generally tends to ignore, although it may be that piping and deep-weathering processes (section 2.3.2, below) play a role in this. Stream table experiments by Gomez and Mullen satisfy this initial requirement by grading sediments to form a converging channel prior to sapping network generation "to encourage the development of a trunk stream" (Gomez and Mullen, 1992 p.467). However, two-dimensional experimental work by Howard and McLane (1988) using cohesionless sediments generated headward erosion by sapping processes without the need for an initial channel. Another factor stressed by many studies (e.g. Howard *et al.*, 1988) is the need for a lithological or structural inhomogeneity in order for sapping processes to proceed. This may not always be the case, as experimental studies by Howard and McLane (1988) and Gomez and Mullen (1992) generated headward erosion by sapping processes which was controlled primarily by the position of the water table.

(b) Drainage and valley networks produced by groundwater sapping

Valley systems and drainage networks considered as being formed primarily by the process of sapping have been described by a number of authors from highly variable environments and at a range of scales, including ephemeral microdrainage nets on coastal beaches (e.g. Higgins, 1984), submarine canyons (e.g. Robb *et al.*, 1982; Robb, 1990) and the massive chasms of equatorial Mars (e.g. Baker, 1980). The form and development of some of these networks will now be considered.

(i) Beach micro-drainage networks

The formation of small dendritic drainage networks commonly seen on most sandy beaches represents the clearest example of drainage developed entirely by groundwater outflow in unconsolidated sediment, being analogous to sapping in larger scale systems (Higgins, 1974, 1982, 1984). During backwash, water flowing downslope progressively thins and breaks into a diamond pattern of flow, forming rhomboidal surface patterns separated by micro-channels. During the falling tide, the beach water table is always above tide level and intersects the beach face near the top of the swash zone, below which the sand is saturated. Water seeps back to the surface once the backwash has drained away, emerging in the micro-channels between the rhomboidal patterns.

It is beyond this stage of development that Higgins (1982, 1984) considers beach microchannels as most analogous to larger systems. Rills develop and continue to develop where groundwater is constricted by an impermeable substrate or where the water table is held at the same level for an extended period. Groundwater discharge and sapping develops steep gully heads in the rills at the upper limit of the saturated zone, with mass-wasting at the gully heads. The direction of headward growth is controlled by the direction of groundwater flow, and parallels the orientations of the sides of the original rhomboidal pattern, forming an angulate network. Should the water table fall there may be development of secondary gullies within these networks.

The micro-valley systems have steep, blunt valley heads and walls, joining a flat valley floor at an abrupt angle. The drainage pattern is characteristically angular with few tributaries, and upslope of the networks there is little or no evidence of surface runoff. Despite the generally small areas covered by these networks, Higgins (1984) notes that their size covers around 1.5 orders of magnitude and thus suggests that they may be partially independent of scale and a useful analogy of larger drainage systems.

(ii) Martian channels

As has been mentioned above, the return of remotely-sensed images of the equatorial regions of Mars by Mariner 9 stimulated considerable interest in the channels depicted (e.g. Carr, 1980; Baker, 1982, 1983, 1985; Stiller, 1983). Many of these features were attributed to sapping (Sharp, 1973; Higgins, 1982, 1983; Belderson, 1983; Kochel *et al.*, 1985; Sharp and Malin, 1985) by analogy with terrestrial valleys, despite the circular argument involved as a result of being unable to observe the formative processes. The Martian atmosphere contains little water, and water cannot exist on the surface for very long, so the origin of these "fluvial" features is highly speculative. It has been suggested that ground ice, permafrost or liquid trapped underground may have been released and formed these valleys, but until more evidence can be provided any suggestions are purely conjectural (Higgins 1982, 1984).

Mariner 9 showed huge "outflow" channels, heading at areas of chaotic terrain, analogous to the Badlands of Washington state, U.S.A. (Sharp, 1973; Baker, 1982; Higgins, 1984), indicative of a massive catastrophic outflow of a fluid. Whatever this fluid, it did not remain on the surface for very long, as there remains no sign of lacustrine deposits.

Other valley forms are described by Baker (1980, 1982, 1985), Pieri (1980), Sharp & Malin (1985), Kochel and Piper (1986) and Howard *et al.* (1988), amongst others. These "dendritic" channels (Carr, 1980) are best developed on the south side of Ius Chasma, part of the Valles Marineris equatorial rift system (Higgins, 1984). The largest tributary canyon to Ius Chasma is 135 km long and 10 km across, and is greater than 2 km deep at its mouth, with all tributaries ending abruptly in the Sinai Planum plateau, with no visible catchment. The tributaries show abrupt headscarps and an angulate pattern, with a definite asymmetry since valleys are best developed to the south of Ius Chasma. It has been suggested that this is due to the northern regional dip in the area, so that any "groundwater" would flow towards the valley and produce sapping landforms preferentially on the south side (Higgins, 1984). Pieri (1980) describes cuspate terminations at the heads of small tributaries feeding into Nirgal Vallis and suggests that this is a product of sapping. The tributaries show marked parallelism and a lack of competition with low junction angles (<25°). Cross-sectional areas of tributaries are similar to that of the main valley trunk despite being of lower magnitude.

Attempts to calculate quantitative planimetric morphometric parameters from Viking images of Mars have been made by Pieri (1980), but it is difficult to calculate traditional measures of drainage basin morphometry from Martian imagery as no drainage divides are evident (Howard *et al.*, 1988). Kochel and Piper (1986) have analysed the morphology and morphometry of valleys on Mars, and demonstrate the similarities between these sapping valleys and networks developed experimentally. Howard *et al.* (1988) suggest that Martian valleys were most probably produced by sapping because of the presumed greater ease of formation by sapping as compared to runoff. Sapping does, however, need large volumes of water in order to produce a well developed network and is most efficient in unconsolidated sediments. Experiments by Howard and McLane (1988) show that a minimum of 10 times more water than sediment are necessary, with a ratio of 100 to 1000 times being more likely (Howard *et al.*, 1988).

(iii) Terrestrial sapping networks

Drainage networks attributed to development by groundwater sapping have been described from various parts of the globe (Laity *et al.*, 1980), including England (Sparks and Lewis, 1957-58; Small, 1964), Australia (Jennings, 1979; Baker, 1980; Young, 1986), New Zealand (Schumm & Philips, 1986), Hawaii (Kochel and Piper, 1986), Libya (Peel, 1941), Egypt (Maxwell, 1979; El Baz *et al.*, 1980), the Colorado

Plateau of the U.S.A. (e.g. Howard *et al.*, 1988; Laity *et al.*, 1980; Pieri *et al.*, 1980) and SE Botswana (Shaw & De Vries, 1988). The valleys of Libya, Colorado, Hawaii and Botswana will now be considered.

Libya

Peel's description of wadis from the Gilf Kebir plateau in Libya was one of the earliest studies to consider the role of sapping in an arid environment (Peel, 1941). He describes wadis showing "many departures from the normal valley form", with straight sections connected by abrupt angles, and tributaries entering at various angles. Valley width varies between wide reaches and narrow gorge-like stretches, with evidence of lateral undercutting of cliffs and basal retreat (Peel, 1941). The valleys have a stepped longitudinal profile, terminating at an abrupt headward cliff, with no evidence of surface flow over the surrounding plateau or over the lip of the headward cliff.

Peel (1941) discounted wind erosion as a significant formative process and suggested that the wadis show the appearance of being "cut out from below" rather than "let down from above", attributing this to groundwater sapping. Peel inferred that sapping proceeded due to rainfall infiltrating the sand surface and re-emerging as springs at the cliff base. Over wide areas of the Gilf Kebir, silicified sandstone outcrops at the surface and would provide a protective capping, thus retaining the steepness of the valley walls.

Hawaii

Drainage networks on the Hawaiian islands with high level aquifers due to control of water movement by lava flows have been used to provide a terrestrial analogue to Martian valleys. Networks have been described by Hinds (1925), Baker (1980) and Kochel and Baker (*in* Baker, 1990) from many islands, with, for example, long parallel valleys exhibiting v-shaped cross sections and abrupt amphitheatre-like heads on Mauna Kea. Similar valleys are described on the NE slopes of Kohala and Haleakala volcanoes on Maui, in areas of very humid climate.

Valleys considered to be influenced by groundwater sapping have smooth floors, steep walls with evidence of mass wasting, often have hanging valleys, and amphitheatre headwalls. A small basin area to catchment ratio is common, with a low drainage density and a paucity of tributaries downstream. Tributaries often show asymmetry due to regional structure and join at irregular angles, with valleys commonly maintaining their width or even widening upstream (Baker, 1980; Kochel and Piper, 1986; Howard *et al.*, 1988; Kochel and Baker [*in* Baker, 1990]). Principal components analysis clearly separated Hawaiian valleys formed by runoff and by sapping (Kochel and Piper, 1986).

Colorado

Numerous studies have described the valley networks of the Glen Canyon region and other parts of the Colorado Plateau, which have many features characteristic of development by sapping. Howard *et al.* (1988) report elongate basins and networks within the Navajo Sandstone with a simple network structure showing little or no branching. Most networks are of Strahler order 1 or 2 with none greater than order 3, with valleys exhibiting relatively straight longitudinal profiles, hanging valleys and a comparatively small catchment area compared to the size of the valleys. Alcoves and tafoni occur along the valley sides and valleys often end at amphitheatre heads. Groundwater currently discharges near the base of the cliff walls, and is usually confined to a zone 2m thick, although at valley heads the zone may be 20-25m (Howard *et al.*, 1988).

Strong structural control is also evident from the Colorado valleys, with networks often highly asymmetric and showing a high consistency of tributary junction angles. There is also evidence of "pervasive parallelism" of tributaries over wide geographical areas (Howard *et al.*, 1988). Theatre-headed valleys appear to grow in an updip direction, on surfaces with an overall regional dip of 1° to 4° (Howard *et al.*, 1988). Pieri *et al.* (1980) note a marked trellis pattern, best developed where the main valley elongation is perpendicular to major NE-SW structural trends. Laity *et al.* (1980) and Laity (*In* Baker, 1990) consider joints as important lines of weakness or paths for pipes of sub-surface flow, with persistence of valley forms due to rock strength. The asymmetry of drainage networks may be due to groundwater capture on the updip side of a slope, with groundwater flow influenced by changes in sedimentological facies (Howard *et al.*, 1988).

Groundwater emergence and sapping in the Navajo Sandstone is partially controlled by diagenetic factors (Laity, 1983). Groundwater emerges in a region of reduced permeability at the base of the formation. This permeability change is due to clay-minerals, quartz overgrowths and extensive carbonate deposition which block pores; by a process of positive feedback, sapping in turn causes diagenetic changes to occur. Calcite precipitation appears to be the principal agent for loosening sand particles, resulting in the growth of tunnels to accommodate the volume of converging groundwater outflow at the valley head.

As a result of studies in Colorado, Howard *et al.* (1988) have theoretically investigated the effect of climatic change upon sapping processes. They conclude that increased rainfall leading to increased groundwater outflow may not necessarily lead to increased sapping. Whilst there would be more moisture available for processes such as freeze-thaw, if runoff occurred or groundwater output were too high, it would preclude the accumulation of minerals and salts which act as important mechanical weathering agents.

Southeast Botswana

As noted in Chapter 1 of this thesis, Shaw and De Vries (1988) suggest that the dry valleys in the region of Letlhakeng in southeast Botswana were formed by groundwater erosion processes, citing the valley form, occurrence of relict spring lines and suites of duricrusts as evidence for this. The valleys show a number of features which are suggested to support a sapping origin; a rectilinear pattern, low tributary bifurcation ratios and small catchment areas, as well as a lack of surface drainage evidence.

Conditions suitable for groundwater discharge are suggested to have been provided by a decreased hydraulic gradient and an increase in clay content leading to a reduction in permeability. Shaw and De Vries (1988) suggest that to form valleys of the dimensions seen in parts of Botswana (up to 50 m deep and 2 km across), with abrupt valley heads and gorge-like stretches, a long term fall in the water table would be necessary, possibly provided by regional uplift.

(iv) Morphometric features of sapping networks

From the above discussion it is suggested that if all or most of the following morphometric features are present in a valley network, it is liable to have developed from groundwater sapping;

a) Abrupt initiation of channels, with amphitheatre headwalls, and little or no sign of surface runoff into valley head-regions,

b) Alcove development, springs and possible tufa development (Marker, 1988) in the headward region,

c) Steep valley walls with an abrupt angle to a fairly flat valley floor, with evidence of mass-wasting,

d) A small basin area to canyon area ratio,

e) A low drainage density,

f) A long main valley with a constant valley width, or evidence of widening in a headward direction,

g) Short first order tributaries with a paucity of tributaries downstream, and possible hanging valleys,

h) Asymmetry of tributaries controlled by regional structure, with possible parallelism of tributaries and constant junction angles,

i) A longitudinal profile which is flat and may show step-like features.

2.3.2 *In situ* deep-weathering and valley development: the case of African dambos

In addition to erosion by groundwater sapping and fluvial processes, the role of *in situ* deep-weathering processes have also been implicated as a factor in dryland valley development. This suggestion has been made within the context of the formation and development of southern African dambos (also termed *fadamas, vleis, bas-fonds* and *bolis* in other parts of the continent; Boast, 1990). Dambos are defined by Mäckel (1974) and Acres *et al.* (1985) as broad, shallow, seasonally waterlogged, grassed depressions without a marked stream channel, occupying valley floors, which commonly occur at the headwaters of stream networks. Boast (1990) discusses the variable geomorphological and geographical meanings of the word "dambo"; within this review, "dambo" will be used to describe features conforming to the above general definition. There are two schools of thought concerning dambo formation, which are broadly analogous to the present study. One sees fluvial erosion and slope transportational processes dominating development (Mäckel, 1974), whilst the other implies that dambos have developed by pseudo-karstic *in situ* deep-weathering independent of the fluvial network (McFarlane, 1989, 1990). As such, it is instructive to briefly consider the morphology and arguments concerning the formation of dambos; full literature reviews for dambos are provided in Thomas and Goudie (1985), Boast (1990) and Bullock (1992).

(a) Dambo location, morphology and hydrology

Features to which the term "dambo" can be applied are found over much of Africa, particularly central southern regions, and may also occur in India and South America (see Acres *et al.*, 1985 p.69). They are usually shallow with a concave profile, slope angles of less than 6° and an axis gradient of less than 1% (Mäckel, 1974). The planimetric form of dambos often shows a "club-head" and is commonly influenced by geological structures and changes in lithology (Boast, 1990), and by differences between the angle of slope of the regional water table and the dambo longitudinal gradient (Acres *et al.*, 1985). Dambo soils are usually gleys, with textural properties dependent upon the underlying geology. Within the valley occupied by the dambo, soil characteristics change from the hill crest to the dambo floor, with an increasing clay content down-profile (Mäckel, 1974). Dambo vegetation cover is determined by soil and drainage characteristics, with the seasonally waterlogged dambo floor commonly supporting grass and sedge communities and the interfluves colonised by woodland species (Whitlow, 1985).

Dambo hydrology is characterised by seasonal waterlogging due to a shallow, fluctuating watertable, with the supply of water arising from a combination of direct precipitation and throughflow from interfluves (Whitlow, 1985; Boast, 1990). Bullock (1992) notes the possible influence of dambo properties on the movement of water, and the effect this has upon drainage lines down-valley of dambos. Dambos may influence the response of drainage systems to rainfall by acting as "sponges", but there is no concensus as to whether a high density of dambos in the headwater areas of a network significantly affects base-flow or flood-peak magnitudes (Bullock, 1992). Boast (1990) concludes that if dambos are net receivers of throughflow but do not provide baseflow outputs, then there must be substantial water loss from the dambo surface by evapotranspiration during the dry season.

(b) Dambo formation

Whilst dambos exhibit a variety of forms, their gross characteristics are controlled by three main factors (Boast, 1990); climate, geology and relief. Acres *et al.* (1985) note that dambos are found in areas with strongly seasonal rainfall regimes, with total annual precipitation in the range 600 to 1500 mm, with Bond (1967) suggesting an optimum rainfall of 875 mm p.a. It would, however, appear that climate is not the overriding control of contemporary dambos, since the majority of Zimbabwean dambos would otherwise be relict features (Whitlow, 1985). Geology and the influence of lithology upon soil characteristics are also important in determining drainage, whilst a flat, gentle relief appears to be a prerequisite for formation (Mäckel, 1974), except in the development of "perched dambos" (Acres *et al.*, 1985). A low relief allows the infiltration of surface water, thus reducing erosion, and results in large areas of water table outcrop (Boast, 1990).

As noted above, two contrasting origins have been put forward to explain the formation of dambos; development by fluvial and colluviation processes, and by *in situ* deep-weathering. Each of these will now be discussed.

(i) Fluvial activity and slope transportational processes

The model of dambo formation by fluvial activity considers development to be dominated by erosion at the headwaters of drainage systems to which dambos are invariably linked. As rivers cut back, the resulting valley may be infilled by channel and translocated slope deposits. Mäckel (1974) regards sheet-wash as the main slope transportational process, with the steepest parts of the dambo slope gradually retreating as a result of the downslope transportation of material. This produces a gradual lowering of the dambo margin which causes an extension of the saturated groundwater outcrop area and reduces the area of potential woodland occupation. In this way, grasslands extend and the dambo margin migrates. The collapse of underground pipes is also suggested as a method of headwater extension and for the removal of material from the dambo, thus promoting surface lowering (Mäckel, 1974).

Mäckel (1974) identifies two types of dambo, based upon the extent of sedimentation in the dambo floor. The first, termed a "degradational dambo", has surfaces which are continually lowered but the resultant material is lost from the system. "Aggradational dambos" have an increased influence of fluvial activity, with cut and fill sequences reflecting the alternating dominance of river incision and subsequent deposition.

(ii) *In situ* deep-weathering processes

In contrast to development by fluvial activity, McFarlane (1989) suggests that dambos developed without the action of rivers, with chemical and biochemical corrosion as the main weathering mechanisms. This hypothesis is based on the identification of a number of dambos displaying features untypical of fluvial valley, most notably where dambos cross drainage divides. However, as noted during the discussion of Australian palaeodrainages (above), this feature may be due to localised crustal uplift after valley initiation. The presence of favourable lithologies, concentrations of fractures, joints and faults within bedrock promotes deeper or more extensive weathering, with surface lowering proceeding by solute leaching (McFarlane, 1990). Sub-surface flows beneath the valley axis remove solutes, and either move them towards streams or into groundwater storage. McFarlane (1989) suggests that geologically-controlled advanced weathering in time produces a topography of highs and lows, with dambos aligned with the trends of major geological lineaments. The coincidence of surface lowering associated with dissolution of bedrock along lines of fractures has also been noted in eastern Botswana by Gieske and Selaolo (1988), where fractured aquifers are characterised by shallow surface depressions.

However, *in situ* weathering and fluvial erosion should not be regarded as mutually exclusive processes. McFarlane (1989) recognises the significance of potential rejuvenation of the fluvial network and the influence this would have on dambo development. Fluvial activity would cause incision and possibly the removal of sediment from the dambo floor by flushing, with a concomitant lowering of regional water tables and increase in slope gradients. McFarlane proposes two separate systems of water flow within dambos. Shallow subsurface flows occur above and within sediments deposited above the clay soil commonly found in the dambo floor, whilst deeper flows are supplied from the interfluves. It is these lower flows that remove solutes from beneath the dambo. This deeper water may also emerge during periods when the water table is high at discrete seepage sites and springs around the perimeter of the dambo-floor clay zone.

The model for dambo development thus proposed by McFarlane (1989) envisages formation by deep-weathering, primarily during periods of little or no fluvial activity. The dambos produced by this long-term process may be modified by fluvial activity, but *in situ* weathering is the primary evolutionary process. The presence of fluvial sediments in the floors of many dambos indicates that they are not inactive, but it is proposed that deep-weathering is the most important agent for surface lowering.

2.3.3 Groundwater and pan development

Deep-weathering by groundwater has also been proposed as a mechanism in the formation of pans (also known as playas, sebkha, kavir and bays). Pans are depressions with no outlets and rarely inlets, containing water for at least part of the year. Various processes have been implicated in their formation, including tectonism, volcanic activity, meteorite impact, biogenic activity and many geomorphological processes including deflation and solution (Stow, 1873; Alison, 1899; Weir, 1969; De Brulyn, 1974; Lancaster, 1978*a*; Grobler *et al.*, 1987; Shaw, 1988*a*; Shaw and Thomas, 1989), the most popular being deflation. Pans and *mekgacha* development are closely-genetically related, with pans often developing within dry channel alignments. Pans also form the "headwater" areas of a number of Kalahari *mekgacha*, such as tributaries to the Okwa Valley system.

(a) Pan development: deep-weathering and hydrologic control of deflation

The role of groundwater in pan formation has been most closely considered by workers in Australia (e.g. Bowler, 1986) and the United States (e.g. Osterkamp and Wood, 1987), with other basins and sabkhas fed by groundwater seepage described in Tunisia by Coque (1962), in Algeria by Boulaine (1954) and in many other locations in North Africa by Glennie (1970).

Pans were traditionally considered as being "closed-basins", a view which Torgersen *et al.* (1986) consider to be incorrect and misleading. A salt-lake is;

> A "dynamic system with fluxes of water and salts along fluid dynamically and kinetically determined pathways which include reversible and irreversible chemical reactions and transients in the water and salt balances" (Torgersen *et al.*, 1986, p.9).

Within the system of water fluxes experienced in pan environments, the role of groundwater is the least understood. However, two major roles for groundwater can be identified, these being as a control and factor in pan/playa development and in the deposition of pan sediments (Bowler, 1986; Torgersen *et al.*, 1986).

The role of groundwater in pan development is twofold; percolating groundwater can lead to pan-floor subsidence by direct dissolution processes (Osterkamp and Wood, 1987), and the groundwater table can act as a base level for wind deflation (Bowler, 1986). Osterkamp and Wood (1987) and Wood and

Osterkamp (1987) propose a lithologically specific groundwater model for playa development based upon observations in the Southern High Plains of Texas and New Mexico, substantiated by mass-balance calculations. Whilst playas are likely to be initiated by a number of processes, these authors suggest that the expansion of the pan floor area occurs essentially by dissolution and removal of material beneath the playa surface. The infiltration, weathering and downward transport of solutes by percolating groundwater, along with the removal to the subsurface of clastic material along solutional pipes, is envisaged as leading to the gradual subsidence of the playa surface.

The other possible geomorphic function of groundwater in pans or sabkhas is via seasonal fluctuations in the water table which cause wetting and drying of basin clays (e.g. Bowler, 1986; Thomas *et al.*, 1993). Such clays accumulate due to the tendency for fines to migrate to topographic lows (Lancaster, 1978; Thomas, 1984; Shaw, 1988*a*; Shaw and Thomas, 1989). Given sufficiently strong winds, deflation of the basin down to the groundwater table can occur, often leading to the formation of clay-dunes (Bowler, 1973, 1986; Thomas *et al.*, 1993). This groundwater controlled deflation leads to concentrations of salts (as efflorescences or as saline clay layers) in the surface layers which can further weather clays.

The alkaline environment formed by such basins has a high potential for chemical weathering and includes conditions ranging from calcic soils to widespread salt deposits (Shaw, 1988*a*). A complex balance occurs between groundwater inflows and outflows and mineral precipitation and removal (cf. Torgersen *et al.* (1986). Salts commonly found in pan environments include sodium and calcium sulphates, sodium chloride and sodium carbonate, of which the former is particularly effective as a weathering agent (Goudie, 1986*b*).

Studies in the Kalahari (e.g. Lancaster 1977, 1978*a,b*, 1979*b*; Goudie & Thomas, 1985, 1986) have generally regarded the role of groundwater in pan development as minimal. Indeed, Lancaster (1978*a*) suggests that groundwater is not a major factor in the formation of Kalahari pans, since the majority of Botswana pans are located on Kalahari Sand; if groundwater were important in pan development then most pans would be located in areas of pre-Kalahari bedrock which provide better aquifers. Thus Lancaster considers that seepage from groundwater is only of local significance, although in later papers (Lancaster 1986*b*, 1988) he recinds this view.

Significant evidence of the role of groundwater in pan development, does, however exist (Shaw, 1988*a*). Detailed geophysical and geochemical investigations of Mogatse Pan and Lokwane Pan near Kukong (Butterworth, 1982; Farr *et al.*, 1982) reveal alteration and calcrete formation in sediments underlying pans to a depth exceeding 30m (the limit of drilling). Shaw suggests that this shows the great antiquity of pans, as multiple layers of calcrete are reported beneath pans by Butterworth (1982).

The studies by Osterkamp and Wood (1987) and Wood and Osterkamp (1987) in the southern United States and the work on the distribution of pans in the Kalahari appear to have a number of common themes. In both the U.S.A. and Botswana, landforms associated with pans are usually those which impede local drainage or lead to preferential movement of groundwater, such as sand-dunes and fossil valleys (Shaw, 1988*a*). Structural control of pan location is also implicated in both regions by virtue of the role of deep-weathering in pan development, which Shaw (1988*a*) suggests is controlled by the presence of sub-surface lineaments.

(b) Pans and palaeoenvironmental reconstruction

The use of sediments and former lake levels associated with pans for purposes of palaeoenvironmental reconstruction is well established in the geomorphological literature (e.g. Street and Grove, 1976; Street, 1980, 1981; Street-Perrott and Roberts, 1983). However, caution is needed in the interpretation of sedimentological data, since the deposition and preservation of sediments is strongly affected by groundwater-sediment interactions. Recent studies in Australia (Bowler, 1986; Torgersen *et al.*, 1986) and the Kalahari (Thomas *et al.*, 1993) reveal a complex set of processes which determine the erosion and deposition of pan sediments and the development of associated lunette dunes. The interaction of groundwater, runoff, surface-water, porewater and sediments, aeolian transport, and chemical and biological reactions affect the geomorphic features of lakes (Torgersen *et al.*, 1986). These processes control the presence of pan sediments by affecting sediment deposition, deflation and re-deposition, composition and diagenesis, and the lacustrine biota. Thus the balance of these processes controls the evolution of the lake, the means of recording that evolution and any processes that modify or destroy palaeoenvironmental records. These processes should be fully understood before any palaeoclimatic and palaeoenvironmental information can be usefully drawn from pan sediments (Thomas *et al.*, 1993).

2.4 Chapter summary

This chapter has reviewed the role of fluvial and groundwater processes in the development of valleys from a variety of studies undertaken in many different environments. Perhaps the main distinction which can be drawn between the two sets of processes is the extent to which they require major climatic changes in order to operate. Fluvial erosion may have taken place due to the presence of either perennial or ephemeral flow. In order for perennial rivers to exist in an environment similar to that of the Kalahari, a major change in regional climate, most probably combined with drainage diversions, and the clearance of aeolian sediments from valley floors would be required. The operation of erosion by ephemeral streams would similarly require the removal of sediment, but not the same extensive regional climatic and drainage changes although clearly some increase in rainfall would be necessary.

In contrast, given suitable geological conditions, groundwater sapping processes would only require a higher water table, which could have several causes, not just increased local precipitation. Deep-weathering processes may actually be occurring at present in areas where water collects on a seasonal basis. These generalised scenarios, however, do not take into account the regional and tectonic context necessary for the understanding of long-term evolution of drainage systems, a factor which has been stressed throughout much of this chapter. The following chapter provides that context for the Kalahari and, in particular, introduces the main geomorphological lines of evidence for Quaternary environmental change in the region.

Chapter 3
The place of *mekgacha* within the Kalahari environment

3.1 Introduction and general information

The purpose of this chapter is to provide a setting for *mekgacha* within the context of other aspects of the Kalahari environment, in particular the geological context and relationship to other water-formed features. Following an introduction in which the Kalahari physiographic area, regional climate, soils and vegetation are briefly discussed (section 3.1), the geology (3.2), groundwater and hydrogeology (3.3) and landforms (3.4) of the Kalahari Desert are considered. This last section concentrates particularly upon landform interpretation and includes a chronology of Kalahari landform development.

(a) The definition of the Kalahari

The area of the Kalahari Desert is usually taken as being synonymous with the Republic of Botswana in central southern Africa, as it occupies 80% of that country (Jones, 1982). However, defining the location, limits and extent of the Kalahari is not as simple as this, and a number of possible schemes exist. Confusion arises since the name Kalahari has been used to define a physiographic area (e.g. Wellington, 1955), an ecozone (Werger, 1978) and has also been given to a group of sediments (Passarge, 1904). Furthermore, whilst the area is traditionally regarded as a "desert" (despite being dominantly semi-arid) the name "thirstland" has also been used to describe its environment (Schwarz, 1920; Debenham, 1952; Grove, 1969).

The earliest attempt at delimiting the Kalahari was by Passarge (1904) who split the region into the Northern (northwest of the Okavango Delta), Middle (the Okavango-Makgadikgadi area) and Southern Kalahari (south of the Makgadikgadi Depression and including the Bakalahari Schwelle drainage divide). Since then various geological, biogeographical and geomorphological clarifications of this definition have been attempted, including those of Du Toit (1927) and Grove (1969).

The generally accepted definition of the Kalahari today is shown in figure 3.1, and consists of the geologically and structurally defined "Kalahari physiographic region" (or Mega Kalahari; Thomas, 1984*b*), within which is located the "Kalahari Desert". The Kalahari Desert is bounded in the north by the Zambezi-Okavango-Etosha swamp zone and in the south by the Orange River (Grove, 1969). The western extent is marked by higher country rising towards the Great Escarpment and by an erosion scarp in central eastern and southeastern Namibia (Du Toit, 1927). The eastern limit is less clearly defined, with the exception of eastern Botswana where there is a definite break between hardveld and sandveld areas in the vicinity of the Limpopo watershed (Thomas and Shaw, 1991*a*). Perhaps the most difficult Kalahari boundary to delimit is that in the western Zimbabwe/northeastern Botswana area, which is marked in places by an abrupt scarp but is so well vegetated that the designation of this area as a desert is difficult to support (Thomas and Shaw, 1991*a*).

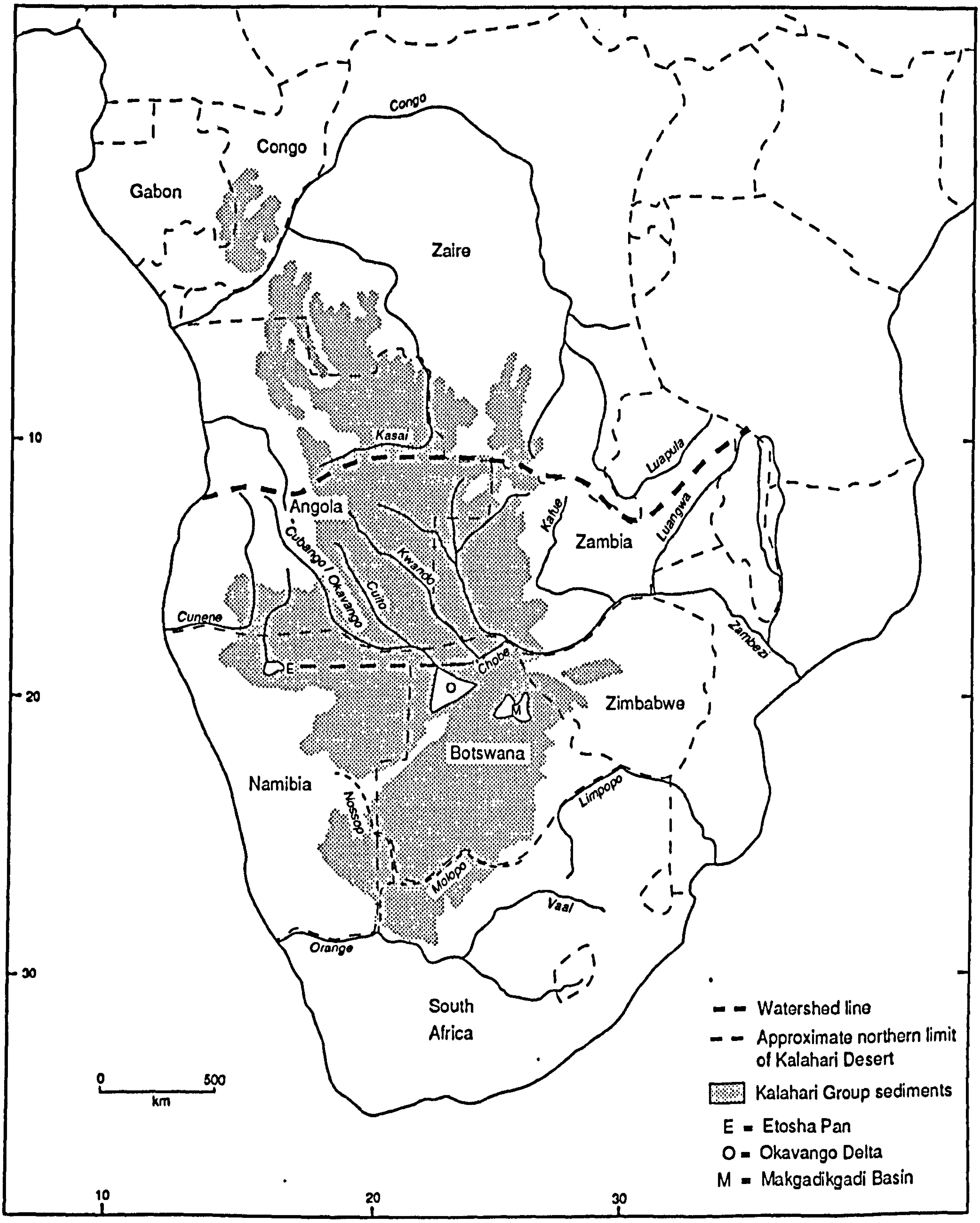


Figure 3.1: The Kalahari physiographic region. The Kalahari Desert occupies the area from the Orange River in the south to a line between the Zambezi River, Okavango Delta and Etosha Pan in the north (from Thomas and Shaw, 1991*a*).

The Mega Kalahari has developed due to sedimentary infill of a number of depositional continental basins (Thomas, 1988*a*) since the Jurassic. The major surface unit of the Mega Kalahari is the Kalahari Sand of the Kalahari Group sediments (SACS, 1980), which cover an area of some 2.5 million km^2 (Baillieul, 1975), including parts of South Africa, Namibia, Botswana, Zimbabwe, Angola, Zambia, Zaire, Congo and Gabon. The mean altitude of the Mega Kalahari is approximately 1,000 metres asl (Jones, 1982).

(b) Climate, vegetation and soils

The climate of the Kalahari Desert (as opposed to the Mega Kalahari) is best characterised by the contrasting annual wet (summer) and dry (winter) season, and the marked year-to-year variations in moisture availability. Kalahari synoptic climatology has been described more fully elsewhere (e.g. Pike, 1971*a*; Schulze, 1972; Andersson, 1976; Westgate 1976; Tyson, 1979, 1986). In general, the climate is controlled by the disrupting effect of the southern African landmass upon the development of the southern hemisphere high pressure belt. The climate is mostly affected by movements of this high pressure belt, especially during the summer, and has increased continentality due to the altitude of the region.

Precipitation in the Kalahari Desert increases to the north and east, with the southwestern Kalahari receiving on average less than 200 mm per annum. Rainfall figures for northern Botswana are around 600 mm per annum. The Kalahari Desert is essentially a summer rainfall zone with precipitation dominated by high intensity rainfall events, usually associated with convective thunderstorms. There is high seasonality, with over 80% of annual rainfall occurring between October and April (Thomas and Shaw, 1991*a*); the summer rains in Botswana commonly begin in November and last until April (Cooke, 1979*c*). There is, however, an interannual variability in precipitation, which exceeds 45% in the southwestern Kalahari (Tyson, 1986).

Temperatures in the Kalahari are generally higher and show less diurnal variability during the wet season. This is primarily due to the coincidence of the wet season with the southern hemisphere summer (Thomas and Shaw, 1991*a*). Relative humidity decreases from east to west and north to south, and is lowest in the dry season. This generally low relative humidity combined with warm to hot temperatures leads to high rates of evapotranspiration, thus contributing to desert conditions. Typical mean daily temperatures and annual potential evapotranspiration rates for selected locations within and near the Kalahari Desert are shown in Table 3.1.

Kalahari soils are generally poorly developed, due to the prevailing climatic and sedimentological setting, with profile development seldom present. Soils developed on the Kalahari Sand are classified as arenosols, although finer textured soils are noted within depressions such as *mekgacha* where higher nutrient levels are attained (Leistner, 1967). Soil characteristics vary markedly with underlying geology, especially where sand cover is thin. An example of this variation is the development of calcisols on the calcium carbonate-rich rocks of the Ghanzi Ridge.

Table 3.1: Mean daily temperature, temperature extremes and annual potential evapotranspiration rates for selected locations within and near the Kalahari (after Thomas and Shaw, 1991*a*).

		Daily mean temperature (°C)						
	Overall Mean	Wet season Max.	Wet season Min.	Dry season Max.	Dry season Min.	Temp. extremes Max.	Temp. extremes Min.	Ann. Pot. Evap. (mm)
Upington	21.9	32.0	16.7	22.8	5.8	43.0	-8.0	3805
Tsabong	19.5	32.1	15.2	24.4	3.6	42.0	-11.0	n.a.
Keetmanshoop	21.0	32.3	16.0	23.8	8.0	42.0	-4.0	3903
Ghanzi	20.5	31.7	14.5	26.2	6.4	42.0	-7.0	3305
Maun	22.0	35.5	18.6	28.0	8.5	43.0	-6.0	3058

n.a. indicates data not available.

Much of the vegetation in the Kalahari can be broadly described as grass-, shrub- or tree-dominated savanna, although there are considerable spatial variations in species diversity and community composition. Species diversity and biomass generally decreases to the southwest, in line with falling mean annual precipitation levels, although this general pattern is locally disrupted by variations in soil type, the effects of bush-fires and various geomorphological factors (Leistner, 1967; Flint and Bond, 1968; Weare and Yalala, 1971; Thomas, 1984*a*). In savanna areas, grass species are dominated by *Aristida, Eragrostis* and *Stiapgrostis* spp., with the most common trees and shrubs being *Acacia*, *Colophospermum*, *Commiphora* and *Terminalia* spp. (Leistner, 1967; Weare and Yalala, 1971).

The Okavango Delta has by far the most diverse range and number of plant species and soil types. Soils types include histosols and arenosols in perennial channels, calcic arenosols on islands, and luvisols, fluvisols and arenosols in seasonal channels and floodplains (Thomas and Shaw, 1991*a*). Vegetation communities range between savanna woodland, riverine forest, aquatic grassland and perennial swamp (McCarthy *et al.*, 1986, 1988*b*).

3.2 The geology and structure of the Kalahari

As noted in the preceding chapter, any consideration of long term drainage development needs to be placed within a geological and tectonic context. In the case of Kalahari *mekgacha* development, an understanding of the deposition of the Jurassic to Recent Kalahari Group sediments and the nature of underlying pre-Kalahari lithologies is required. In addition, the structural development of the region needs to be considered particularly in terms of the potential role of groundwater processes in *mekgacha* development.

3.2.1 Pre-Kalahari lithologies

The Kalahari Sand represents the surface expression of a sequence of sediments in places exceeding 300 m in thickness deposited within the Kalahari Basin (Thomas, 1988*b*). With such vast thicknesses of sediments, exposures of pre-Kalahari lithologies are comparatively rare except in peripheral settings or as inliers. In many areas the only available geological information has been provided by aeromagnetic, geophysical and gravity surveys (Greenwood and Carruthers, 1973; Reeves and Hutchins, 1976; Reeves, 1978*a*; Terra Surveys Ltd., 1978; McEwen, 1979; Lüdtke, 1986) in conjunction with lithological information from boreholes.

Detailed accounts of pre-Kalahari geology are given by a number of authors both in general terms and for specific localities (e.g. Rogers, 1907, 1934, 1936; MacGregor, 1916; Boocock and Van Straten, 1961, 1962; Jennings and Crockett, 1961; Kingston *et al.*, 1961; Boocock, 1962, 1966; Crockett and Jennings, 1962, 1964, 1965; Green, 1966; Smit, 1972; De Villiers and Simpson, 1974; Baldock *et al.*, 1976; Albat, 1978; Wright, 1978; Coates *et al.*, 1979; Mallick *et al.*, 1981; Jones, 1982; Litherland, 1982; Dingle *et al.*, 1983; Visser, 1983; Smith, 1984; Aldiss, 1984, 1987*b*, 1988; Aldiss *et al.*, 1989; Thomas and Shaw, 1991*a*; Aldiss and Carney, 1992). As such, the following descriptions of pre-Kalahari lithologies are necessarily brief. All terms follow the recommended nomenclature of the South African Commission for Stratigraphy (1980).

(a) Precambrian rocks

Surface exposures of Precambrian rocks are limited within the Kalahari, especially in central areas, due to the thickness of Karoo and Kalahari cover. However, studies have revealed two areas of basement rocks; a southeastern area of Archaean cratonic rocks and a northwestern zone of Proterozoic rocks, separated by a southwest-northeast Medial Rift (Reeves, 1977; Jones, 1982).

The southeastern Archaean basement comprises rocks of the ancient Kaapvaal and Zimbabwean cratons, representing many periods of volcanism, sedimentation and tectonic activity prior to 2600 Ma. The craton includes greenstone belts and gneisses, overlain in parts by Proterozoic Ventersdorp Supergroup lavas, Transvaal Sequence sandstones and dolomites (Griquatown Group, Dolomite Group and Blackreef Formation) and Waterberg Group sandstones. In places, ultrabasic lavas of the Bushveld Igneous Complex are intruded into these Groups. (Boocock and Van Straten, 1962; Mallick *et al.*, 1981; Jones, 1982).

To the northwest of the Medial Rift, Precambrian rocks are mostly of a Proterozoic age (Mallick *et al.*, 1981). They are covered by Karoo rocks between the Medial Rift and the Ghanzi Ridge, but northwest of the Ghanzi Ridge they lie unconformably beneath the Kalahari Group. Three groups are represented; the

Kgwebe Formation volcanics, Ghanzi Group meta-sediments and the Damara Sequence sediments (Walker, 1973*a*; Wright, 1978; Mallick *et al.*, 1981; Hegenberger, 1982; Litherland, 1982).

(b) The Upper Palaeozoic to Mesozoic Karoo Sequence

Rocks of the Upper Palaeozoic and Mesozoic Karoo Sequence rest unconformably upon Precambrian lithologies. Despite having a total thickness in excess of 1000 m, they still have poor surface exposure within the Kalahari. Karoo rocks are described in detail by, amongst others, Boocock and Van Straten (1962), Green (1966), Jones (1982) and Smith (1984). The geology of the Karoo Sequence can be simply divided into lower sediments and upper volcanics, with five series of rocks being recognised. These are the Dwyka Formation, and the Ecca, Beaufort, Stormberg (or Lebung in Botswana) and Drakensberg (or Stormberg Lava) Groups (SACS, 1980; Smith, 1984).

Karoo sedimentary deposition occurred in a number of basins as Gondwanaland drifted equatorward (Smith, 1984; Weijermars, 1989), and thus begins with Permo-Carboniferous glacial tillites (Visser, 1983) and ends with desert "red bed" formations. Triassic sedimentation in the Kalahari ended with the extrusion of the series of lavas of the Drakensberg or Stormberg Lava Group, with basalt thicknesses of up to 375 m recorded from boreholes by Smith (1984). The final episode of late- or post-Karoo volcanic activity coincided with the breakup of Gondwanaland, and comprises a swarm of dolerite dykes up to 100 km wide, trending northwest-southeast across the Kalahari (Green, 1966; Reeves, 1978*a*; Mallick *et al.*, 1981). The events associated with this breakup are discussed in the following section.

3.2.2 The development of the Kalahari Basin

(a) The break-up of Gondwanaland

The development of the Kalahari Basin, and thus the accumulation of sediments within it, was closely linked to the separation of Africa from the super-continent of Gondwanaland during the Mesozoic. The breakup of Gondwanaland occurred between 200 and 160 million years ago (Tankard *et al.*, 1982) with rifting and horizontal separation along the line of the Mozambique Ridge (Wellington, 1955). A further period of rifting due to the opening of the South Atlantic occurred around 135-130 million years ago (Tankard *et al.*, 1982). As a result of rifting, thermal expansion of the crust adjacent to the rift-zone caused uplift, generating a hingeline or flexural bulge (De Swardt and Bennet, 1974; Summerfield, 1985*a*). This ultimately generated the Great Escarpment which now fringes southern Africa, with subcontinental interior basins developing from gentle downwarping or rifting within Precambrian cratons (Thomas and Shaw, 1991*a*; figure 3.2); younger extra-cratonic marginal basins were also formed (Summerfield, 1985*a*). These basins became areas of sedimentation during the Cretaceous, with further impetus for sedimentation provided by isostatic readjustments of the crust postdating thermal uplift due to rifting (Summerfield, 1985*a*).

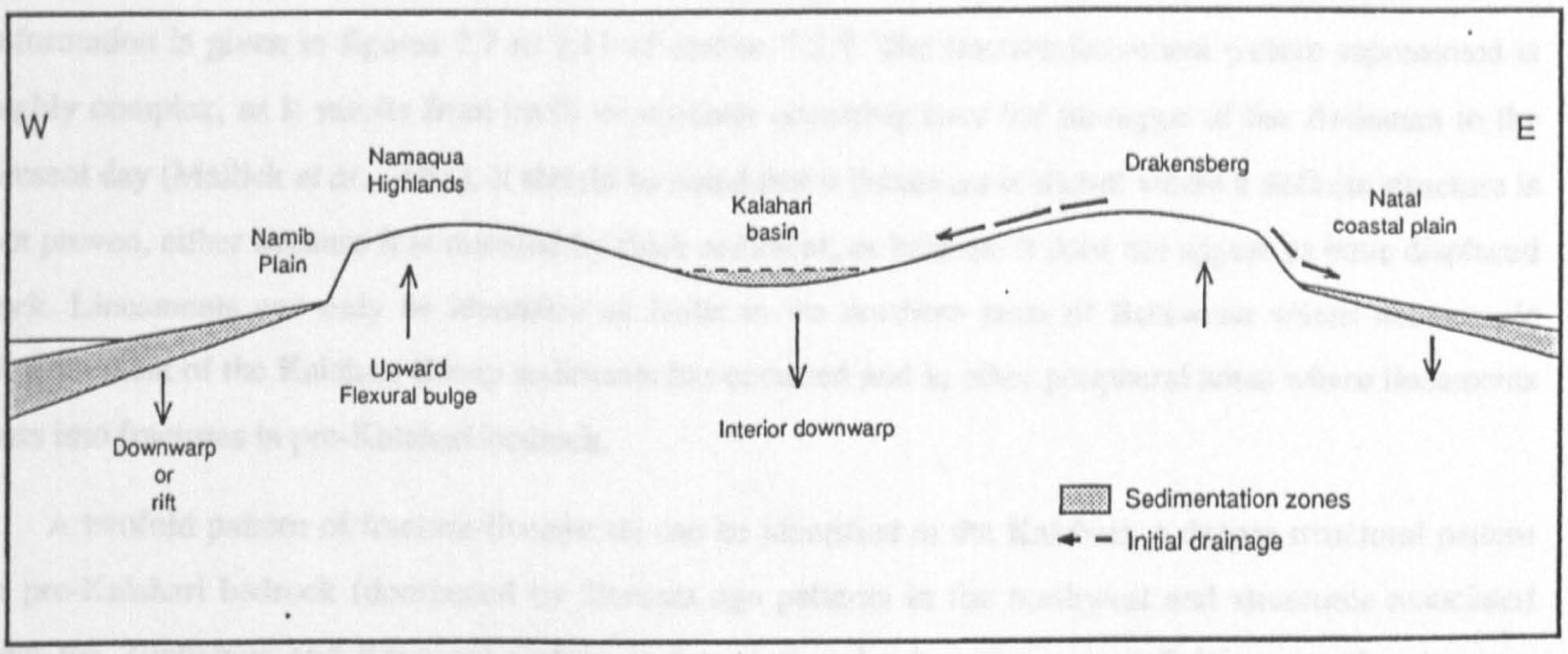


Figure 3.2: Cross-section of southern Africa, indicating areas of uplift and downwarping associated with the breakup of Gondwanaland (from Thomas and Shaw, 1991*a*).

(b) Neotectonics in the Kalahari

As well as the large scale effects due to rifting, neotectonic movements (broadly defined as tectonic movements which have taken place after the establishment of a major plate configuration; Summerfield, 1987) have also significantly influenced sedimentation in the Kalahari Basin, particularly in the Okavango Delta and Makgadikgadi Depression. Du Toit (1926*a*) was the first to propose a connection between the Okavango Delta and the East African Rift System, this being given credence by the location of a seismic axis beneath the Delta (Reeves, 1972, 1978*b*; Scholz, 1975; Scholz *et al.*, 1976; UNDP/FAO, 1977). The Okavango Delta is terminated by the NE-SW trending Thamalakane and Kunyere faults (Hutchins *et al.*, 1976*a,b*; Mallick *et al.*, 1981) with further faults perpendicular to these controlling the "panhandle" of the Delta. The result of gradual rifting beneath the delta was the accumulation of up to 100 m of alluvial deltaic sediments (Hutchins *et al.*, 1976*b*), with other areas of alluvial deposition due to downwarping associated with the Groot Laagte and the Linyanti-Chobe swamps.

The Makgadikgadi Depression has also been influenced by tectonics (Baillieul, 1979), although this is discussed in more detail in section 3.4.1. The effects of tectonic activity on the landforms of Northern Botswana are discussed by Cooke (1975, 1977, 1979*a*, 1980), Grey and Cooke (1977), Helgren and Brooks (1983), Cooke and Verstappen (1984), Shaw (1988*a,b*) and Thomas and Shaw (1988) amongst others. The timing of neotectonic activity is uncertain, but can be relatively dated by the position of faults which cut linear sand dunes in NW Botswana (Wright, 1978; Mallick *et al.*, 1981), and faulting in the Gcwihabedum (Cooke, 1975, 1980). It is probable that seismic activity has been continuous since the breakup of Gondwanaland, if contemporary results are assumed to be representative of past crustal movements.

(c) Fracture-lineament patterns in the Kalahari

The general pattern of geological structures in the Kalahari is given in figure 3.3, although more detailed information is given in figures 7.7 to 7.11 of section 7.2.2. The fracture-lineament pattern represented is highly complex, as it results from earth movements occurring over the timespan of the Archaean to the present day (Mallick *et al.*, 1981). It should be noted that a lineament is shown where a definite structure is not proven, either because it is mantled by thick sediment, or because it does not appear to have displaced rock. Lineaments can only be identified as faults in the northern parts of Botswana where neotectonic displacement of the Kalahari Group sediments has occurred and in other peripheral areas where lineaments pass into fractures in pre-Kalahari bedrock.

A twofold pattern of fracture-lineaments can be identified in the Kalahari; a deeper structural pattern in pre-Kalahari bedrock (dominated by Damara age patterns in the northwest and structures associated with the Zimbabwe and Kaapvaal cratons in the east) and a less dense superficial expression in areas covered by Kalahari Sand. It is the latter group of superficial structures upon which this section will concentrate, with the deeper structures defined by Reeves and Hutchins (1975), Reeves (1977) and Terra Surveys (1978). Many of the superficial lineaments represent faulting in the younger Kalahari Group sediments, but may also reflect deeper fractures. Lineaments are often only identifiable due to vegetational differences above buried fractures as a result of the effect of these deep faults upon groundwater circulation (Mallick *et al.*, 1981).

The superficial fracture-lineament pattern can be divided into two areas which are separated by the Makgadikgadi Line (Reeves, 1977), a zone of fractures extending from the southwest corner of the Makgadikgadi Depression to Union's End on the Nossop Valley. To the north of this narrow zone, the pattern is dominated by Post-Karoo rifting, with NE-SW fractures partly controlled by older structural directions being most prevalent, whilst the south and southeast Kalahari has a pattern dominated by WNW and WSW lineaments (Mallick *et al.*, 1981). As structures mostly within the Kalahari of Botswana are considered in this study, the subdivision into Northern, Western and Southeastern areas of Botswana by Mallick *et al.* (1981) will be used in the following brief descriptions.

(i) Northern Botswana

The pattern of faulting in northern Botswana is dominated by the major NE-SW and NW-SE trending faults at the distal end of the Okavango Delta and controlling the "Panhandle" of the Okavango River. The NE-SW faulting is also apparent in the vicinity of the Makgadikgadi Depression. Mallick *et al.* (1981) identify NW-SE fractures as gravity anomalies beneath the alluvium of the Okavango Rift, which suggests they predate the rift zone which existed prior to the division of Gondwanaland (Rust, 1975). The most important faultline in relation to Kalahari *mekgacha* is the Gomare Fault, which has disrupted the drainage of northwest Botswana (Wright, 1978). In the extreme northwest, a number of fracture lineaments have a N330-340° trend, especially in the vicinity of the Tsodilo Hills (Hegenberger, 1982).

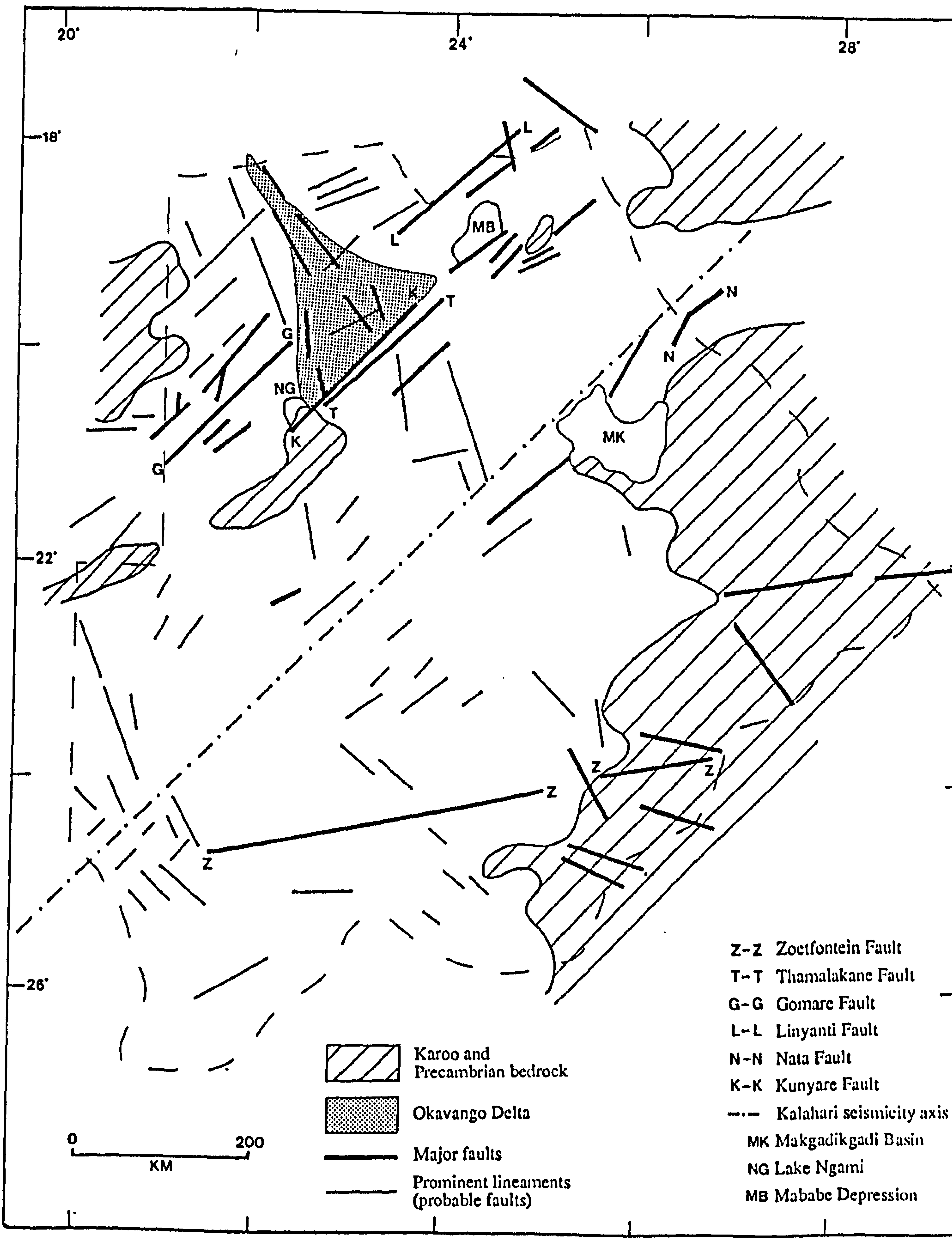


Figure 3.3: Major structural trends in Botswana (after Thomas and Shaw, 1991*a*). Data from Mallick *et al.* (1981) with additional miscellaneous sources as listed in section 7.1.5*d*.

(ii) Western Botswana (north of the Makgadikgadi Line)

Four patterns of fractures can be identified in this area; fractures following the Makgadikgadi Line; faults following the strike of the Ghanzi Group (N045° to N070°); fractures trending N330° to N350° to the west of the Makgadikgadi Depression; and other lineaments/faults in the Ngami Basin area. NE-SW trending faults are prevalent in the area of the Okwa Valley, with many tributaries of the Okwa apparently following fault lines (Jennings and Crockett, 1961; Crockett and Jennings, 1962; Aldiss, 1984; 1987*b*; Aldiss and Carney, 1992).

(iii) Southern Botswana (south of the Makgadikgadi Line)

This area is dominated by a number of structures, including intrusions and dykes associated with the late Karoo dyke swarm and the major Zoetfontein Fault trending at N250° (Mallick *et al.*, 1981). The Zoetfontein Fault can be traced for 500 km across southern Botswana (Terra Surveys, 1978), although only 80 km of this is at the surface. Oblique to the fault are a number of major post-Karoo faults trending N290° to N300°. The surface termination of the Zoetfontein Fault in Botswana includes faulting at N315°, with displacement of Kalahari Group silcrete and calcrete by these faults apparent in the vicinity of Lephephe. In the extreme southeastern Kalahari, fractures trend northwest, east-northeast and northeast, with northwesterly fractures apparent, for example, beneath the Naledi *mokgacha*. The southwest Kalahari is relatively fault-free (Thomas *et al.*, 1988) with only small normal and reverse faults present.

(d) The development of today's landscape

In addition to influencing patterns of sedimentation, the tectonic evolution of southern Africa ultimately shaped landscape development (Partridge and Maud, 1988). This occurred particularly through the development of the Great Escarpment and the effects this had upon drainage (section 3.4.1*a*). The resulting macro-scale landscape consists of a gentle coastal plain backed by the steep slopes of the Great Escarpment, the Highveld inland from this, with the flat interior Kalahari Basin in the centre of the subcontinent. The evolution of this landscape has generated considerable debate, particularly in terms of the establishment of a denudation chronology for development centred around the idea of periodic uplift and erosion cycles (e.g. Dixey, 1958*a*,*b*; King, 1978; Mabbutt, 1955, 1957; Partridge and Maud, 1987, 1988). The general sequence of landscape evolution is reviewed by Partridge and Maud (1988).

3.2.3 Kalahari Group sediments and duricrusts

The tectonic framework in southern Africa following the breakup of Gondwanaland provided the setting for the deposition of terrestrial sediments over the Mega Kalahari, an area of 2.5 million km^2 (Thomas, 1984*b*). These sediments, given the name Kalahari Group (SACS, 1980), exhibit considerable facies variations, resulting in a complex stratigraphy. The understanding of this stratigraphy and attempts at regional intercorrelation have been hampered by the general lack of exposures in the Kalahari, although increasing information attained from exploratory boreholes has improved the situation in recent years. This section briefly discusses the deposition, stratigraphy and lithologies of the Kalahari Group, although the significance and development of Kalahari duricrusts are dealt with more thoroughly in Chapter 6.

(a) History of research

The sediments of the Kalahari first received attention through the work of Passarge (1899, 1904) who proposed the first stratigraphic sequence for the area based upon studies in Ngamiland in Botswana. The evolving understanding of the Kalahari Group has been adequately discussed elsewhere (e.g. SACS, 1980; Jones, 1982; Thomas and Shaw, 1990, 1991*a*) based upon exposures both within and at the periphery of the Kalahari in a variety of countries. These include the studies in Botswana by Passarge (1904), Rogers (1934, 1936), Boocock and Van Straten (1962), Walker (1973*c*), Baillieul (1975), Wright (1978), Coates *et al.* (1979) and Mallick *et al.* (1981), in SWA/Namibia by Range (1912), Mabbutt (1955, 1957), Albat (1978), Hegenberger (1982, 1985/86 and 1986/87) and McDaid (1985), in South Africa by Smit (1977) and Thomas *et al.* (1988), in Zimbabwe by Lamplugh (1902, 1907), MacGregor (1916), Maufe (1920, 1939), Jones (1944), Bond (1946, 1948) and Dixey (1950), and in Zaire and Angola by Veatch (1935) and Cahen and Lepersonne (1952) and in Zambia by Dixey (1950) and Money (1972).

(b) Depositional history and setting

The development of the Kalahari-Cubango-Congo basins following rifting initiated major sedimentation in the subcontinental interior. The inter-cratonic Kalahari basin has clearly existed since the late Palaeozoic, as evidenced by the extensive Dwyka tillite deposits within the basin from the Permo-Carboniferous glaciation of Gondwanaland (Visser, 1983). Initial sedimentation was provided by an extensive endoreic drainage system centred on the basin (De Swardt and Bennet, 1974) with deposition occurring in two separate basins, divided by the tectonically stable Precambrian Ghanzi Ridge (Wright, 1978). Deposition continued throughout the Tertiary, with the thickness of deposits depending upon the pre-Kalahari relief and local neotectonic downwarping; deposits thin towards the periphery of the Kalahari basin.

The sequence of sedimentation in the Kalahari Basin appears to represent a progressive increase in the aridity of the region. Wellington (1955) suggests that this may be attributed to the increased rain shadow effect provided by the Miocene uplifts in the eastern Kalahari margin. Sedimentation began during the Cretaceous (Mabbutt, 1955; Thomas and Shaw, 1991*a*) with the deposition of gravels, conglomerates and marls, followed by sands (which have been subjected to post-depositional modifications and reworking). The region was ultimately blanketed by the Kalahari Sand, which was deposited during the late Tertiary (Maufe, 1935), Pliocene (Mabbutt, 1957), Plio-Pleistocene (King, 1962) or Pleistocene (Dixey, 1958*b*; Bond, 1963). Deposits are generally thickest to the north of the Ghanzi Ridge, with over 300m of sediment in northern Namibia and in the Okavango graben. The thickness of the Kalahari Group sediments constructed from borehole records is shown in figure 3.4 (after Thomas, 1988*b*).

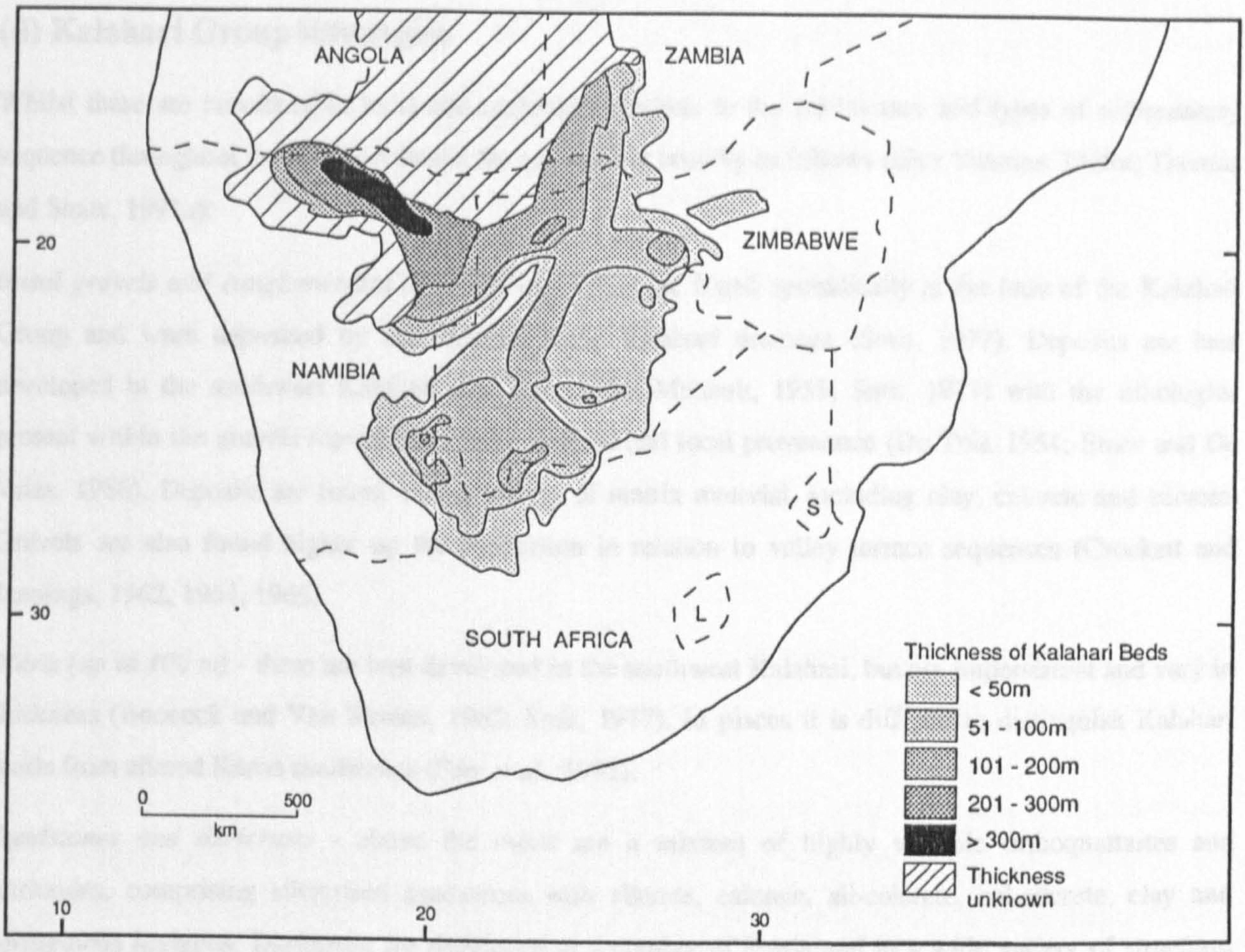


Figure 3.4: The thickness of the Kalahari Group sediments (from Thomas and Shaw, 1991*a*; after Thomas, 1988*b*).

(c) Stratigraphy and classification of the Kalahari Group

Many investigators have regarded the Kalahari Group sediments as a straightforward stratigraphic succession, and have tried to correlate sedimentary sequences between regions. This approach dates back to the seminal work of Passarge (1904, chapter 34) who classified the "Kalahari Formation" into five groups, implying chronostratigraphic relationships between similar beds in various parts of the basin. This was recognised by Rogers (1936) as incorrect because of the possibility of diachronous deposition of similar lithologies over such a large area. Passarge's "Kalahari Formation" thus became the Kalahari Beds (Boocock and Van Straten, 1962), and is now named the Kalahari Group in order to remove any chronostratigraphic implications (SACS, 1980).

It is foreseeable that correlation of the Kalahari Group sediments may occur in the future with the identification of unconformities within the sediments. Sedimentation was probably not continuous, as evidenced by sedimentary sequences of comparable age from other parts of southern Africa (particularly coastal sequences). These indicate hiatuses during the Oligocene-Miocene and the Late Pliocene-Early Pleistocene (Thomas and Shaw, 1991*a*). These breaks in sedimentation broadly correlate with periods of hingeline uplift, although Summerfield (1985*a*) concludes that there is insufficient evidence to clearly define separate periods of uplift and sedimentation.

(d) Kalahari Group lithologies

Whilst there are considerable local and regional variations in the thicknesses and types of sedimentary sequence throughout the Kalahari Basin, the sequence is broadly as follows (after Thomas, 1988*a*; Thomas and Shaw, 1991*a*):

Basal gravels and conglomerates (up to 90 m) - these are found sporadically at the base of the Kalahari Group and were deposited by the endoreic early Kalahari drainage (Smit, 1977). Deposits are best developed in the southwest Kalahari (Du Toit, 1954; Mabbutt, 1955; Smit, 1977) with the lithologies present within the gravels representing both regional and local provenance (Du Toit, 1954; Shaw and De Vries, 1988). Deposits are found with a variety of matrix material, including clay, calcrete and silcrete. Gravels are also found higher up the succession in relation to valley terrace sequences (Crockett and Jennings, 1962, 1964, 1965).

Marls (up to 100 m) - these are best developed in the southwest Kalahari, but are impersistent and vary in thickness (Boocock and Van Straten, 1962; Smit, 1977). In places it is difficult to distinguish Kalahari marls from altered Karoo sandstones (Farr *et al.*, 1981).

Sandstones and duricrusts - above the marls are a mixture of highly variable orthoquartzites and duricrusts, comprising silcretised sandstones with silcrete, calcrete, sil-calcrete, cal-silcrete, clay and ferruginous horizons. Duricrusts are developed at a number of levels and in a wide variety of structural connotations, with Helgren (1984), for example, identifying over a dozen silcrete facies in the Makgadikgadi area alone. Kalahari duricrusts are discussed more fully in Chapter 6.

Kalahari Sand - the spatially extensive unconsolidated Kalahari Sand mantles all underlying Kalahari Group lithologies and varies in colour, mineralogy and provenance. Baillieul (1975) identifies four distinct sand types, and suggests that aeolian processes were probably important in their deposition, with rivers and local weathering of sandstone also contributing. The greatest thicknesses of sand occur in northern Namibia with 200-300 m of sand covering Kalahari marls and gravels (Thomas, 1988*b*), and in the southwest Kalahari, where up to 30 m of dune sand overlie an ancient calcareous surface (Mabbutt, 1955). The sand is largely composed of quartz grains and varies in colour from red (Maufe, 1930) or ochre (Cahen and Lepersonne, 1952) to bleached white (Wright, 1978). Various theories concerning the provenance and sedimentary characteristics of the Kalahari Sand are discussed by Thomas and Shaw (1991*a*) based upon studies by Baillieul (1975), Mallick *et al.* (1981) and Thomas (1984*b*, 1988*b*) amongst others.

Lacustrine and alluvial deposits - the most recent sediments are localised alluvial, lacustrine and pan deposits. These are found most extensively in the Okavango-Zambezi swamp areas, the Makgadikgadi, Ngami and Mababe depressions, and on the floors of Kalahari *mekgacha*. Sediments associated with these landforms include mainly silts and clays mixed with Kalahari Sands, but other deposits such as diatomaceous earths (Passarge, 1904; Rogers, 1936; Smit, 1977; Coates *et al.*, 1979; Shaw, 1985*a*; Heine, 1987; Shaw and Thomas, 1988), evaporites (Shaw and Thomas, 1989) and shell beds (Passarge, 1904;

Rogers, 1936; Heine, 1982, 1987; Shaw, 1985*a*; Shaw and Cooke, 1986; Shaw and Thomas, 1988) are also common.

3.3 Groundwater and hydrogeology in the Kalahari

The use of groundwater in the Kalahari is essential for the survival of the human population and of livestock, with an estimated 75% of Botswana's human and cattle population dependent at least partially upon groundwater for their water supply (Hyde, 1971). As has been noted in the preceding chapters, an understanding of regional hydrology and hydrogeology is important in the assessment of the role of groundwater in *mekgacha* development.

3.3.1 Kalahari hydrogeology

The hydrogeology of the Kalahari is perhaps best considered by means of a systematic study of the water-bearing strata known from boreholes. The Kalahari Group sediments are the only formation to contain water in primary porosity (i.e. as porewater) with all other lithologies storing water in secondary porosity; joints, fissures and fractures (Hutton and Loenhert, 1977). The following summaries are based largely upon Boocock and Van Straten (1962), Jennings (1969, 1974), Hyde (1971), Farr *et al.* (1981), and Republic of Botswana/VIAK (1984*a*,*b*,*c*) in stratigraphic order after Key (1977) and Thomas and Shaw (1991*a*).

(a) Precambrian Groups

Precambrian lithologies are generally regarded as poor aquifers, with the exception of the following; the Transvaal Sequence where groundwater occurs in fractured quartzites, but only in areas where rock is exposed or where sand cover is thin and impersistent; the Waterberg Group which yields good quality water (up to 500 mg l^{-1} dissolved solids) from sheared quartzitic sandstones; and the Ghanzi Group where groundwater occurs mostly in fissures along the strike of quarzites or intercalated shale beds, with some supplies from intercalated lava beds. In all cases, supplies are generally best adjacent to areas of recharge and where sand and Kalahari Group cover is thin.

(b) Karoo Sequence

In terms of groundwater yield, the Karoo Sequence is the most important geological unit within the Kalahari, principally because it underlies a large area. In addition to this lithologies are generally stratiform and contain a significant proportion of unmetamorphosed arkosic sediments (Farr *et al.*, 1981). The hydrogeology of the Groups within the Karoo Sequence (Smith, 1984) is as follows;

Dwyka Group - no potable supplies have been found in this group, with any groundwater being highly saline.

Ecca Group - this series is an important aquifer, although Upper Ecca shales generally give poor yields. However, supplies from the arenaceous Middle Ecca have been tapped by many boreholes e.g. in the region of Letlhakeng, Kweneng District, where rocks outcrop in dry valley networks (probably providing the main area of recharge for these beds in the eastern Kalahari). In the central Kalahari, supplies are commonly encountered at depths of over 100 m beneath a thick cover of Kalahari sands, and provide an important supply for the cattle route from Ghanzi to the railheads at Lobatse (Boocock and Van Straten, 1961). The Ecca Group is a complex aquifer system with generally low salinities, in the range 300-700 and 50-100 mg l^{-1} (Foster *et al.*, 1982). A detailed account of the Ecca aquifer can be found in Farr *et al.* (1981).

Beaufort Group/Tlhabala Formation - this unit does not constitute a major aquifer, but does act as an impermeable upper aquiclude to the underlying Ecca aquifer.

Stormberg/Lebung Group - within the Stormberg Group, the contact zone between the Cave or Ntane Sandstone and Stormberg Lava Group provides one of the best water bearing zones in Botswana (Hutton and Loenhert, 1977). The Cave/Ntane Sandstone is important to the south and southeast of the Okwa Valley, whilst the Stormberg Lava gives best yields in the vicinity of the Makgadikgadi Depression, even beneath a thick mantle of Kalahari Sands. Of the two stages, the Cave Sandstone is the more important aquifer, as discussed by Farr *et al.* (1981).

(c) Kalahari Group

Groundwater occurs in limited amounts in calcareous sandstones and conglomerates, where they underlie surface exposures of duricrusts. However, the Kalahari Group does not constitute a major aquifer, despite its vast areal extent. The major use of Kalahari Group aquifers is in the Boteti-Rakops and Nata areas of Botswana where thin lenses of fresh water occur above deeper saline groundwaters (Coates *et al.*, 1979). In the central regions of the Kalahari, supplies are often seasonal and only found in the vicinity of pans and drainage lines, with perched water tables being commonplace. Basal Kalahari marls provide good potable supplies of groundwater in south-eastern Botswana, whilst in the southern Kalahari of Gordonia, a number of localised closed basins exist with unconfined to semi-confined aquifers (Verhagen, 1983).

3.3.2 Groundwater movement

At present groundwater movement in the Kalahari is only partially understood, and then only for parts of the populated eastern margin, primarily because of the extensive cover of Kalahari Sands over much of the country. Groundwater is relatively plentiful in the northern Kalahari, particularly near to the Okavango Delta. However, this groundwater resource is spatially limited since the Kunyere and Thamalakane faults at the distal end of the Delta effectively act as a barrier, preventing groundwater movement southwards (Wilson and Dincer, 1976; UNDP/FAO, 1977).

Studies in the eastern Kalahari, particularly in the vicinity of Mochudi village (Buckley and Zeil, 1984; Zeil *et al.*, 1991), suggest that aquifers in pre-Kalahari bedrock are mostly linear and essentially vertical features. Groundwater occurrence is usually restricted to linear features associated with intrusion and faulting, with aquifers having a channel-shaped form (i.e. they are narrow and elongate and can be considered as vertical to sub-vertical features). These elongate aquifers can be identified by standard seismic and magnetic techniques, and are ideal locations for borehole siting (Farr *et al.*, 1981). The aquifer system can be regarded as consisting of main groundwater flowpaths occupying the major fracture and crush zones, with lesser flow in minor joints away from the main fractures. Groundwater movement is negligible in rock masses with little or no structural deformation away from the main fracture zones.

Highest rates of groundwater circulation were found by Buckley and Zeil (1984) to occur within main aquifer channels, although these were not necessarily the most obvious lineaments (usually shear zones) identified on aerial photography and satellite imagery (VIAK Ab, 1983). The major circulation of groundwater occurs through tensional joints and fractures which are related to the major shear zones. Circulation rates were found to be highest at the intersection of fractures, which suggest that these intersections must lie within a tension zone (Buckley and Zeil, 1984).

Dietvorst *et al.* (1991) have recently questioned primary aquifer control by fracture systems, at least for the extreme southeast of Botswana. Here, major aquifers are confined to lows provided by a "chess-board" pattern of intersecting undulations in the pre-Kalahari Basement rocks. The major Zoetfontein Fault forms an aquifer only where it intersects such tectonic basins, although small fractures do however, locally enhance permeability.

3.3.3 The question of recharge

As a result of the increasing reliance upon groundwater resources, the question of whether groundwater replenishment can occur within an environment such as the Kalahari is an important issue. This is particularly as a result of fears that supplies of potable groundwater would be disastrously depleted with the increased use of deep boreholes. It is also of importance in explaining why water tables are presently at great depth beneath the floors of many *mekgacha*.

Two conflicting arguments exist regarding the question of recharge. Until recently, the general concensus was that beyond a thickness of 4 to 15 m the Kalahari Sands prevent diffuse recharge (Passarge cited in Mazor, 1982 p.139; Debenham, 1948; Wayland, 1953; Van Straten, 1955, 1963; McConnell, 1956; Boocock and Van Straten, 1962; Baillieul, 1975, using permeability curves derived by Masch and Denny [1965]; Foster *et al.*, 1982). However, it has also been suggested that localised replenishment is possible, especially in the vicinity of pans and ephemeral drainage lines where sand cover is thin and duricrust outcrops occur (Boocock and Van Straten, 1961; Foster *et al.*, 1982; Verhagen, 1983; Arad, 1984). Other studies suggest that there is some replenishment of Kalahari aquifers although the extent of recharge is not known (e.g. Mazor *et al.*, 1974; Verhagen *et al.*, 1974; Foster *et al.*, 1982; Mazor, 1982; Verhagen, 1983; Arad, 1984; De Vries and Von Hoyer, 1988).

The main evidence has come from isotopic analyses of pumped groundwaters, with analyses of Tritium (^{3}H), ^{13}C, ^{14}C and ^{18}O carried out in the southern Kalahari (Verhagen *et al.*, 1978; Verhagen, 1983, 1984) and the central Kalahari Basin (Mazor *et al.*, 1974, 1977; Verhagen *et al.*, 1974; Foster *et al.*, 1982; Mazor, 1982). These studies suggest that recharge occurs mainly during years of above-average rainfall, with vertical recharge commonly occurring along preferential lines. However, these results must be regarded with caution due to possible biasing of borehole sampling locations and wellhead contamination (Foster *et al.*, 1982) and no actual rate of recharge can be formulated.

Studies of groundwater nitrate levels, salinity and chemical composition have been carried out in a number of Kalahari aquifers. Where waters are highly saline, this has been used to suggest the antiquity of groundwater supplies and a lack of active recharge (Debenham, 1948; Hyde, 1971). Conversely, the low levels of groundwater salinity in Kweneng District have been used by Foster *et al.* (1982) to indicate periodic flushing of groundwater during a number of periods in the past. Studies of nitrate levels in the Stampriet aquifer near to the Auob headwaters indicates significant groundwater recharge between 35,000 to 26,000 years BP and 14,000 to 8,000 years BP (Heaton *et al.*, 1983).

A variety of models have been suggested for regional groundwater movement and recharge (e.g. Farr *et al.*, 1981; De Vries, 1984, 1985) with Hutton and Loenhert (1977), Foster *et al.* (1982) and Jones (1982) following Grey and Cooke's (1977) suggestion of recharge during pluvials tentatively dated at 18,000-13,000 and 2,500-750 years BP. De Vries (1984) concludes that the last major period of general recharge ceased at around 12,500 years BP. Farr *et al.* (1981) propose a historical recharge-discharge model which suggests that there would be recharge to the Upper Karoo aquifers around the Kalahari Basin margins (as well as within the Okwa Valley, near the Ghanzi hills, and in the *mekgacha* in the vicinity of Letlhakeng) and subsequent flow towards the Makgadikgadi Depression. The lower Ecca aquifer would be recharged from the lake and would produce flow in an opposite direction, probably discharging into the rivers flowing along the easterly edge of the Kalahari basin.

3.4 The landforms of the Kalahari

The landforms of the Kalahari are a product of a variety of geomorphological processes including the action of wind and water on the land surface, and chemical activity due to the movement of groundwater beneath the surface. These processes have operated within a framework of tectonic and climatic change to produce the distinctive Kalahari landscape of today.

Kalahari landforms have attracted the attention of scientists since the time of Livingstone (1858), with the work of Schwarz (1920), Du Toit (1926*a*) and Kokot (1948) further developing interest in the region. Our knowledge of the landforms of the Kalahari has greatly improved since the availability of remotely-sensed imagery provided an overview of the landscape (e.g. Mallick *et al.*, 1981). Particular interest has been given to landforms suggesting climatic and environmental change, first described in studies by Du Toit (1926*a*), Wayland (1954), McConnell (1959*b*) and Grove (1969). Since then work has focussed on specific aspects of the landscape, including fluviolacustrine features (e.g. Cooke 1977, 1979*a,b*, 1980, 1984; Cooke and Verstappen, 1984; Heine, 1978*a*; Lancaster, 1978*a,b*; Shaw, 1985*a,b*, 1988*a*), cave landforms and sediments (e.g. Cooke, 1975, 1984; Cooke and Verhagen, 1977; Brook *et al.*, 1990) and dunes (e.g. Goudie, 1970; Lancaster, 1980, 1981, 1988; Thomas, 1984*a*, 1988*a*). This section briefly summarises studies of these landforms, together with an outline of their possible palaeoenvironmental significance and a chronology of Kalahari landform development.

3.4.1 Fluviolacustrine landforms

An understanding of the drainage systems and lacustrine landforms in and around the Kalahari region is essential in any consideration of *mekgacha* development. Considerable changes in the configuration of the regional drainage of southern Africa have occurred during the timescales over which *mekgacha* have developed. Additionally there is well-documented evidence for former extensive lakes in the Makgadikgadi, Ngami and Mababe depressions during the Late Quaternary, which relate to both tectonic and climatic changes as well as diversions in regional drainage.

(a) Drainage evolution and the perennial rivers of the Northern Kalahari

(i) Drainage evolution.

The division of Gondwanaland had a major impact upon the drainage of southern Africa, with coincident development of a dual system of endoreic and exoreic rivers (figure 3.2). Relatively short rivers with steep gradients drained from the seaward side of the Great Escarpment to the coast, whilst longer, lower gradient rivers drained inland to the Kalahari Basin (De Swardt and Bennet, 1974; Thomas and Shaw, 1988). It was these endoreic rivers which initiated deposition of the Kalahari Group sediments (Thomas and Shaw, 1990), under conditions that were probably hydrologically different to today. In particular, Dardis *et al.* (1988) note that botanical (particularly the absence of angiosperms) and climatic differences at this time would have resulted in potentially higher rates of erosion and sediment yield than the present day.

Since the establishment of this dual drainage system, the more aggressive exoreic rivers have progressively captured the internally draining ones (De Heinzelin, 1963). This process of capture has been described by many studies, using evidence from a variety of river systems (e.g. the Molopo-Vaal-Orange system by Smit, 1977; McCarthy, 1983; the Fish River by Wellington, 1955; the Limpopo and Zambezi by Bond, 1963; Du Toit, 1926*a*, 1933; Thomas 1983/1984; Moore, 1988; Thomas and Shaw, 1988, 1992; Nugent, 1990, 1992). Of these studies, the development of the Zambezi and Limpopo, and the Molopo-Vaal-Orange systems are of most relevance to Kalahari *mekgacha*. These will now be considered briefly; full reviews are provided by Thomas and Shaw (1991*a*) and Dardis *et al.* (1988).

The modern course of the Zambezi River is believed to have been established as recently as the mid-Pleistocene (Bond, 1975), as a result of the joining of the independently developed Upper and Middle Zambezi Rivers. Evidence for this has been described elsewhere (see Thomas and Shaw, 1988, 1991*a*, for recent summaries), but is supported by a number of lines of geomorphological evidence as well as studies of fish populations (Bell-Cross, 1975). Many observers have suggested that the Zambezi once flowed on a more southerly course, joining the Orange (Lister, 1979) or, more probably, the Limpopo (Du Toit, 1927, 1933; Bond, 1963; Wellington, 1955).

A proposed pattern of development of the Zambezi is shown in figure 3.5 (after Thomas and Shaw, 1988). The development of the individual Upper and Middle Zambezi Rivers was disrupted by regional uplift along the line of the Kalahari-Zimbabwe Axis and downwarping in the Okavango and Makgadikgadi regions (Du Toit, 1926*a*; Bond, 1963). A complex series of tributary captures are proposed (Bond, 1963; Williams, 1975; Wellington, 1955; Lister, 1979; Thomas and Shaw, 1988), with the drainage pattern of today ultimately emerging. Renewed uplift along the Kalahari-Zimbabwe Axis is proposed as a reason for the erosion of the Victoria Falls and the Middle Zambezi Gorge (Bond, 1975), with the Victoria Falls representing the current position of receding nick-point. The movements of the Kalahari rim also affected drainage in the southern and eastern Kalahari, with Wellington (1955) suggesting that the Molopo-Auob-Nossop valley network has lost its headwater areas due to capture by the Fish River in Namibia and the Limpopo to the east. Uplift of the Kalahari-Zimbabwe Axis since the late-Tertiary (Cooke, 1980) is also proposed to have caused the gradual incision of the headwaters of the Mmone/Quoxo *mekgacha* system (Shaw and De Vries, 1988).

On a grander scale than the development of the Zambezi is the proposed Trans-Tswana River which is thought to have once drained much of southern Africa entering the Orange River from the north (McCarthy, 1983). Figure 3.6 (after Dardis *et al.*, 1988) shows the suggested course of this river, together with superimposed *mekgacha* networks, which is supposed to have existed prior to the establishment of the present course of the Zambezi in the late Pliocene or early Pleistocene (Thomas and Shaw, 1988). The suggestion was originated by Schwarz (1920) and has been substantiated by the work of McCarthy (1983) on the basis of studies of the provenance of gravel deposits in the vicinity of the Orange River. McCarthy (1983) proposes that this system was disrupted by crustal flexuring, particularly along the Griqualand-Transvaal and Kalahari-Zimbabwe axes, and further suggests that the Okavango Delta, Makgadikgadi and Ngami Basins are the only remaining remnants.

(ii) Perennial rivers of the northern Kalahari.

Due to higher available precipitation levels, the northern Kalahari contains all of the Kalahari's perennial rivers. These systems are all allogenic (i.e. they rise in sub-humid or humid areas beyond the margins of the desert; Wilkinson, 1988) and flow through the northern Kalahari. They can be grouped from west to east as follows; the Cubango-Cuito-Okavango, the Kwando-Chobe and the Zambezi. It is not proposed to describe these river systems in detail; where they impinge upon other areas of study within the Kalahari they are discussed in the following section.

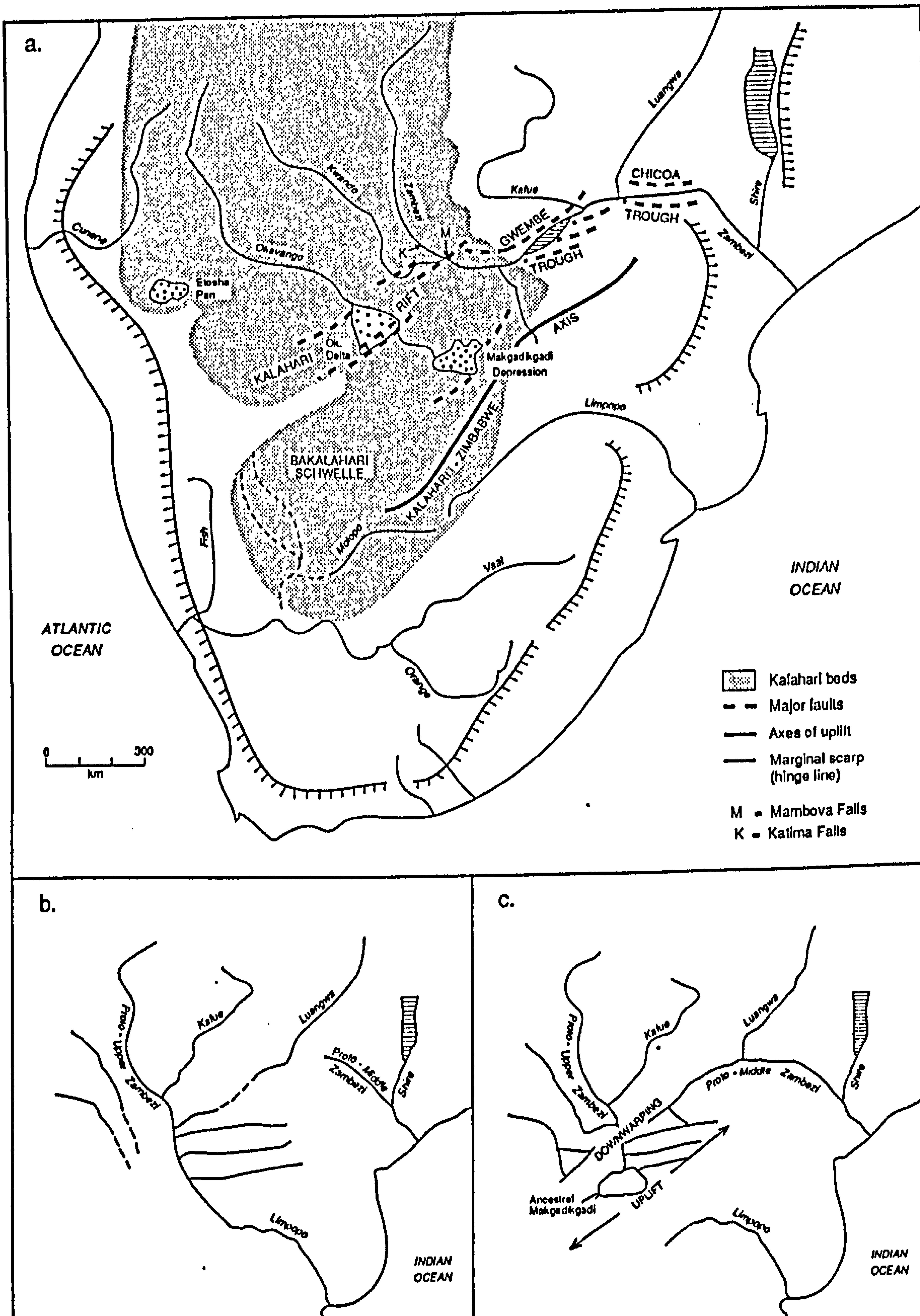


Figure 3.5: Southern African drainage development (from Thomas and Shaw, 1991*a*; after Thomas and Shaw, 1988): (a) Modern drainage lines and areas of crustal uplift and downwarping, (b, c) The development of the Zambezi drainage system (b; prior to the breakup of Gondwanaland, c; prior to the joining of the Middle and Upper Zambezi in the early Pleistocene).

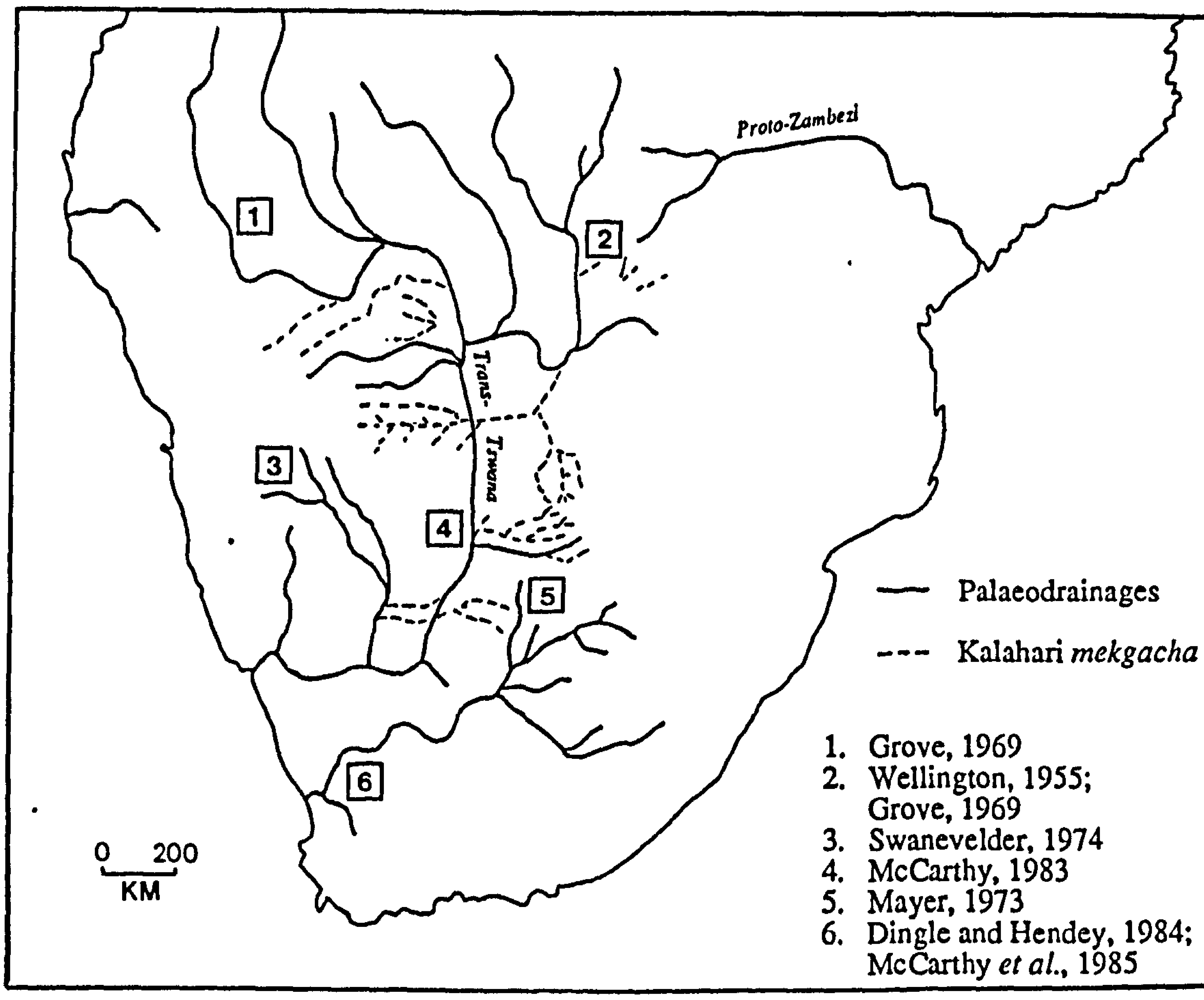


Figure 3.6: The course of the proposed Trans-Tswana palaeodrainage system (after Dardis *et al.*, 1988), plus the present courses of major Kalahari *mekgacha.*

(b) Lakes and channels of the Middle Kalahari

The landforms of the Middle Kalahari are dominated by the swamplands of the Okavango Delta and Chobe River, and the Makgadikgadi, Ngami and Mababe Basins. The complexity of the landforms of this area have long been recognised and described (Livingstone, 1858; Passarge, 1904; Schwarz, 1920; Du Toit, 1926*a*; Wellington, 1955; Grove, 1969) with more recent studies by Heine (1978*a*, 1979) Shaw (1984, 1985*a,b*, 1986, 1988*b*), Shaw and Cooke (1986), Shaw and Thomas (1988) and Shaw *et al.* (1988). They are reviewed by Shaw (1988*a*, 1989) and Shaw and Thomas (1989).

As has been mentioned in section 3.4.1*a* above, neotectonic movements have played a major role in shaping the drainage of the Middle Kalahari. Together with climatic changes, tectonic movements once led to the formation of Lake Palaeo-Makgadikgadi, encompassing the Ngami, Mababe and Makgadikgadi Basins, the Okavango Delta and parts of the Chobe-Zambezi confluence area, covering a total area in excess of 60,000 km^2 (Cooke, 1980). The component parts of Lake Palaeo-Makgadikgadi will now be briefly considered, although the chronology of the development of the palaeolake is discussed in section 3.4.4.

(i) The Okavango Delta

The Okavango Delta is formed by the dispersal of the fault controlled Okavango River close to Seronga in Botswana, into a series of anastomosing distributaries. The delta, essentially an alluvial fan (or more correctly, a delta-fan; McCarthy *et al.*, 1988*a*), covers an area of between 6,000 and 13,000 km^2, depending upon the time of year, precipitation characteristics and flood conditions (UNDP/FAO, 1977; Shaw, 1988*a*). The Okavango Delta is terminated at its distal end by the Kunyere and Thamalakane Faults.

Flow in the delta varies annually, but it is estimated that the mean annual water budget is approximately 15.5 x 10^9 m^3 yr^{-1} of which some 96% is lost to evapotranspiration or groundwater (UNDP/FAO, 1977). Major fluctuations have occurred in the hydrological regime of the delta (Shaw, 1984), with routes of flow affected by minor tectonic movements (Pike, 1971*b*) as well as channel blockages due to sedimentation, vegetation and animal- or human-induced diversion (McCarthy *et al.*, 1986, 1988*b*).

Of greatest interest to this study is the extension of older alluvium to the west of the Okavango Delta, which terminates the Xaudum, Qangwadum, Eiseb, Gcwihabedum and Groot Laagte. The Groot Laagte has deposited deltaic sediments over an area of some 4,600 km^2 to the west of these older Okavango alluvial sediments (Thomas and Shaw, 1991*a*).

(ii) The Ngami and Mababe Basins

These depressions are located at the southwestern (Ngami) and northeastern (Mababe) ends of the Kunyere-Thamalakane fault system, with both basins developing as a result of downthrow along the fault line. The basins have respective areas of 1,800 km^2 and 3,000 km^2, with the Ngami Basin linked to the Okavango Delta by the Thamalakane, Nchabe and the fossil Thaoge River systems, and Mababe linked via the Thamalakane-Mokhokelo, Khwai and (tenuously via) the Magwegqana-Savuti channels (Shaw, 1985*a*, 1986). This complex linkage of drainage is due to the tectonically unstable nature of the area, with rivers particularly susceptible to disruption or reversal due to their low gradients (Shaw, 1985*a*).

A series of concentric former beach ridges and shorelines have been identified in the western and southwestern areas of both depressions. In Ngami, the shorelines are collectively known as the Dautsa Ridge with the comparable feature in the Mababe Basin termed the Magikwe Ridge, the ridge complexes extending for 25 km and 100 km respectively. In both basins, ridges are developed at a number of levels, the levels common to both being 940-945 m asl, 936 m, 930 m, 926 m and 923 m, with additional ridges in Ngami at 934 and 928 m and in Mababe at 929 and 927 m (Thomas and Shaw, 1988, 1991*a*) and a common sump level at 919 m asl. The relationships between lake levels and channels in the area are discussed in Shaw (1985*a*).

(iii) The Makgadikgadi Basin

The Makgadikgadi Depression covers an area of 37,000 km^2, linked to the Okavango by the Thamalakane and the misfit Boteti Rivers (Cooke, 1980). A number of channels and *mekgacha* feed into or are directed towards Makgadikgadi, including the water bearing Boteti and Nata Rivers from the north, the Okwa and Letlhakane from the south, the Passarge and Deception from the west and the Semowane, Mosetse and Lepashe sand rivers from the east. The orientation and form of the basin is controlled by a series of SW-NE Basement and Karoo structures (MacGregor, 1930, 1931; Baillieul, 1979; Cooke, 1980; Smith, 1984), as clearly defined by the shape of the main Ntwetwe and Sua Pans which occupy the bulk of the depression.

The basin contains a suite of palaeolake shorelines, on a much larger scale than the Ngami and Mababe depressions, dominated by the 250 km long Gidikwe Ridge (Grove, 1969). Major levels have been identified at 940-945 m, 920 and 912 m, with a sump at 890 m, and additional levels at 904, 905 and 908 m in Sua Pan and 910 m in Ntwetwe Pan (Grey and Cooke, 1977; Ebert and Hitchcock, 1978; Cooke, 1979; Cooke, 1980; Mallick *et al.*, 1981). It is probable that the Gidikwe, Dautsa and Magikwe Ridges all formed as offshore bars, as evidenced by sediment characteristics and gastropod species (Cooke, 1980).

Of greatest interest in terms of this study are the landforms of western Makgadikgadi, particularly a series of deltaic deposits built up by the Boteti River and the Okwa, and an area of lagoonal sediments to the west of the Gidikwe Ridge which terminate the Passarge and Deception valleys but which the Okwa crosses (Cooke and Verstappen, 1984). The sequence of ridges within Makgadikgadi represent fluctuating lake levels, which only the Okwa and Boteti break. This suggests that both rivers had sufficiently stable regimes to enable them to breach the ridge (Cooke, 1980), although the Boteti shows at least three levels of ponding indicating that the Gidikwe Ridge has, at times, acted as a barrier to flow (Shaw *et al.*, 1988).

(iv) The Chobe-Zambezi confluence: Lake Caprivi

Water levels in the Chobe and Zambezi rivers are controlled by the Mambova Falls, a basalt dyke at 926 m asl which acts as a bottleneck to flow during periods of flood. A large area of younger alluvium associated with this ponding of the Zambezi is identifiable on satellite imagery. Shaw and Thomas (1988) have identified an additional area of older alluvium in the Chobe-Zambezi confluence area, with associated ridges at 932-936 m asl and a terrace level at 934 m. The rivers of northern Botswana show clear evidence of structural control, being orientated along a series of NW-SE and SW-NE faults giving a rectilinear pattern.

(v) The Middle Kalahari palaeolakes: a summary

From the numerous lines of data collected on the palaeolakes of the Middle Kalahari, two main lake stages have been identified. The higher 940-945 m stage, Lake Palaeo-Makgadikgadi, covers all three depressions described above, as well as the lower Okavango Delta and probably parts of the Zambezi trough (Grey and Cooke, 1977; Shaw *et al.*, 1988), an area of up to 80,000 km^2 (Mallick *et al.*, 1981). A second stage is identified at the 936 m level, linking the landforms in the Ngami and Mababe Basins along the Thamalakane fault line with Lake Caprivi, with overflow to the 920 or 912 m levels in the Makgadikgadi Basin. Shaw (1988*b*) terms this the Lake Thamalakane Stage, with an estimated area of 7,000 km^2 in the delta region alone. The chronology and interpretation of this sequence of palaeolakes is discussed in section 3.4.4.

3.4.2 Caves and cave sediments

Cave systems are found in the calcareous rocks of the Damara Sequence to the west of the Okavango Delta (Pickford and Mein, 1988) and in the Proterozoic Transvaal Sequence of the southern and southeastern Kalahari margin. The most important cave site is Drotsky's Cave in the Gcwihaba Hills of northwest Botswana, described in detail by Cooke (1975). The horizontal, predominantly solutional cavern system contains evidence of four distinct cycles of sinter formation separated by aeolian sediments, which suggest possible arid intervals. Two deep sinkholes have been found in the nearby Aha Hills (Cooke and Baillieul, 1974) but these contain no significant deposits.

In southeastern Botswana, caves in the Lobatse area have yielded flowstones and sinters, with recent palaeoenvironmental studies of this site summarised by Shaw and Cooke (1986). The caves of the Gaap Escarpment on the southern margin of the Kalahari in South Africa have been studied both for geomorphological interest (e.g. Butzer, 1974; Butzer *et al.*, 1978) and by archaeologists in conjunction with discoveries of the "Taung child", a skull of *Australopithecus africanus*, in Hrdlicka Cave at Taung (Peabody, 1954). Other cave sites in South Africa include Wonderwerk Cave in the Kuruman Hills (described by Beaumont *et al.*, 1984; Butzer, 1984*a,b*; Peabody, 1954), Echo Cave in the Transvaal (Brook, 1982) and Equus Cave at Taung (Klein *et al.*, 1991).

3.4.3 Aeolian landforms

The extensive fields of fixed sand dunes in southern Africa have been studied by a variety of authors (e.g. Du Toit, 1927; Lancaster, 1979*a*; Thomas, 1984*a*; Thomas and Goudie, 1984). Systems of parallel, mainly linear, often degraded, vegetation-fixed dunes occur throughout the Kalahari from the Orange river to southern Angola and southwest Zambia (Lancaster, 1981, 1984), although they are less apparent in central areas. Three systems are recognised (Thomas, 1984*a*), forming a broadly arcuate pattern (Goudie, 1970) with a radius of around 1,000 km (Lancaster, 1979*a*). The dune fields and their sediments have been described by numerous authors including Lewis (1936), Grove (1969), Goudie (1969, 1970), Coates *et al.* (1979), Lancaster (1979*a*, 1980, 1981, 1984), Mallick *et al.* (1981), Thomas (1984*a*, 1987*a*, 1988*a*) and Cooke (1984). Recent work has concentrated on aspects of the sedimentology, morphology and morphometry of the dunefield (e.g. Lancaster, 1986*a*, 1988; Thomas, 1988*a*; Thomas and Martin, 1987) and has questioned aspects of their "fossil" status (Thomas and Tsoar, 1990; Thomas and Shaw, 1991*b*; Thomas, 1992). Lunette dunes found on the margins of many pans in southern Africa have been discussed by Grove (1969), Lancaster (1978*a,b*, 1979*a,b*) and Goudie and Thomas (1985, 1986).

3.4.4 Chronology and interpretation of landforms

As has been noted, the most common application of evidence from landforms, sediments, faunal and archaeological remains in the Kalahari has been in establishing a chronology for landform development, particularly in relation to climatic changes. The majority of this work, with certain exceptions, has been restricted to the last 100,000 years, the maximum timespan within which radiocarbon dating can be applied. However, there is some evidence for conditions prior to the Late Quaternary. The palaeoenvironmental inferences which can be made from Kalahari landforms will be considered for the remainder of this chapter. For further consideration of the evolving chronology and interpretation of southern African environmental changes see Van Zinderen Bakker and Coetzee (1972), Van Zinderen Bakker (1976, 1980, 1982), Nicholson and Flohn (1980), Klein (1984), Vogel (1984, 1985, 1989), Partridge (1985) and Williams (1985).

(a) Conditions prior to the Late Quaternary

Evidence for conditions within the Kalahari during the Cretaceous to the Late Quaternary (i.e. the period during which the Kalahari Group sediments have been deposited) is generally fragmented and difficult to interpret. The sequence of Kalahari Group sediments from basal conglomerates to aeolian deposits has been suggested to indicate increasing aridity in the region (e.g. Smit, 1977). This interpretation is difficult to support primarily because the deposition of gravels may be more closely related to tectonic changes and the development of the drainage network in southern Africa than to climatic changes. Furthermore, deposits may be more closely related to climates within source areas than the Kalahari *per se*, and need to also be placed within palaeobotanic contexts, as noted above. Some of these source areas may even have occurred beyond continental Africa prior to the breakup of Gondwanaland (Dardis *et al.*, 1988). There is however, evidence from cave sites to the west of the Okavango Delta to support increasing aridity (Cooke, 1975, 1980), suggested by the presence of windblown sand fills dating to the late Tertiary near the base of cave deposits in the Gcwihaba Hills. The caves themselves must have developed by solution during earlier more humid conditions (Cooke, 1980).

A more complex picture of changes prior to the Quaternary has been indicated from sites around the southern Kalahari margin, which suggest fluctuating conditions. The sequences of tufas from the Gaap Escarpment interspersed with other deposits suggests a variety of environmental conditions including high intensity rainfall, cold periods conducive to the operation of freeze-thaw processes, spring activity, semi-arid conditions and perennial streams (Butzer *et al.*, 1978; Butzer, 1984*a*,*b*). In total, six facies changes can be identified form the Gaap Escarpment dating from the Late Tertiary to the Pleistocene, with the last period of tufa formation occurring around 250,000 years BP (Vogel *in* Butzer, 1984*a*). Complex environmental changes are also recognised from sedimentary sequences deposited in dolines developed within calcretes at Kathu Vlei in the Northern Cape (Butzer, 1984*a*,*b*). Peat deposits are suggested to indicate wet condition, with periods of increased aridity implied by ash deposits resulting from peat fires and flash-flood deposits. These sequences have been dated by means of archaeological and faunal remains.

(b) Conditions during the Late Quaternary

(i) Valley sites

The main studies of Kalahari valley sites are from the Dobe Valley to the west of the Okavango Delta and the Auob-Nossop-Molopo confluence area in the southwestern Kalahari. Investigations in the Dobe Valley (Helgren, 1978; Helgren and Brooks, 1983) indicate the presence of lakes within the valley, the earliest dated to Middle Stone Age times and a later lake at around 23,000 to 22,000 years BP. It is proposed that these lakes dwindled to a series of interconnected pans during intervening dry periods. In the southwestern Kalahari, Heine (1981, 1982) has suggested perennial flow within the Molopo, Nossop and Auob valleys in the period 16,600 to 12,500 years BP.

(ii) Middle Kalahari palaeolakes

The sequences of strandlines and configurations of the palaeolakes which formerly occupied the Makgadikgadi, Ngami and Mababe basins have been discussed in section 3.4.1*b*. Radiocarbon dating, mostly of calcretes, but also shell material, reed casts and peat deposits has been carried out in conjunction with a number of studies (e.g. Street and Grove, 1976; Heine, 1978*a*, 1979; Cooke, 1979*b*, 1984; Cooke and Verstappen, 1984; Helgren, 1984; Shaw, 1985*a*, 1986; Shaw and Cooke, 1986; Shaw and Thomas, 1988; Shaw, Cooke and Thomas, 1988; Shaw, Thomas and Nash, 1993).

These studies provide information on the relative timing of the two main lake configurations, the 940-945 m Lake Palaeo-Makgadikgadi and 936 m Lake Thamalakane Stages. Absolute dates for Lake Palaeo-Makgadikgadi are only indicated from the Makgadikgadi Depression, where a high lake level was attained between 40,000 to 35,000 years BP. Low lake levels are indicated from 35,000 to 26,000 years BP, with a succession of lakes at the 920 m level between 26,000 to 10,000 years BP (including lows at 21,000 and 19,000 years BP). The record for the Mababe and Ngami Basins begins with low levels between 26,000 to 24,000 years BP, with the Lake Thamalakane stage occurring between 17,000 to 13,000 years BP concomitant with levels at 920 m in the Makgadikgadi Depression. Lake levels progressively fell during the period 12,000 to 9,500 years BP.

Various suggestions have been put forward for the sources of these vast palaeolakes. The Lake Palaeo-Makgadikgadi Stage is generally agreed to have required diversion of a major river such as the Zambezi to fill it (Grove, 1969). This suggests that climatic factors were not as important in its development, with specific tectonic events being largely responsible for the presence of the palaeolake. The shoreline of Lake Thamalakane has not been tectonically disrupted (Thomas and Shaw, 1991*a*) and could have been supplied by increased precipitation within the Middle Kalahari.

(iii) Cave sites

A chronology based upon radiocarbon and U/Th dates from speleothems in Drotsky's Cave within the Gcwihaba Hills have resulted from studies and discussions by Cooke (1975, 1980, 1984) and Shaw and

Cooke (1986). Drotsky's Cave is possibly the most important palaeoenvironmental location within the Kalahari, as it provides the only closed site with a major sequence of sedimentary deposition. Major periods of cave sinter deposition occurred during 45,000 to 37,000, 34,000 to 29,000 and 16,000 to 13,000 years BP, with further development between 6,000 to 5,000 and at 4,000 years BP. Cave sinters from Lobatse Cave II in southeast Botswana formed during 26,000 to 22,000 and 18,000 to 17,000 years BP (Shaw and Cooke, 1986).

Studies of cave sites at the southern Kalahari margin have also revealed evidence for environmental changes. Beaumont *et al.* (1984), Butzer (1984*a,b*), Deacon and Lancaster (1988) and Klein *et al.* (1991) summarise the findings from the Equus and Wonderwerk caves and Kathu Vlei sites. The main environmental patterns arising from these sites will be discussed below.

(iv) Aeolian activity

The interpretation of the possible palaeoenvironmental significance of the three wholly or partly vegetated dune systems is discussed by a variety of authors, including Lancaster (1980, 1981, 1988), Thomas (1982, 1984*a*, 1988*c*, 1992) and Thomas and Shaw (1991*a,b*). The dating of periods of aeolian activity is usually based on the identification of "windows" within the humid chronology (Thomas and Shaw, 1991*a*), although present studies utilising the OSL dating technique should provide absolute dates for dune activity.

A number of Quaternary arid periods have been recognised. The southwestern Kalahari dunefield is suggested to have been active prior to 32,000, between 19,000 to 17,000, between 10,000 to 6,000 and between 4,000 to 3,000 years BP (Lancaster, 1989). However, periods of activity in the Northern and Eastern dunefields of Thomas (1984*a*) are less well known, with Cooke (1975, 1980) suggesting a Late Tertiary age based upon aeolian sediment within Drotsky's Cave.

(c) Conditions during the Holocene

Holocene environments are indicated by a variety of studies. Lake levels in the Middle Kalahari were initially low, but levels rose to 932 m in Ngami at around 6,000 years BP (Shaw, 1985*a*). This date coincides with reactivation of cave sinter deposition in Drotsky's Cave (Shaw and Cooke, 1986). It is suggested that the Lake Thamalakane stage was reoccupied at 2,500 to 2,000 years BP. Wetter conditions than present are indicated from Kathu Vlei for the period 7,400 to 4,500 years BP.

During the late Holocene, periods of increased fluvial activity are recognised in southern African rivers away from the Kalahari by Smith (1991, 1992). In particular, increased flood activity in the Orange River between 1650 and 1720 AD coincided with the Little Ice Age, and floods around 2,400 years BP possibly relate to cooler wetter conditions identified elsewhere (Cooke, 1984)

(d) Chronology of landform development

Figure 3.7 shows the palaeoenvironmental information from all the major Kalahari sites mentioned in this chapter in addition to data from western Kalahari pans (Lancaster, 1986*b*; Deacon and Lancaster, 1988). The major observation which can be made from this figure is that humid phases indicated by high lake levels, flow in rivers and sinter growth in caves are not always coincident. This is in part due to the inherent climatic variability occurring over such a large area as the Kalahari, but also to the varying nature and quality of available evidence and the general reliance upon geomorphological as opposed to more sensitive palynological sources (Thomas, 1987; Nash *et al.*, 1993). The general patterns can be summarised as follows (after Deacon *et al.*, 1984; Deacon and Lancaster, 1988; Thomas and Shaw, 1991*a*);

50,000 to 20,000 years BP

There is conflicting information from Makgadikgadi, Drotsky's Cave and the Dobe Valley, which may be explicable in terms of climatic differences between the Angolan Highlands which would have supplied the palaeolake and the more local conditions affecting the area to the west of the Okavango Delta. However, humid conditions are generally indicated for 35,000 to 22,000 years BP, with regional variations including a dry period indicated at around 25,000 years BP by low lake levels.

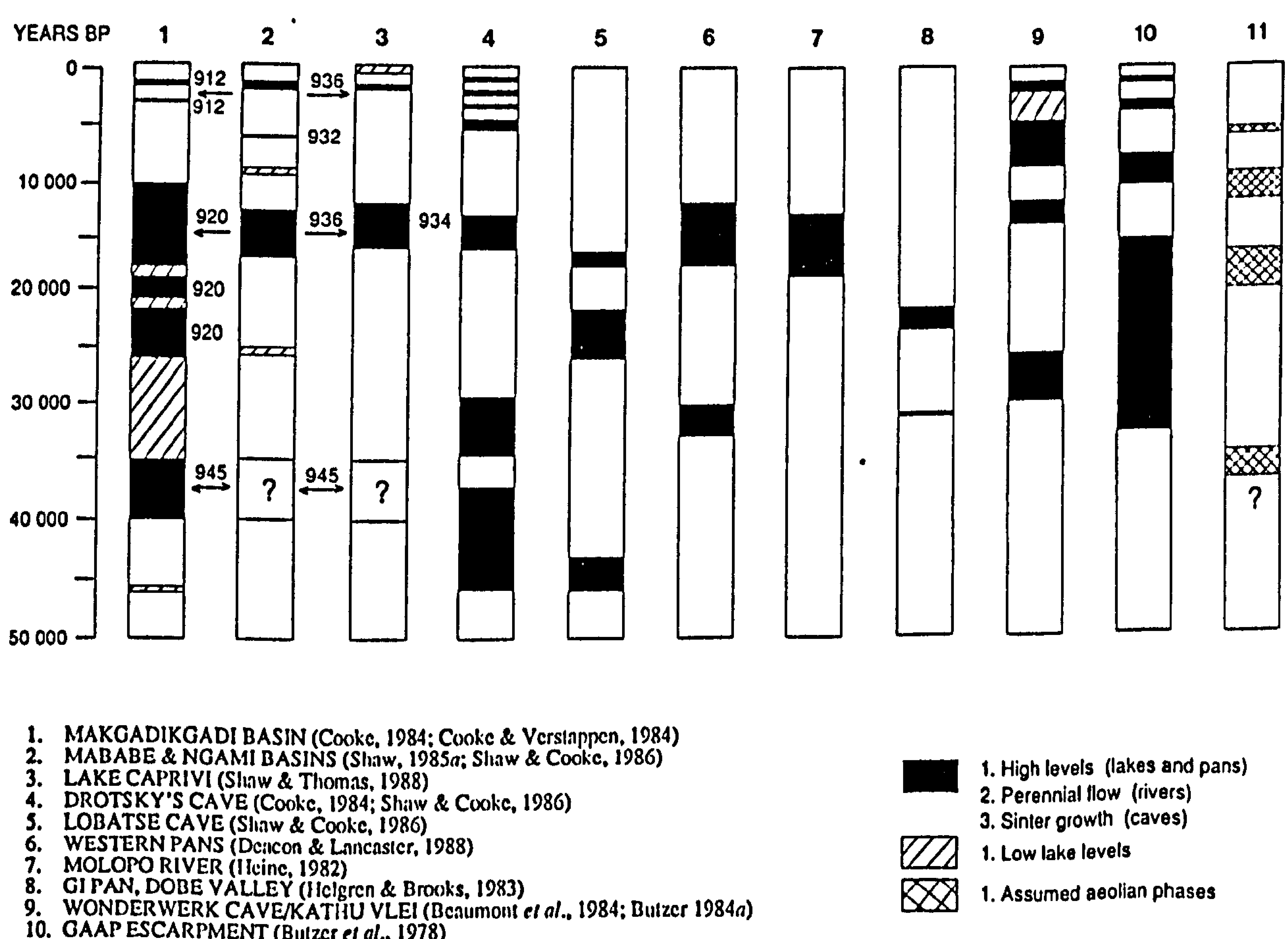


Figure 3.7: The evidence of past climates from locations within the Kalahari (after Thomas and Shaw, 1991*a*).

20,000 to 12,000 years BP

Cold dry conditions are indicated at the start of this period, apart from the Gaap Escarpment where tufa deposition continued. Between 17,000 to 12,000 years BP a wetter period occurred, probably occurring over an area from the Orange River (Kent and Gribnitz, 1985) to the Zambezi.

12,000 years BP to the present

Drier conditions returned at the start of the period over much of the Kalahari, accompanied by cooler temperatures in the south. Humid episodes occurred at Drotsky's Cave during the Holocene, as described above, with similar but not necessarily synchronous wetter periods in the southern Kalahari.

Part 2

Hypotheses, methods and analysis.

Chapter 4
Research hypotheses and methodology

4.1 Introduction to Part 2

The background to Kalahari *mekgacha* development has been described in the preceding chapters, with reference made to both past and present environmental settings and the factors involved in valley evolution. This chapter addresses the possible hypotheses for the development of Kalahari dry valleys (section 4.2) and introduces the methods of analysis used to assess their relative importance in valley evolution (section 4.3). The remaining three chapters in Part 2 give full details and results for the main methods of study, all including reference to previous research.

4.2 Hypotheses of valley development

From chapter 2, two broad hypotheses can be put forward to explain the origins and evolution of Kalahari *mekgacha*. These are that valleys formed either as a result of erosion due to former fluvial activity (from perennial rivers or low-frequency high-magnitude flood events) or through the influence of groundwater processes (deep-weathering along preferential groundwater flowpaths and erosion by groundwater seepage or sapping).

It should be stressed that these two hypotheses for valley origin are not mutually exclusive since it is possible for both processes to interact; purely "fluvial" or purely "groundwater" origins must be viewed as the two extremes of a process spectrum. Neither should it be necessary that one explanation is wholly appropriate for all localities nor at all timescales. Thus it is possible that distinct spatial and temporal variations in the significance of each process can be distinguished.

Since a background to both sets of valley-forming processes has already been given in chapter 2, with particular detail to groundwater weathering and erosive processes, the two hypotheses will only be briefly discussed.

4.2.1 Fluvial hypotheses

The traditional explanation for the presence of Kalahari valley systems is that they are a result of erosion during periods of excess available moisture with sufficient surface water to maintain permanent flowing rivers. This idea can be traced back to Passarge (1899) who postulated that the Groot Laagte in northwestern Botswana must once have contained a river comparable in size with the Okavango (i.e. a mean annual budget of 15.5×10^9 m^3 yr^1 and a possible flow rate of 85 m^3 sec^{-1} during the July flood peak; Wellington, 1955; Thomas and Shaw 1991*a*).

The suggestion that valleys were formed by fluvial erosion (particularly by perennial rivers) is invariably taken as evidence for former widespread wetter conditions; by inference, Kalahari *mekgacha* are accorded palaeoclimatic significance (e.g. Heine, 1982). This link between valleys and past hydroclimatic changes is in accordance with the traditional view of the significance of landforms in the Kalahari. As has already been discussed in chapter 3, many studies considered vegetated linear dunes as being indicative of former aridity (e.g. Lancaster, 1981), whilst palaeolakes have been cited as evidence for increased moisture availability (essentially precipitation) in the past (Grey and Cooke, 1977; Heine, 1978*a*).

However, there has been a tendency to move away from viewing Kalahari landform development in terms of simple arid to humid climatic swings, particularly with the recognition of the importance of other factors at a regional scale. Recent developments have increasingly recognised the importance of subtle tectonic activity and drainage diversion (first suggested by Du Toit, 1933) in the formation and later desiccation of Lake Palaeo-Makgadikgadi (Grey and Cooke, 1977; Cooke 1980; Shaw and Thomas, 1988). The geomorphological significance of vegetation upon dune activity has also been questioned, with controversy existing regarding the environmental and palaeoenvironmental conditions represented by vegetated linear dunes (Thomas and Tsoar, 1990).

Thus, one hypothesis for Kalahari valley development is that *mekgacha* are essentially a "fossil" landform representing a period (or periods) of greater effective precipitation. This traditional viewpoint is, of necessity, largely dependent upon geomorphological interpretations, since there is little palynological or faunal evidence for environmental change in the Kalahari (Thomas, 1987*b*). However, one recent palaeontological study of cave macrofauna (Klein *et al*, 1991) suggests that, for the southern Kalahari margin at least, climatic shifts may have been much more subtle than once suggested.

If this is the case, then a second fluvial hypothesis for *mekgacha* development may be more appropriate, namely that valleys are a product of high magnitude-low frequency flood events. Such flash-flood events (as discussed in chapter 2) are more typical of fluvial regimes in arid and semi-arid zones (Reid and Frostick, 1989), and would not necessarily need to be explained by major shifts in climate or prolonged periods of increased precipitation. Flash-floods are uncommon today in *mekgacha*, particularly in the endoreic systems, with only one documented report of flow away from headwater areas (in the Letlhakane valley in 1969; Mazor *et al*, 1977). This is almost certainly due, at least in part, to the presence of an extensive cover of Kalahari Sand both within valleys and catchment areas of these *mekgacha*. Flooding has occurred in the exoreic Auob, Nossop, Molopo and Kuruman valleys on a number of occasions during the period of historical records, as described in sections 5.3.1 and 5.3.2. Such flooding is almost certainly enhanced because these systems rise either at springs or have headwaters in upland hardveld areas. Increased erosion by flash-flooding would require a shift in climatic parameters if it were to be of significance to endoreic *mekgacha*, but not of the order needed to sustain perennial flow within valleys. However, whether this would have been the case prior to the emplacement of the Kalahari Sand is unclear.

4.2.2 Groundwater hypothesis

A third major hypothesis for *mekgacha* development was first proposed for the Kalahari by Shaw and De Vries (1988); valley systems are a result of groundwater erosion processes. This suggestion was made on the basis of studies of valleys in the vicinity of Letlhakeng, Botswana, with the recognition of a number of distinctive valley forms and unusual suites of diagenetic sediments. The atypically deep valleys with extensive sequences of duricrusts were attributed to formation by sapping processes. These valleys closely parallel subsurface fractures, and lithological samples taken from water boreholes sunk within them show evidence of substantial deep-weathering (Von Hoyer *et al*, 1985).

From section 2.3, groundwater activity can be seen as having two major contributions to valley development: through deep-weathering (Shaw and De Vries, 1988) and through seepage or sapping processes (Howard *et al*, 1988; Baker, 1990). Deep-weathering processes may influence valley location, especially where groundwater flow is concentrated along preferential flowpaths such as sub-surface fractures or faults. Groundwater sapping processes effectively erode valleys "from below" (Peel, 1941) rather than surficial erosion by fluvial processes. Valleys produced by such processes have distinctive form and network characteristics, as summarised in 2.3.1*b(iv)*.

The significance of a groundwater hypothesis in explaining *mekgacha* development is that, like erosion due to high magnitude-low frequency flood events, it does not necessarily require recourse to major hydroclimatic changes in order for processes to operate. On the contrary, Howard *et al* (1988) suggest that in areas where sapping processes are known to be operating, greater rainfall and groundwater outflow may not lead to an increased rate of erosion; this would hinder the accumulation of salts and other important mechanical weathering agents. It does, however, require water tables to be close to the ground surface, a situation which does not occur in Kalahari *mekgacha* at the present day. This is in part due to recent groundwater extraction (Thomas and Shaw, 1991*a*; Shaw *et al*, 1993) but is also a reflection of a lack of significant extensive recharge since at least 12,500 years BP (De Vries, 1984). Both groundwater emergence and subsurface movements of groundwater along preferential flowpaths (such as linear aquifers developed in fractures; Buckley and Zeil, 1984) are closely linked to bedrock permeability and water table fluctuations; the latter can be tectonically as well as climatically controlled.

In summary, Kalahari *mekgacha* development can be explained by two, not necessarily mutually exclusive, sets of processes. These are that either fluvial and/or groundwater activity have been responsible for valley development, with the possibility of the different processes having varying spatial and temporal significance. The hypotheses for valley development also have varying significance in terms of the degree of hydroclimatic change they require in order to operate. Whilst the presence of perennial flow within valleys would have required either a shift in climatic parameters or some form of past drainage diversion, explanations of development by high-magnitude-low frequency flood events or groundwater processes would need less recourse to climatic change.

4.3 Introduction to methodology, and its significance in hypothesis evaluation

Three main approaches have been used to evaluate the hypotheses for Kalahari *mekgacha* development. These were the use of field investigations of valley form coupled with aerial photograph interpretation, studies of duricrusts exposed within valleys and an analysis of the drainage network structure of the major valley systems.

Detailed accounts of methodology and results from each of these approaches are given in chapters 5, 6 and 7. The remainder of this chapter briefly introduces the significance of each approach to hypothesis evaluation.

4.3.1 Field studies and remotely-sensed data

Field investigations have been employed primarily in order to assess variations in valley form and to identify any localised sedimentological deposits or structures indicative of former flow within valleys. Field study sites were selected on the basis of analyses of aerial photography and, in the case of specific valleys, upon locations cited in previous research. Air photographs were also used in the identification of larger features, such as terraces or relict channels, suggestive of the role of flowing water in valley development. Additionally, interpretations of Landsat imagery were used as part of an analysis of drainage network orientation.

The results of investigations of aerial photography are of most significance to the evaluation of hypotheses suggesting fluvial activity. In many cases, landforms are invisible on the ground due to the overall low relief of the Kalahari, and can only be identified from aerial photography. Field studies of valley form were used in order to identify valleys showing evidence of excessive incision, amphitheatre valley heads or abrupt widening and deepening, all of which are of significance if groundwater sapping has played a role in valley development.

4.3.2 Duricrust analysis

The role of studying duricrusts in order to evaluate *mekgacha* development is slightly more complex, and a full rationale and explanation of the various methodologies employed are given in chapter 6. In summary, duricrusts exposed in valley sides and floors have been studied at both meso- and micro-scales, utilising a number of field and laboratory techniques in order to assess relationships between the timing and environmental conditions associated with duricrust formation and diagenesis, and hence establish how these factors relate to valley formation.

One aspect of study was the need to identify the presence or absence of stratigraphy within duricrusts exposed in valley flanks, which would enable the relative timing of duricrust formation and valley development to be established. Duricrust suites could have existed prior to valley formation, having been formed by pedogenic processes, or be a product of surface or sub-surface water movements acting within a pre-existing valley. Lithological borehole records were used to determine spatial variability in duricrust types in the vicinity of valleys, and also to identify any evidence of deep-weathering. Additionally, areas of well-exposed duricrusts were studied in a series of profiles and in thin-section to determine any clear stratigraphic relations.

Thin-sections were also studied to assess the presence or absence of micro-morphological features indicative of formation within weathering profiles, as identified by Summerfield (1978, 1982). Samples were studied from areas where duricrusts were well-exposed, these same samples also being analysed by x-ray fluorescence techniques to enable quantification of sample chemistry. X-ray fluorescence also enables the evaluation of the concentrations of certain oxides (notably TiO_2), identified by Summerfield (1983*a*) as indicators of environmental conditions at the time of duricrust formation. Discriminant analysis was carried out on bulk chemistry data for silcrete samples in order to assess geographical variations in chemistry and also to compare composition with samples from Summerfield (1982, 1983*d*).

The significance of these techniques for *mekgacha* is discussed fully in chapter 6, but can be directly related to the two main hypotheses. A number of attempts have been made to define an overall stratigraphy for the Kalahari Group sediments, mainly by the use of lithostratigraphical correlation between duricrust exposures in various regions. Having assumed a stratigraphy exists, it is usually inferred that valleys have fluvially incised into a pre-existing sequence of rocks (e.g. R.J. Thomas *et al*, 1988). Clearly, establishing a stratigraphy for a terrestrial sedimentary sequence which covers over 2.5 million km^2 is not without complications. Whilst many of the older duricrusts at depth within the Kalahari Group can be lithostratigraphically correlated, it may be the case that comparatively recent exposures of duricrusts within *mekgacha* were formed primarily due to the presence of *mekgacha* themselves. This would suggest that valleys have not formed as a result of simple incision by fluvial processes, but that valley and duricrust development are at least partly contemporaneous, with a significant involvement of laterally moving water contributing to duricrust development.

The recognition of whether duricrust formation occurred under weathering or non-weathering profile conditions is of importance for establishing local environmental parameters. If duricrusts (particularly silcretes) can be shown to have developed in a weathering profile and there is evidence for deep-weathering beneath valley floors, then an interaction between groundwater activity and valley development can be broadly inferred. The possible use of silcrete as a palaeoclimatic indicator has been suggested by Summerfield (1983*a*,*b*, 1986) making analyses of the TiO_2 content of silcrete samples of additional interest, although this will be discussed more fully in chapter 6.

Finally, in order to ascertain the relative age of duricrust exposures, samples containing materials suitable for absolute dating were collected. These samples, mainly calcretes, containing freshwater molluscs (*Lymnaea natalensis* and *Melanoides tuberculata* species) and casts of algal growth on reed stems were collected from low terrace levels in the Okwa and Xaudum valleys and from the Okwa Gorge where the valley cuts the Gidikwe Ridge. Whilst there are problems of interpretation of the significance of such shell and organic deposits set in a calcrete matrix, the dates are interpreted in chapter 6 and by Shaw *et al* (1993).

4.3.3 Network orientation analysis

A macroscale analysis of the relationship between valley orientation and geological structures was also undertaken. In the account of geological structures within the Kalahari (section 3.2) it has been noted that fractured and faulted pre-Kalahari bedrock is often buried by Kalahari Group sediments up to 100 m thick. A detailed methodology is given in chapter 7, with statistical analysis being based upon the method outlined by Abdel-Rahman and Hay (1981), and the sampling strategy designed specifically for the study of *mekgacha*. Essentially, the study involves measurements of the angular difference between randomly selected valley segments and structural lineaments. Structural information was identified from a variety of sources including interpretation of Landsat imagery by Mallick *et al* (1981). From this analysis, sections of valleys exhibiting a strong structural control can be identified. These results were then related to the thickness of Kalahari Group sediments, to establish any link between degree of structural control and sediment thickness.

The significance of this technique is most apparent where valley sections showing strong geological control as a result of structures buried by Kalahari Group sediments are identified. Close parallelism between lineaments and valleys in areas of bedrock outcrop or subcrop can be simply attributed to structural control of surface drainage. However, where close parallelism occurs associated with fractures developed in bedrock buried by sediment and there is evidence for deep-weathering from lithological borehole logs, then the role of groundwater erosion in valley location and possibly formation can be inferred.

Having introduced the main techniques of study, the following three chapters give full methodologies and results for each of the three approaches. Interpretation of results is included within each chapter and subsequently discussed in Part 3.

Chapter 5
Variations in valley morphology from field studies and remotely-sensed data

5.1 Introduction

This chapter considers the results of field studies and the interpretation of aerial photography to assess variations in the form, both planemetric and morphometric, of Kalahari *mekgacha*. Information from previous studies specific to certain locations is also included where appropriate. Further results of field studies are also included in the following chapter, in conjunction with the mapping, description and analysis of duricrusts.

Not all Kalahari valley systems were studied in the field, the degree of depth of study dependent upon general accessibility and the presence of duricrust exposures. More detailed studies were undertaken in the Okwa, Mmone/Quoxo, Auob, Kuruman and Serorome valleys, whilst the Nossop, Moselebe, Ncamasere, Xaudum, Groot Laagte, Hanchai and Rooibrak/Passarge valleys were studied at a reconnaissance level only. The Molopo Valley satisfies both access and duricrust exposure criteria, yet was also studied at a reconnaissance level, having already been the subject of a thesis by Smit (1977). This chapter systematically considers field study information, with valleys subdivided according to whether their courses are directed endoreically or exoreically; i.e. whether they trend towards the Makgadikgadi Depression or Okavango Delta or ultimately connect with the Orange River via the Molopo Valley (figure 1.1). The most detailed studies were carried out in the endoreic valley systems. Information regarding variations in valley form and depth of incision is largely descriptive due to the absence of topographic survey data or maps for the majority of the Kalahari within Botswana. Figures given for widths and depths are often best estimates, mainly due to the high width to depth ratios and low relative relief of most valleys. Where a valley was sufficiently narrow to allow cross-sectional profiling, this information is included.

As noted in the introduction to this thesis, a number of regional words meaning "valley" are often attached to the name of a particular valley, particularly in northeastern Namibia and northwestern Botswana (McConnell, 1956). The nomenclature used in chapter 1 is adhered to throughout this chapter. When discussing valley systems in general, they are referred to both as the SeTswana term *mekgacha* (singular *mokgacha*) and as "valleys".

5.2 Endoreic drainage systems

In this section, the Okwa/Hanchai, Mmone/Quoxo and central Kalahari valleys (Deception and Rooibrak/Passarge) which drain towards the Makgadikgadi Depression, and the northern valley systems (Rukange, Ncamasere, Tamacha, Xaudum, Qangwadum, Eiseb, Epukiro and Groot Laagte in northeast

Namibia and northwest Botswana) which drain towards the Okavango Delta region are considered (figure 1.1).

The endoreic systems have been the subject of the two previous general accounts of *mekgacha* morphology (Boocock and Van Straten, 1962; Thomas and Shaw, 1991*a*). Three different morphological stages have been identified; headwater regions (away from bedrock outcrops) are typically flat, with a morphology similar to the dambos discussed in chapter 2. This gentle form gives way abruptly to an incised gorge-like section with steep sides and a flat valley floor, with the valley eventually dwindling to a further broad and flat stage and often becoming completely sand-choked.

5.2.1 Okwa and Hanehai valleys

(a) Previous studies

The Okwa and Hanehai valleys form the longest of the Kalahari valley systems, with the Okwa over 600 km long (figure 5.1). Both systems rise in Namibia, the Okwa having its headwaters on Damara Sequence rocks which form a westward extension of the Ghanzi Ridge to the southeast of Gobabis. Some confusion regarding the names of both valleys occur where they cross the international boundary. The Okwa is indicated on Namibian topographic maps as the Chapman's River, whilst the Hanehai is variously referred to as the Otjimbindwe (a Damara name meaning "blood"; Schwarz, 1920), Buitsivango or Rietfontein Valley.

As with most other Kalahari valley systems, the Okwa and Hanehai mainly appear in the literature in association with non-geomorphological studies. Possibly the earliest reference to either valley is by Andersson (1856 p. 374), who crossed the Omuramba Otjimbindwe (the Hanehai) during his travels, describing it as a sandy and dry watercourse. Baines (1864 p. 119) also crossed the the Hanehai, noting that;

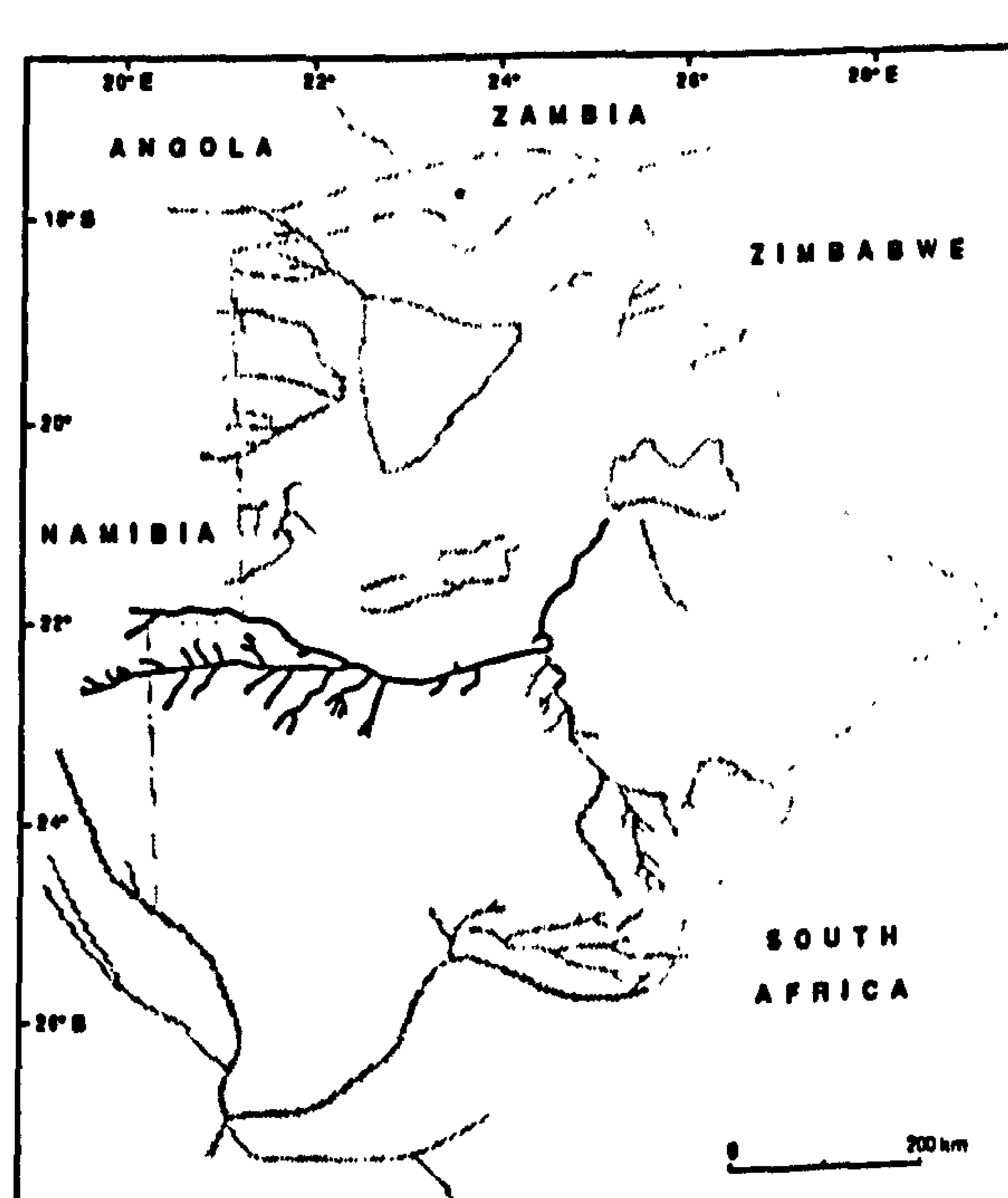


Figure 5.1: Location of the Okwa and Hanehai valley systems.

> "it's breadth was from 100 to 150 yards, with low banks and ridges of sandstone here and there: and the grass in it was as dry, white and feathery as if water had never flowed there, and never could."

Hodson (1912 p.96), whilst travelling to Kalkfontein in the Ghanzi District of Botswana;

> "...passed the Okwe River [sic], a river only in name as it was quite dry".

Subsequent references have been made as part of the consideration of former drainage configurations in southern Africa. Rogers (1934, 1936) dismissed postulated links between the Okavango and Limpopo Rivers, and also Schwarz's (1920) possible link between the major northern rivers and the Orange River, on the basis of differences in freshwater shell populations. The concept of a Trans-Kalahari river was based upon a description by Schwarz (1920), who envisaged linkages between sections of the Epukiro, Hanchai and Okwa valleys (of which only the latter two are actually connected) flowing towards the Makgadikgadi Depression and then heading southwards to join the Molopo River by means of the Mmone and Moselebe valley systems. The findings of Du Toit (1926*a*) indicate that such a connection could not exist today, mainly due to neotectonic activity having affected the elevation of drainage courses. The notion of a Trans-Tswana River persists in the literature (e.g. Main, 1987; Moore, 1988) primarily due to the discovery of extensive conglomeratic deposits in the vicinity of the Orange River in Griqualand, believed to have been derived from a northerly source (McCarthy, 1983).

The majority of references to the Okwa and Hanchai valleys have been made in the context of geological and Quaternary geomorphological studies. Geological work has mainly considered pre-Kalahari Group formations with little reference to valley geomorphology (e.g. Aldiss and Carney, 1992). However, Crockett and Jennings (1962, 1964, 1965), Jennings and Crockett (1961), Lawrance and Toole (1984), Litherland (1982), Union Carbide (1980*c* and *d*) and Aldiss (1987*b*, 1988) do include some interesting observations. Litherland (1982) considers the geology of parts of the western Okwa Valley close to the Namibia/Botswana border, noting that the minor headwaters of the Okwa draining the Mamuno escarpment are generally controlled by bedrock jointing systems. The study also notes the occurrence of riverine calcrete deposits up to 10 m thick containing bivalves up to 7 cm long. Flow within the valley generally ceases within 7 km east of the village of Makunda i.e. 18 km east of the Namibia/Botswana border. Lawrance and Toole (1984) include reference to buried headwater sections of southern tributaries to the Okwa in the vicinity of Takatswaane (22°41'S, 21°55'E) and Lone Tree (22°57'S, 22°07'E) villages. They note a tendency for the valleys to dwindle away to a line of pans in their headwater areas.

The remaining references concern the Okwa in the vicinity of Tswaane Veterinary borehole (22°21'40S, 21°50'30"E) where the Jwaneng-Ghanzi road crosses the valley and a major inlier of Precambrian bedrock occurs e.g. Boocock and Van Straten (1962). The studies by Crockett and Jennings (1962, 1964, 1965) suggest a complex history of development, noting two main terrace levels, gravel deposits and distinct valley asymmetry in the area of Tswaane borehole. Crockett and Jennings (1964) propose that the present landscape was attained prior to the most recent incision of the Okwa. These authors also suggest that the Okwa is structurally controlled and may have formed along the line of a proto-valley, possibly having existed since Cretaceous times. Structural control of a particularly straight southern tributary to the Okwa, which joins the main valley to the east of Xade village, is also indicated by Hutchins *et al.* (1979). They propose that the straightness of the valley is due to control by a NE-trending fault associated with the Makgadikgadi Line structural break.

Boreholes drilled by Union Carbide (1980*c* and *d*) near the Hanchai/Okwa confluence reveal lenticular development of calcrete in valley flanks, forming calcrete supported terraces and shoulders. The presence of these remnant shoulders is cited as possible evidence for a wetter period when calcrete deposits were washed away. Aldiss (1984, 1988) notes that silcrete and calcrete is found penetrating basement rocks in the vicinity of Tswaane borehole.

Remaining references to the Okwa Valley have been made in an anthropological and archaeological context (e.g. Aldiss, 1987*a*) and also with regard to the relationship between the valley and the former Lake Makgadikgadi. The sand-choked course of the Okwa within the Central Kalahari Game Reserve is briefly mentioned during studies of San Bushmen by Tanaka (1976) and Silberbauer (1981). Tanaka (1976 p.100) notes no permanent water sources within "...the long depression that is the Okwa Valley". The presence of terrace levels up to 30 m above the present channel bed, diatomaceous earth and calcrete within valley floors and large sinkholes in the floors of the Xade and G/edon!u tributary valleys are noted by Silberbauer (1981).

Finally, the work on the Quaternary geomorphology of the Makgadikgadi Depression considers the Okwa Valley where it meets the shoreline of the former lake at the Gidikwe Ridge. Cooke and Verstappen (1984) provide the most detailed description of the Okwa in this area, mainly via the interpretation of Landsat imagery and aerial photography. They note an area of shallow lagoonal sediments to the west of the Gidikwe Ridge probably caused by ponding of the Okwa, and an area of thin deltaic deposits within the ridge where the Okwa drained into the 920 and 912 m lake levels. The relationship between the 920 m lake level and former flow in the Okwa is also noted by Breyer (1982). Results included in Shaw, Thomas and Nash (1993) will be described below.

(b) Field studies and aerial photography

The Okwa Valley was the subject of extensive field investigations during both 1989 and 1990 field seasons, with the Hanchai studied at a reconnaissance level during 1990 only. Both valleys are comparatively easily accessible, particularly the Okwa in the relatively well-populated section between the Namibia/Botswana border and the Jwaneng-Ghanzi road (figure 5.2). Further east, study within remote parts of the Central Kalahari Game Reserve was not attempted.

(i) The Okwa Valley

The Okwa Valley rises at a height of 1470 m asl at 22°40'S 19°13'E, 35 km southeast of Gobabis in Namibia where it is known as Chapman's River. The headwaters comprise a number of small tributaries draining the Namibian extension of the Mamuno Escarpment, an ENE-WSW trending relief feature over 140 m high consisting predominantly of Late Precambrian meta-arkoses (Litherland, 1982).

Figure 5.2: Locations along the Okwa and Hanchai mentioned in section 5.2.1.

In total over 88 km of the valley course lie to the west of the Namibia/Botswana border, the Okwa ultimately terminating some 600 km from its source area at the Gidikwe Ridge near the Makgadikgadi Depression. Field studies were conducted over approximately half of the valley length, including 25 km of the valley within Namibia, the entire course from the international boundary to the confluence with the Hanehai, and the Okwa Gorge area where the valley cuts the Gidikwe Ridge. The results of these studies will be systematically discussed for locations in a west to east direction.

The Okwa within Namibia

A number of minor tributary valleys to the Okwa were crossed on the Gobabis-Mamuno road, each being gullies cut through bedrock draining off the escarpment. A calcrete exposure occurred on the east side of the tributary to the east of Grauwater Farm, but otherwise no major occurrences of duricrust were noted in association with these tributary valleys.

The main Okwa Valley was first studied at Molly's Farm (figure 5.2) where the channel consists of a 12 m wide trough, approximately 1 m deep. The channel was sand-filled, but shallow bedrock was also seen outcropping in places. At Karolinenhof Farm, 13 km to the east, the valley has a channel 25 m wide and 1.5 m deep with clear evidence of flow (plate 5.1). On the photo, the channel bed contains almost white sand, in contrast with the channel flanks which are a buff sandy colour and an organic-rich terrace surface.

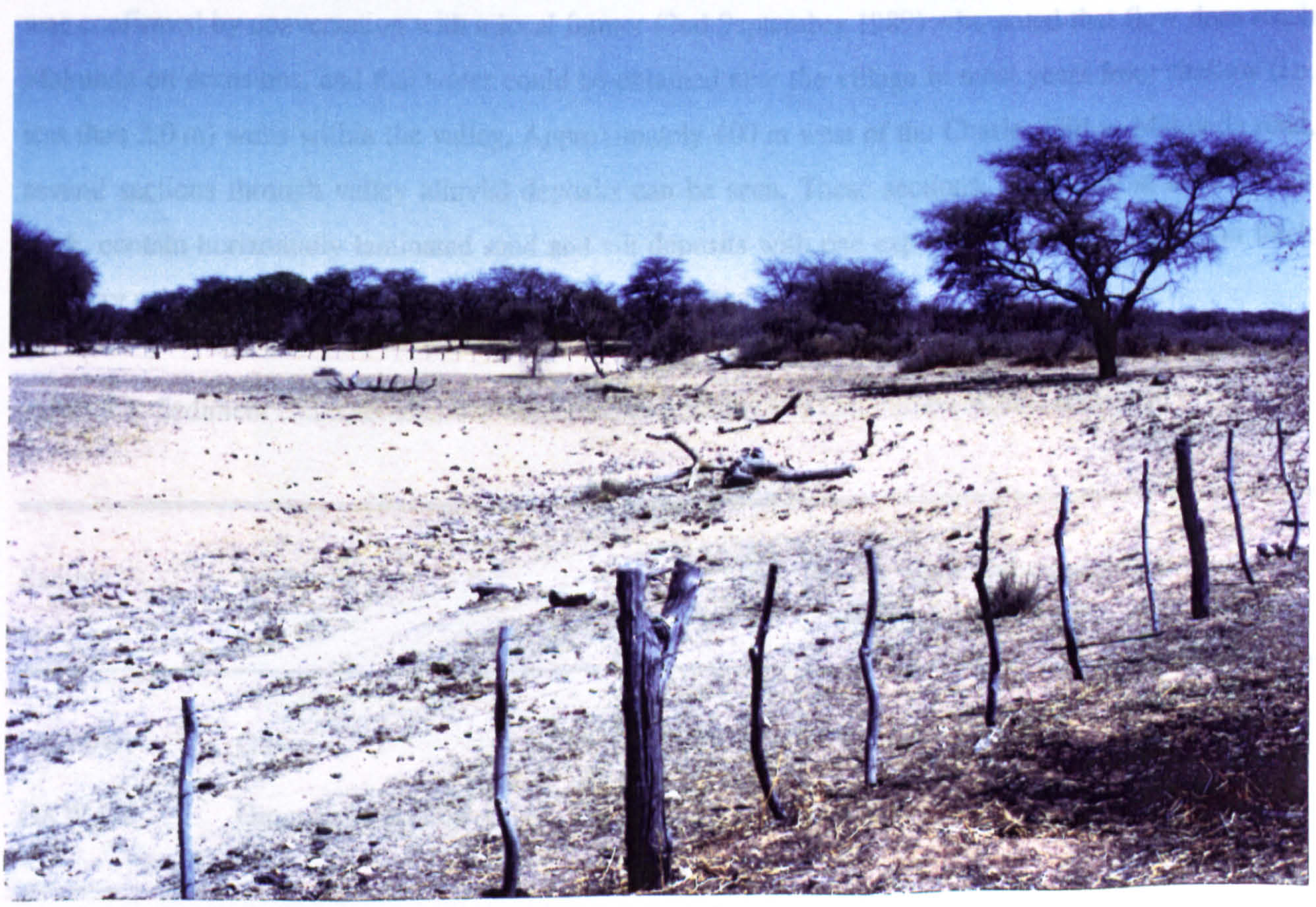


Plate 5.1: The Okwa Valley at Karolinenhof.

Conversation with Mr. Jürgen Eichhoren, owner of Karolinenhof Farm (1st September 1989), gave the following information on recent flows within the Okwa within Namibia. There is usually flow at Karolinenhof at least once a year, with seven periods of flow in the summer of 1985-1986. The magnitude of flow events in the Okwa depends upon the number of headwater tributaries contributing water to the main channel; the highly variable temporal and spatial distribution of rainfall may mean that one or more tributaries contain flow after a rainfall event, but this may not reach the Okwa. The largest floods occur when a number of tributaries contain flow simultaneously, as was the case in 1976 and 1978 when the flood event was enhanced by the bursting of small earth banked dams up-valley of Karolinenhof. Mr. Eichhoren indicated the area of his farm which was flooded during 1978, which suggests that the flood channel was 75-100 m wide on average, attaining a maximum width of nearly 300 m. The depth of flow associated with such surface water widths, based upon the estimated bank-full depth, would be of the order 1.5 to 1.75 m, in comparison with the present 1 m deep channel. This flood also caused a diversion of the Okwa channel which formerly meandered within the farm boundary, but cut a new straight section which subsequent flows have since followed.

The Namibia/Botswana border to Gobololo borehole

Where the Okwa enters Botswana it has a similar form, with a 28 m wide watercourse containing an 11 m wide channel bed 4 km west of Makunda. Valley slopes are less than 1°, and are typically composed of a mixture of alluvial material and Kalahari Sand. The presence of large *Acacia erioloba* and *Boscia albitrunca* trees near the valley indicates groundwater availability beneath the valley floor. This suggestion was confirmed by conversation with a local farmer (2nd September 1989) who noted that flow does reach Makunda on occasions, and that water could be obtained near the village in most years from shallow (i.e. less than 2.0 m) wells within the valley. Approximately 400 m west of the Charles Hill to Makunda road, several sections through valley alluvial deposits can be seen. These sections, which are at most 50 cm thick, contain horizontally laminated sand and silt deposits with one exposure containing an 11 cm thick powdery calcrete layer with calcrete nodules 4 cm above the profile base.

Table 5.1: Sediment characteristics for western Okwa Valley and Omurambwakahue tributary.

Sample	Valley	Folk & Ward statistics (phi)			
		Mean	Sort.	Skew.	Kurt.
OKWA/1	Okwa	2.06	1.11	0.04	1.57
OKWA/2	Omurambwakahue	2.34	0.69	0.12	1.14

A major change in valley form occurs at Makunda, with the Okwa broadening and deepening to a valley 800 m wide and 3-4 m deep. Hardpan calcrete is exposed over much of the valley floor within the village, with the valley flanks and parts of the floodplain area covered by red to buff Kalahari Sand. An intermittent watercourse is still present, reaching a maximum depth of 60 cm, which dwindles to a series of small grassy pans to the east of Makunda.

A contrast in flow regimes is apparent between the main Okwa Valley and its tributaries in this region, distinguished by variations in valley bed sediments. This contrast is most clearly seen 4.5 km east of Makunda where the road to New Xanagas cuts west-east across the confluence of the Okwa and the Omurambwakahue tributary. Two distinct channels are present, with different sediment characteristics reflecting the variations in the persistence of water in each valley (table 5.1). The bed of the Omurambwakahue is dominated by well-sorted sand with the Okwa containing a less well-sorted grey organic- and clay-rich vertisolic soil which contains some clay pellets (up to 2 mm diameter). The Okwa channel is also more deeply incised, supporting the sedimentary indications for more regular flow within the main valley.

Plate 5.2: The confluence of the Okwa and Onjonja valleys at Jerusalem borehole (from aerial photography; scale 1:50,000). Botswana Department of Surveys and Lands photo number Ghanzi 19 (132), dated 7 October 1986.

The next tributary to the east, the Kanai or Otjizumbido, is 600 m wide with a depth of incision of approximately 9 m. The flat valley floor contains no evidence for recent flow, with the valley axis consisting of a 40 m wide zone of fine calcrete rubble. A typical feature at confluences of the Okwa and its tributaries is that valley flank slope angles decrease in both valleys, often associated with broadening of the valley, with maximum slopes of 2° and 3° seen where the Kanai and Onjonja valleys join the main *mekgacha*. These gentler slopes may be associated with locally developed interfluves, and are most typically seen at confluences. There is no evidence for flow within any of the valleys, although outcropping hardpan calcrete at the Okwa-Onjonja confluence (at Jerusalem borehole) contained a localised channel.

Around Jerusalem borehole low paired terraces first appear at heights of between 1.8 and 2.2 m above the valley floor. The southern terrace can be identified on plate 5.2 from the shadow on the southern flank. The typical slope angle from the valley floor to the terrace surface is 4 to 5°, with typical slope lengths of around 30 m. The terraces at Jerusalem are calcrete-supported, with near vertical low cliffs seen in places. The calcrete has developed in valley floor host sediments as opposed to regional bedrock. Outcrops of meta-arkose occur at 2.5 km east of Jerusalem, and show no evidence of calcretisation.

The terrace level is traceable along the Okwa between Jerusalem and Nwaxke at a height of around 2.0 m above the valley floor, being most extensively developed on the inside of bends and in wider valley sections, with a maximum surface distance from the valley axis to the Kalahari Sand flank of 400m. The terrace level suggests that there has been infill of the main valley, with subsequent incision creating a 30 m wide "channel" through the calcretised sediments. Other areas of outcropping bedrock occur along this section of valley, many being immediately overlain by, and possibly supporting, calcretised terrace sediments. Around 1.3 km west of Nwaxke, exposures of well developed calcrete occur, with an 9 m long 12.5° slope of calcrete exposed from the valley floor to the terrace surface. The valley at this point is over 500 m wide, with a total depth of incision of around 15 m, a 22 m wide channel bed between the 2.0 m terraces with gentle convex slopes along the terrace surface to the valley flanks.

The terrace level is less distinct between Nwaxke and New Xanagas, although it is intermittently seen at a height of around 2.0 m. Slope angles decline in this area, with the terrace barely distinguishable in places from the Kalahari Sand valley flanks. The terrace is not seen to the east of New Xanagas, the valley form consisting of a broad shallow depression with a total depth of 16-18 m.

At Nxogodimo borehole there is evidence of a localised shift in channel position of the Okwa. The borehole pump-house is sited on a large low calcretised sandy mound in the centre of the valley with channels to north and south, the main valley being deeper to the south of the borehole. The gentler slopes and lesser degree of calcretisation to the north of the mound suggests that the main channel was once on the northern side of the valley but has subsequently shifted and developed more deeply to the south.

East of Nxogodimo the Okwa Valley narrows considerably to less than 200 m wide in places, and a low terrace level is again evident at around 2 m. This terrace merges into the valley flanks within 4.5 km east of Nxogodimo, but reappears in wider valley sections near Gobololo borehole. At Gobololo the terrace is best developed on the south side of the valley, with the lower 1.2 m consisting of solid calcrete passing upwards into nodular calcretised sediment. The calcrete contains shell material, mostly intact specimens of *Lymnaea natalensis*. Additionally, well-preserved bivalve molluscs were noted in hardpan calcrete in a spoil pit 4.5 km east of Gobololo. The presence of intact and also some comminuted, well-developed mollusca suggests periods of relatively long-standing water within the Okwa Valley.

The Okwa in the vicinity of the Ghanzi-Jwaneng road

The Okwa between 40 km west of the Ghanzi-Jwaneng road (approximately 22°18'S 21°30'E) and its confluence with the Takatswaane Valley (22°24'S 22°07'E) was investigated in 1989 and 1990.

In the westernmost part of this study section, the Okwa has a similar form to that previously described, the valley width and depth varying between 300-500 m and 15-18 m respectively. Low paired terraces are continuously present at a height of 2.0 m above the valley base, the base varying between 12-20 m in width. The terraces are calcrete-supported, although good exposures of calcrete are comparatively rare. A 250 m long exposure of calcrete occurs on the northern bank of the Okwa around 1.2 km east of borehole 3974 (figure 5.2). This 0.85 m outcrop is approximately 0.75 m above the valley floor, and represents the middle to upper section of the terrace. The highly porous hardpan calcrete has developed in an alluvial host material, with a comparatively low sand-size component, and has a pitted surface with a weathering rind up to 6 cm thick. The exposure has a high shell content dominated by mature and immature specimens of *Lymnaea natalensis*, indicative of still near-permanent water (Brown, 1980). A radiocarbon date of 11,890 ± 60 years BP was obtained from a combined shell and calcrete sample taken from near the top of the exposure (GrN 17010), the shell component being insufficiently large to allow separation from the carbonate matrix (Shaw, Thomas and Nash, 1993). This date requires careful interpretation, and can only be regarded as a minimum age for the shell material, but is largely consistent with other dates from the Kalahari. Further interpretation is provided later in this section and in Shaw, Thomas and Nash (1993).

To the east of borehole 3974 the Okwa swings through a series of right angled meanders. The terrace level is identifiable at 18 km and from 8-11 and 3-6 km west of the Ghanzi-Jwaneng road. The terrace has a lobate form, and is extensively dissected by many gullies which disrupt the upper terrace surface. Slope angles vary between 1° and 2.5° only, with slopes generally covered by Kalahari Sand. Slopes in the 11 km immediately west of the main road are distinctly asymmetrical, the south flank being steeper than the north.

Bedrock outcrops between 3 km south of borehole 3974 and 3 km west of the main road, indicating that any sedimentary fill within the valley is thin in this region. Rubbly calcrete exposures are often seen immediately overlying bedrock (e.g. hardpan calcrete overlying mica-schist in a well 7 km west of the road; Don Aldiss, pers. comm., 13th December 1990), but outcrops are not calcretised. Shell material is incorporated within calcrete in a number of locations (e.g. in the small pan 12 km west of the road). The two tributary valleys, 18 and 13 km west of the main road, have a similar overall form, with banks of calcrete rubble up to 1.5 m high located 1.5 to 2.0 m above the valley floor.

Precambrian bedrock outcrops extensively for 10 km east of the Ghanzi-Jwaneng road, with the Okwa showing a different form in this region (plate 5.3). The 2.0 m terrace is present immediately east of the road near Tswaane borehole, and from 1 km west of Old Tswaane borehole, but disappears within the section of continuously outcropping bedrock. The terrace can also be seen in the tributary valley which joins the Okwa 2 km east of the main road. The Okwa Valley floor widens up to 65 m in places, with north and south flanks showing a more pronounced asymmetry than seen further west (plate 5.4). Where rock outcrops are present, the southern valley flanks are almost vertical in places, but more commonly have 10° to 15° slopes which are noticeably steeper than the sandy 6° to 8° northern slopes.

In this bedrock section, the valley floor is flat with an abrupt break of slope at the valley flank. Both flanks are less steep between 3 and 8 km east of the main road, the valley floor widening to between 200-300 m. The southern flank steepens again where the valley swings abruptly to the southeast, and consists of a 5 m bedrock cliff. The structural control of this section of the Okwa has been noted by a number of authors (e.g. Aldiss, 1988) and the valley consists of fault-guided straight sections with abrupt changes of orientation at the intersection of fault zones.

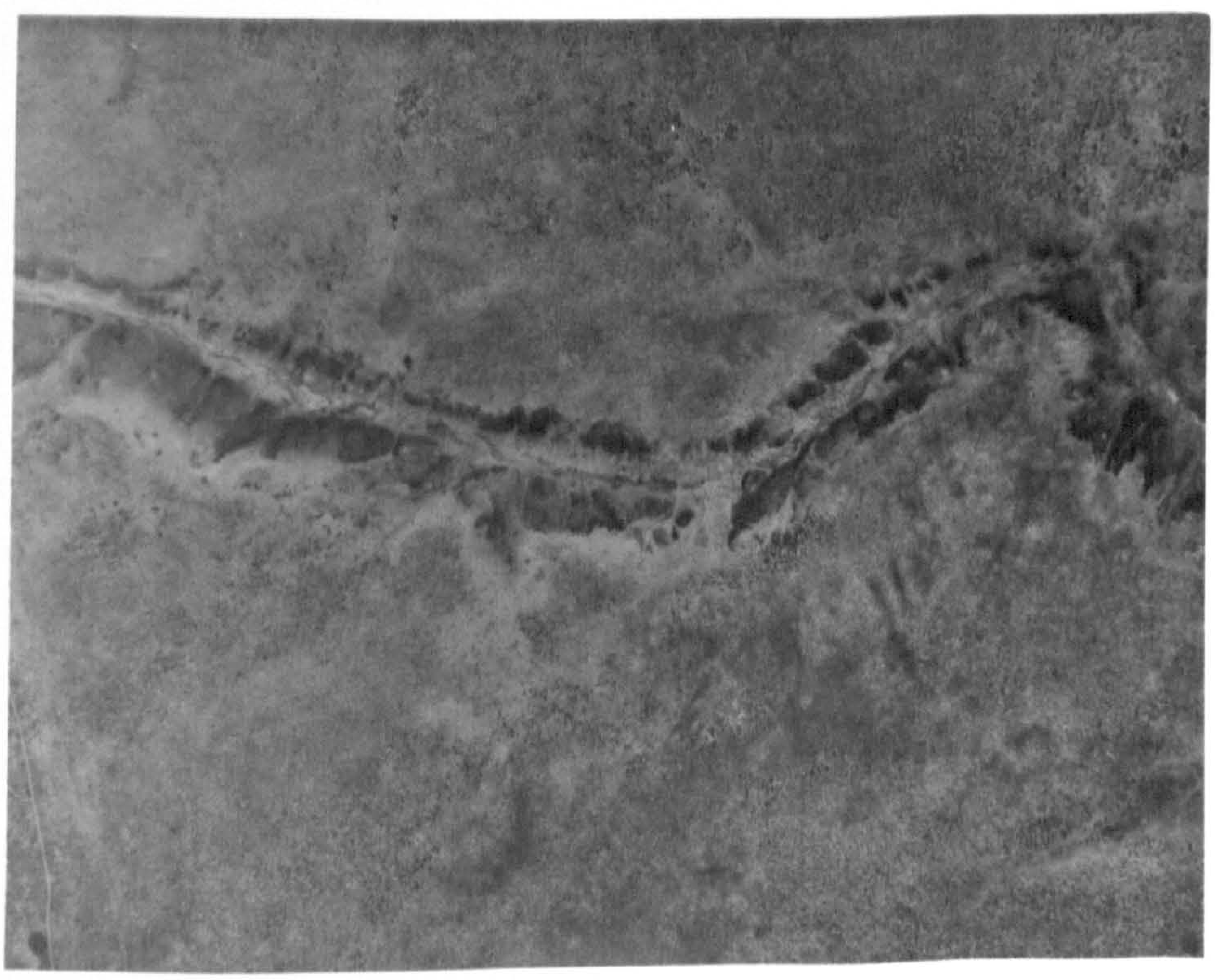


Plate 5.3: The structurally controlled section of the Okwa Valley immediately east of the Ghanzi-Jwaneng road (from aerial photography; scale 1:50,000). Botswana Department of Surveys and Lands photo number Ghanzi 19 (096) dated 7 October 1986.

There is also localised evidence for fluvial erosion within the valley (Don Aldiss, pers. comm., 13th December 1990) from an area of lag deposits and apparently water-smoothed bedrock outcrops (Nash *et al.*, 1993). The lag deposits, consisting of well-rounded gravel clasts up to 40 cm diameter, are exposed in two pits on the south bank of the valley on the inside of the abrupt bend approximately 8.5 km east of Tswaane borehole. The water smoothed outcrops are located on the inner bend of the valley at 5 km east of Tswaane. The presence of such features combined with terrace levels and a flat valley floor consisting of sedimentary infill, suggests that this section of the Okwa was mainly formed by fluvial activity.

The pattern of terrace features becomes locally complicated to the east of Old Tswaane borehole, with a terrace consistently at 3.5 to 4.0 m above the valley floor and the intermittent occurrence of a 2 m deep, 8-10 m wide trough along the valley axis. This latter trough appears to be a continuation of the 2.0 m level seen further east, with the 3.5 m terrace not previously seen. This upper terrace is identifiable until 16 km east of Old Tswaane borehole, beyond which it merges into the valley flanks, which become increasingly gentle to the east (with average slopes of 7 to 8°).

Whilst duricrust rubble is common within the bedrock section, actual outcrops are comparatively rare and where present are usually complex suites of silcretes, calcretes and intermediate duricrust types. Silcrete outcrops occur around 1.0 to 1.2 km east of Tswaane borehole, with the most easterly of these exposures apparently consisting of altered granite. A major silcrete outcrop occurs on the southern valley flank at 4 km east of Old Tswaane borehole, including a 3 m duricrust cliff at the top of a 14 m high 24-28° debris slope. The cliff is dominated by highly indurated terrazzo-silcrete with a chert-like silcrete present within spring lines in recesses in the valley flank. The debris slope is underlain by soft powdery calcrete. This section of the valley shows the most marked asymmetry seen along the Okwa, with southern flanks rising abruptly to a height of 17 m whilst the northern slopes are sand covered and average 2° to 5°. Calcrete outcrops *per se* are uncommon except in lower slopes, and only appear around 1 km west of Old Tswaane borehole, becoming commonplace with the reappearance of terrace levels a further 5 km east.

The marked asymmetry, occurrence of two terrace levels and variations in form of the Okwa in this section require explanation. The asymmetry between north and south valley flanks may be explicable by aeolian infill from north- or northeasterly sand sources. The northern flanks are less steep and contain fewer bedrock outcrops than their southern counterparts, which may be due to a progressive burial of the slopes by Kalahari Sand. In the area 8.5 km east of Tswaane borehole where the valley swings from a northeasterly to a southeasterly orientation, the inner bend is composed of a cliff of bedrock whilst the outer has an atypical gentle sand slope.

Variations in valley form are more complex, but almost certainly relate to the presence of less easily eroded Precambrian bedrock in this area. To the west of the Ghanzi-Jwaneng road the valley form is relatively subdued, but becomes narrower and more box-like where the Okwa has incised through bedrock. This narrowing may have caused increased flow velocity and enhanced erosive potential in this area and to the east, creating the localised lower terrace to the east of Old Tswaane borehole.

The Precambrian bedrock in this area is an inlier surrounded by Kalahari Group sediments, with the valley cutting through both geologies. This would suggest that the Okwa has developed after the deposition of the Kalahari Group, and has incised through to underlying bedrock. The location of the valley within the bedrock inlier is structurally controlled, and this suggests that the valley has followed a line of weakness as it has incised. It would seem unlikely that the present course of the Okwa follows an exhumed valley since the bedrock inlier would have formed a topographic high prior to the deposition of the Kalahari Group (unless neotectonic activity has caused local folding and uplift). Additionally, the present orientation of the Okwa follows the regional gradient towards the local base level provided by the Makgadikgadi Depression.

Crockett and Jennings (1964) hinted at the complex evolutionary history of the Okwa in this section, where it has incised through bedrock and Kalahari Group sediments, has an alluvial infill with terraces and also contains aeolian infill. The two distinct groups of duricrusts present suggest evolution over long time periods, the valley flanks being underlain by highly indurated older silcretes which only outcrop occasionally, and more recent calcretes and sil-calcretes containing mollusca cementing the terrace materials. The deposits suggest that the Okwa was incised into the Kalahari sediments, then underwent a period (or periods) of infill by sandy alluvial material which was subsequently cemented by calcium carbonate and later dissected, this whole suite of landforms being subsequently covered by aeolian sediments.

The Okwa was not investigated further east except in the Okwa Gorge area, with studies within the Central Kalahari Game Reserve based on aerial photography. The form of the valley as far as the Hanehai confluence appears similar to the sections seen east of Old Tswaane borehole (plate 5.5), with a well-defined course visible on air photos. However, east of the confluence the Okwa becomes much less distinct, with an abrupt change at the confluence. The course of the Hanehai up-valley of the confluence also appears poorly defined and sandy. Personal communication with Don Aldiss (13th December 1990) indicates that the Okwa Valley floor is composed of firm grey slightly sandy soil, becoming noticeably sandier after the Hanehai confluence and eastwards into the Central Kalahari Game Reserve. This increase in sand content could have three explanations. Firstly, it is possible that sediment transport along the Hanehai has carried large quantities of sand into the Okwa. Secondly it may be that the Okwa has contained flow more recently than the Hanehai and has cleared much of the sand from its channel as a result, but the quantities of sediment in the confluence area were too great for further transportation. This may relate to flash-flooding, with sediment deposition occurring at the confluence. Thirdly, and more probable in view of regional patterns of aeolian sand transportation, the Hanehai Valley and Hanehai Hills may act as a trap and/or barrier for sediment transported by northeasterly winds, thus preventing sand from choking the Okwa further to the west of the confluence. This has implications for the explanations of valley asymmetry (by a process of aeolian sand blanketing) suggested earlier, but the two mechanisms are not mutually exclusive since a diminished supply of aeolian sediment could be transported across a topographic barrier such as a valley, and local redistribution of sediment is also possible.

Plate 5.4: Asymmetry in the Okwa Valley 3 km east of Old Tswaane borehole, looking east.

Plate 5.5: The Okwa Valley 10 km west of the Hanehai confluence, looking west.

The Okwa Gorge through the Gidikwe Ridge

The easternmost section of the Okwa and the gorge it cuts through the Gidikwe Ridge were investigated during the 1990 field season (figure 5.3). Immediately west of the Gidikwe Ridge the Okwa swings through a series of meanders (plate 5.6), possibly where it was ponded behind the ridge. The valley in this area is a 200-250 m wide sand-filled depression with a total depth of around 6 m (plate 5.7). The outer banks of meander bends are noticeably steeper than inner banks, which would not be expected if the valley were simply infilled by windblown Kalahari Sand.

The Okwa Gorge, whilst only around 10-12 m deep, is a distinctive feature on aerial photography (plate 5.6). It forms the effective termination of the Okwa where it debouched into the Makgadikgadi Basin, with only a faint trace of the valley visible on aerial photography to the east of the ridge. This may suggest only limited flow in the post-lake period, or may be evidence of partial sedimentary infill of a former channel. The gorge is between 100 and 150 m wide with a very flat floor, particularly at its eastern end where Cooke and Verstappen (1984) identify an area of deltaic deposits. The minimum extent of these deltaic sediments identifiable from aerial photography is approximately 350 km^2. In order to assess the age of sediments in this region three samples of shell and calcareous material (collected by P.A. Shaw) were radiometrically dated, from sites within the gorge and on the eastern side of the Gidikwe Ridge, 1 km south of the gorge mouth (figure 5.3). The results of these analyses are included in Shaw, Thomas and Nash (1993). Two separate sites were sampled, both from 80 cm depth in bioturbated sediments and both including specimens of the gastropod *Melanoides tuberculata*, the bivalve *Corbicula africana* and isolated *Bulinus* and *Lymnaea* spp. Radiocarbon dates for the mollusca in the Okwa Gorge (sample GrN 14787) and Gidikwe Ridge (GrN 14786) were 14,490 ± 150 years and 14,070 ± 150 years BP respectively.

Both sites also contained *in situ* calcified algal reed casts up to 10 cm long and 3 cm in diameter, indicative of shallow water environments probably at a lake periphery. A sample of these reed casts from the Gidikwe Ridge site yielded a radiocarbon date of 11,980 ± 60 years BP (GrN 15536). This date has been used in conjunction with dates from other sites around the former Lake Palaeo-Makgadikgadi in an empirical study by Shaw and Thomas (1992). Sets of paired dates (shell and carbonate matrix) were obtained for samples from the 936 m shoreline of the Lake Thamalakane stage in order to date the relative timing within samples for the formation of organic and inorganic carbonates. When dating carbonate material there is a risk of contamination between older and younger carbonate, with any date from a combined sample possibly biased toward the age of formation of the youngest carbonate (in the case of a calcrete sample), particularly where matrix material dominates the sample. The paired dates show a consistent variation of between 15 and 25% in the ages of the "real-time" indicator and the matrix, suggesting a later stage of carbonate development (Shaw and Thomas, 1992). This empirical study is of importance when interpreting the date of 11,890 ± 60 years BP for a combined shell-calcrete sample from the Tswaane area (GrN 17010, described above). This date probably represents the age of the younger calcrete matrix, which (assuming a 15-25% age difference) suggests a "real-time" age of around 13,600 to 14,800 years for the shell material, consistent with the dates from the Okwa Gorge area.

Plate 5.6: The Okwa Valley where it breaches the Gidikwe Ridge to form the Okwa Gorge (from aerial photography; scale 1:50,000). Botswana Department of Surveys and Lands photo number E.K.B."A" Strip 6 (063), dated 2 October 1985.

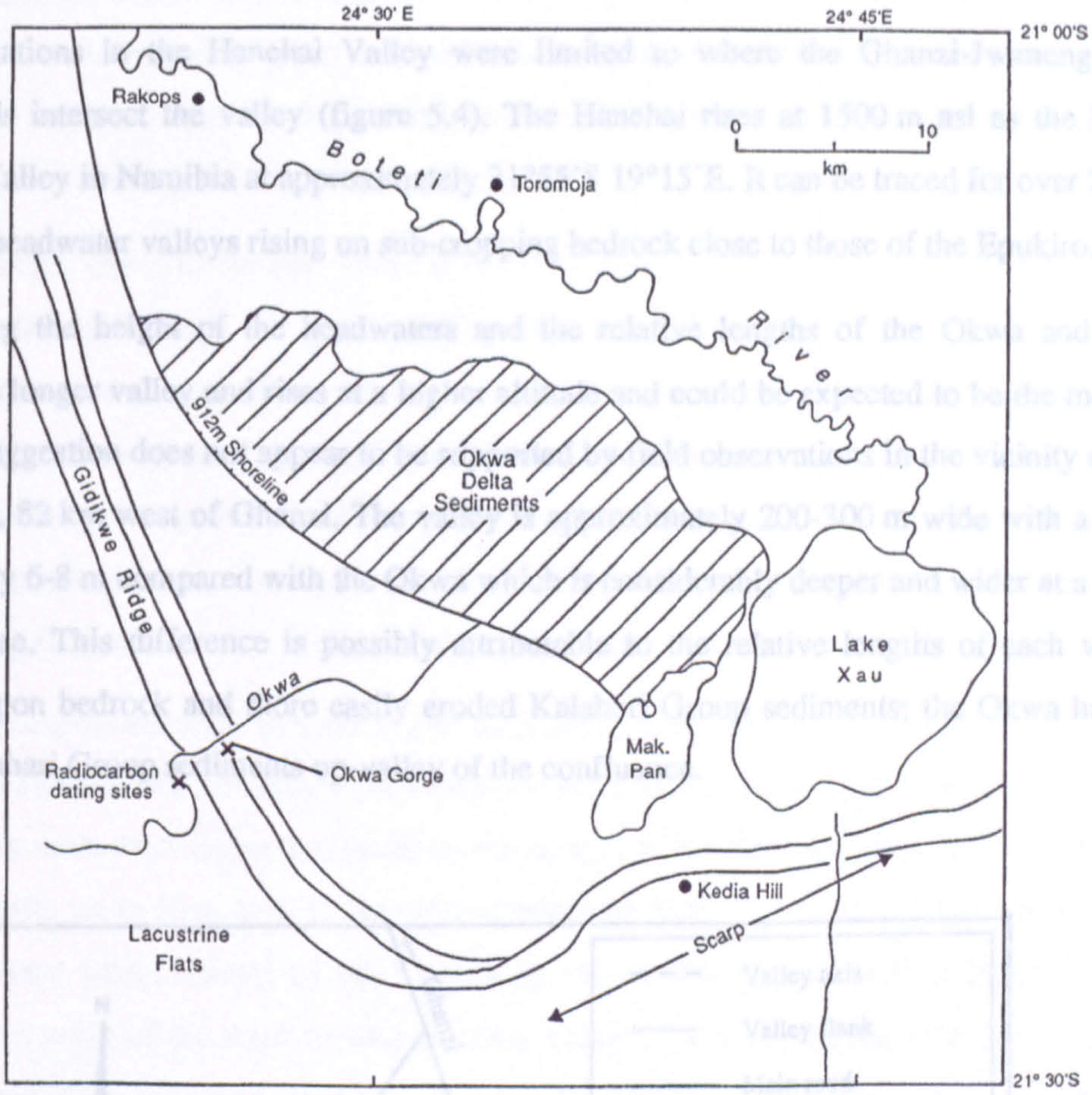


Figure 5.3: The Okwa Gorge through the Gidikwe Ridge.

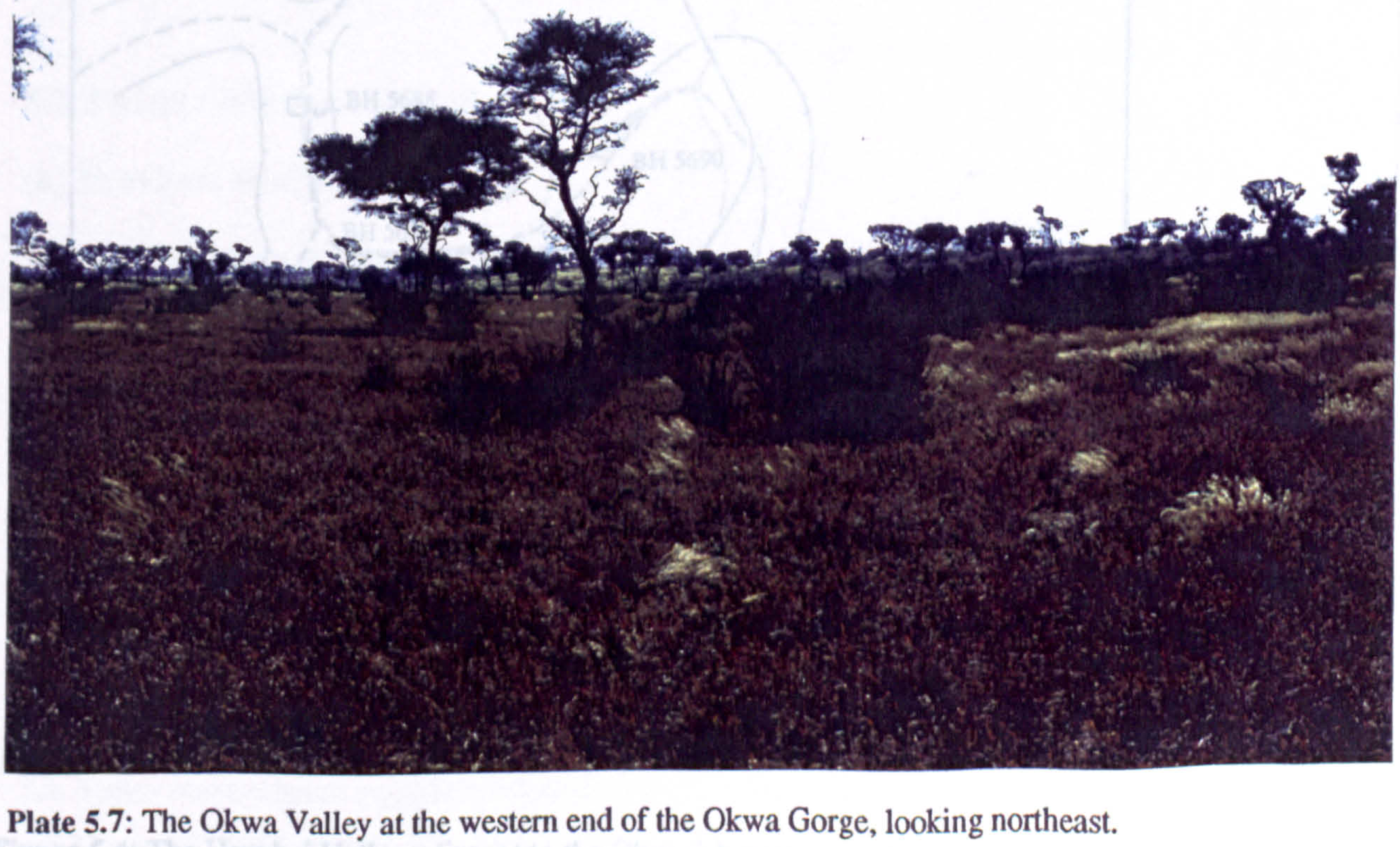


Plate 5.7: The Okwa Valley at the western end of the Okwa Gorge, looking northeast.

(ii) The Hanehai Valley

Field investigations in the Hanehai Valley were limited to where the Ghanzi-Jwaneng and Ghanzi-Mamuno roads intersect the valley (figure 5.4). The Hanehai rises at 1500 m asl as the Rietfontein or Buitsivango Valley in Namibia at approximately 21°55'S 19°15'E. It can be traced for over 200 km within Namibia, the headwater valleys rising on sub-cropping bedrock close to those of the Epukiro.

Comparing the height of the headwaters and the relative lengths of the Okwa and Hanehai, the Hanehai is the longer valley and rises at a higher altitude and could be expected to be the more prominant valley. This suggestion does not appear to be supported by field observations in the vicinity of the Ghanzi-Mamuno road, 82 km west of Ghanzi. The valley is approximately 200-300 m wide with a total depth of incision of only 6-8 m compared with the Okwa which is considerably deeper and wider at a similar length along its course. This difference is possibly attributable to the relative lengths of each valley situated respectively upon bedrock and more easily eroded Kalahari Group sediments; the Okwa has more of its course on Kalahari Group sediments up-valley of the confluence.

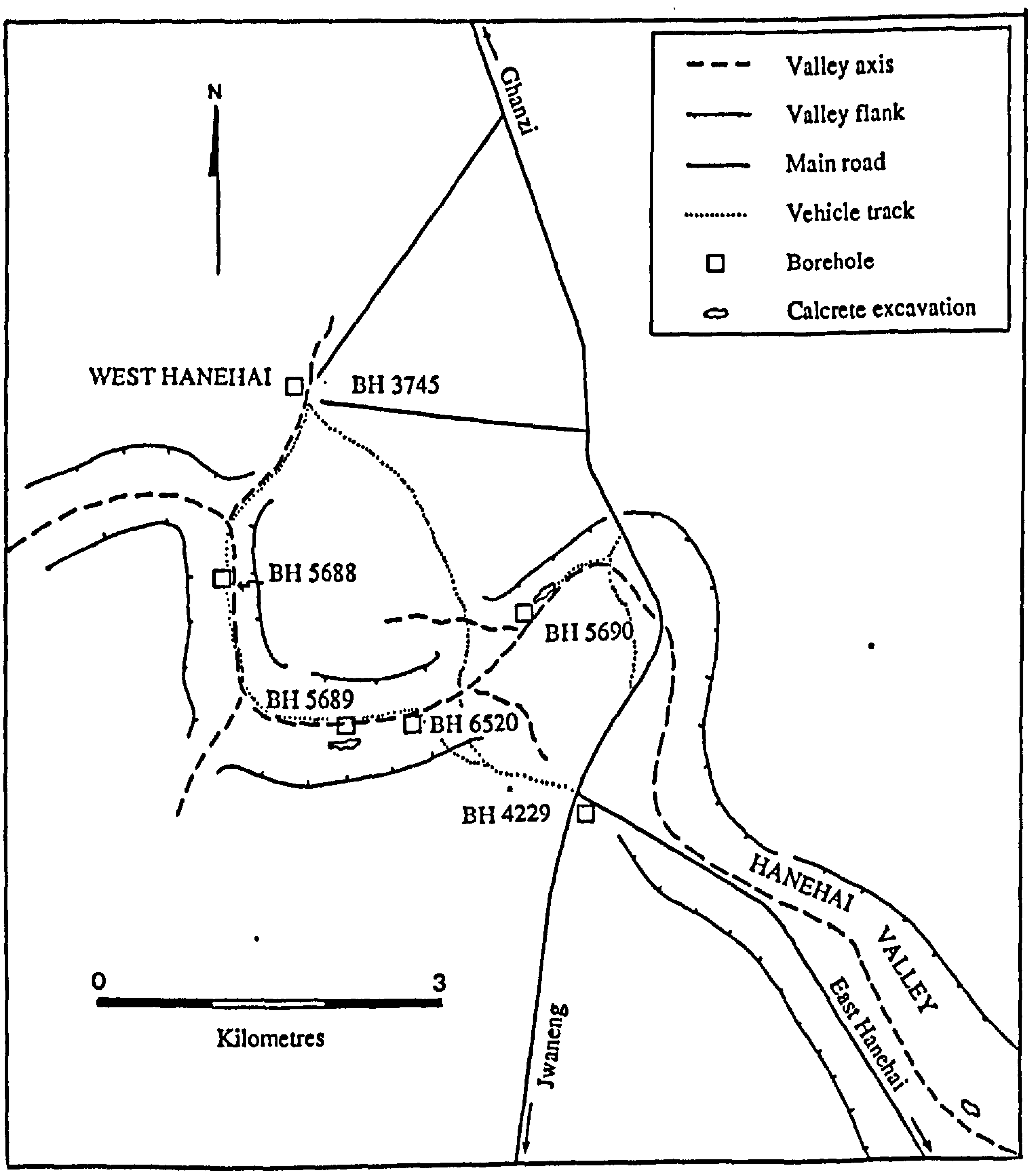


Figure 5.4: The Hanehai Valley adjacent to the Ghanzi-Jwaneng road.

The flanks of the Hanchai in this area are covered by Kalahari Sand, although the valley floor contains grey vertisolic soils typical of other *mekgacha*. It is likely that shallow groundwater exists beneath the valley in this section, the banks supporting an extensive tree and shrub cover, and large specimens of *Acacia erioloba* occurring on the valley floor.

The Hanchai was also studied where it is intersected by the Ghanzi-Jwaneng road, with the valley course meandering through the gently rolling Hanchai Hills. The valley in this region is probably unrepresentative of the Hanchai as a whole, its course being very wide (450-500 m in parts) and incised by between 10-15 m through outcropping Ghanzi Beds. In the immediate vicinity of the road any banks which may be present are obscured by a thick cover of Kalahari Sand which has been exposed due to considerable grazing pressure associated with livestock from nearby villages. However, away from the road the valley floor can be seen to consist of grey vertisolic soil where not infilled by sand.

Duricrust exposures occur in a number of locations, particularly within 5 km west and east of the main road. Calcrete, mostly in a powdery form with some more indurated hardpan exposures, outcrops mostly to the west of the main road. Exposures occur on the eastern valley flank to the north of borehole 5688 and in a series of bluffs up to 10 m high to the west of borehole 5689 (figure 5.4). Some upper sections show a karst-like surface with evidence of silicification in places. Powdery calcrete is also exposed in a 4.5 m deep pit to the north of the track to East Hanchai village, 5 km east of the main road. This pit contains 2.5 m of poorly indurated powdery calcrete overlain by 2 m of a grey-brown silty soil material. Fragments of bedrock and silcrete were also present in a spoil heap adjacent to this pit, but their relative stratigraphic positions are unknown. The base of the pit is below the level of the valley floor, and the duricrust exposures are a form of valley calcrete related to the presence of the valley (as seen in similar pits in the Okwa and Letlhakeng Valley 3). The calcrete/soil interface is abrupt, and dips at an angle of between 2° and 3° towards the valley axis, possibly reflecting the slope of a former higher water table level.

5.2.2 Mmone/Quoxo system

(a) Previous studies

The Mmone/Quoxo Valley system (hereafter referred to as the Mmone system) is the second most extensive of the Kalahari *mekgacha* networks after the Okwa (figure 5.5). The Mmone rises from near the edge of the Kalahari sandveld between Molepolole and Kanye, at the ill-defined watershed separating the drainages directed towards the Molopo River, the Limpopo River and the Makgadikgadi Depression (Aldiss *et al.*, 1989). From here the valley runs northward to join the Okwa in the Central Kalahari Game Reserve, with the Okwa-Mmone systems together sharing a potential catchment of approximately 90,000 km^2 (Thomas and Shaw, 1991*a*). Again, confusion exists regarding the nomenclature of the valley system, with various sections being given a variety of local names. For simplicity, the Mmone can be considered as the overall name, whilst major tributary valleys are treated individually, as shown in figure 5.6. Valley nomenclature and place names for this figure are based upon Botswana Department of Surveys and Lands 1:250,000 topographic maps.

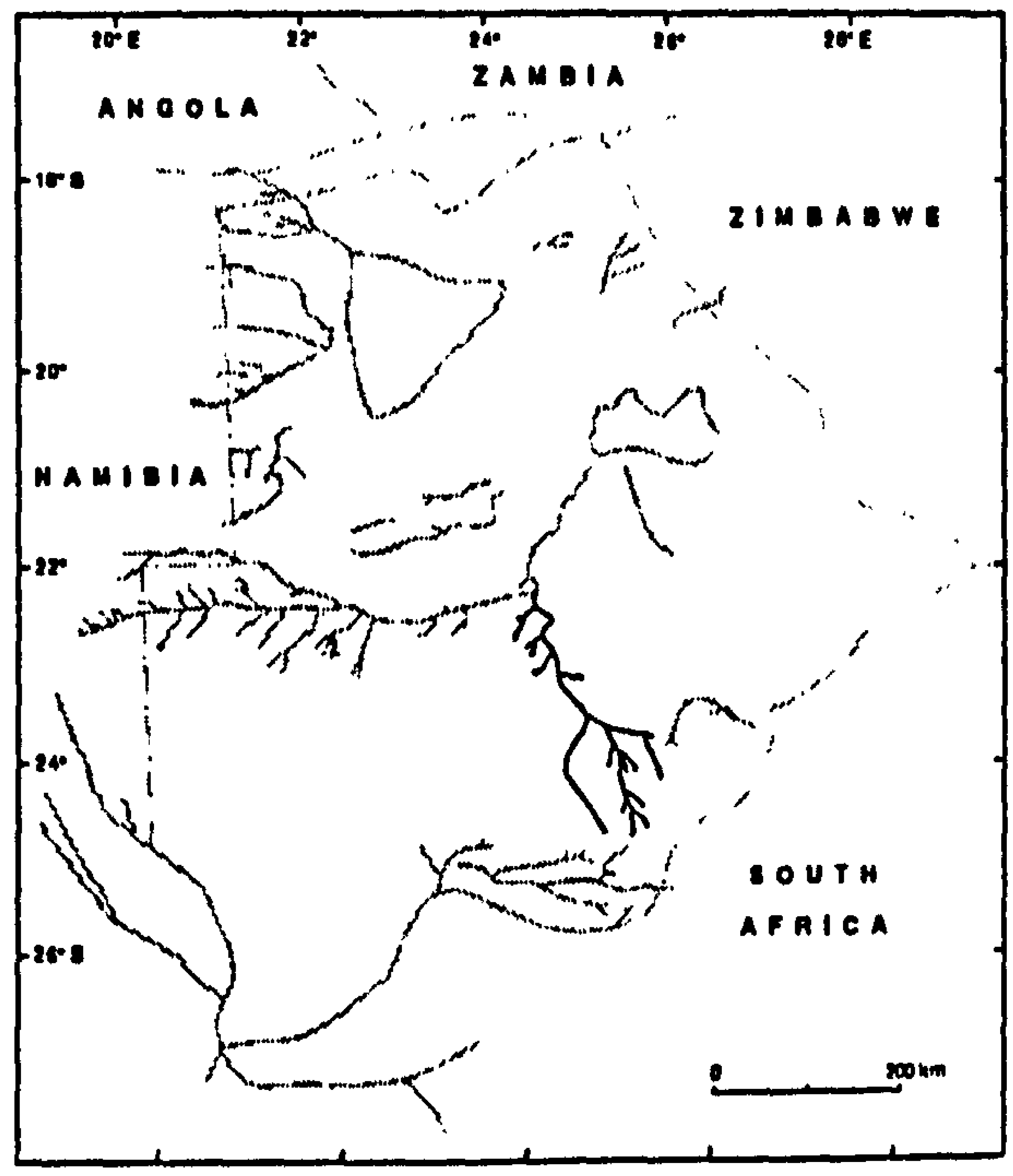


Figure 5.5: The Location of the Mmone/Quoxo valley system.

In comparison with the Okwa Valley, the Mmone has received little reference in the literature. The earliest reference is that of Hodson (1912 pp.79-80) who mentions the Meratswe near to Khudumelapye;

> "The last place of interest we passed was a dry river-bed... whence we could obtain beautiful water at any point by digging. My own theory is that this place, like many other similar dry river beds in the Kalahari has an underground river, otherwise it would be difficult to account for the presence of water so close to the surface."

The valleys in the vicinity of Letlhakeng village have received the most attention in conjunction with groundwater exploration. Indeed, the name "Letlhakeng" means "place of reeds" in local dialects (Campbell and Childs, 1971), suggesting the occurrence of water within the valley in the comparatively recent past. Groundwater and groundwater sapping processes (see section 2.3.1 above) are discussed by Shaw and De Vries (1988) for valleys to the south of Letlhakeng. Boocock and Van Straten (1962) were the first to note a pronounced "nick-point" in these valleys, also describing the Meratswe widening to over 1.5 km in places before narrowing 75 km north of Khudumelapye and eventually degenerating into a series of pans in the Central Kalahari Game Reserve. Farr *et al.* (1981) mention the valley systems as areas for potential groundwater recharge. Other descriptions include those of Buckley (1984) and Timje (1987) who note high groundwater yields in relation with drainage lines, particularly where valleys are associated with fracture systems.

A study in the Naledi Valley close to Jwaneng mine by Thoregren and Löfroth (1989) identified a much larger infilled channel existing beneath the present valley. The study utilised ground probing radar and identified a 50 m wide deposit-filled channel whilst the actual surface expression of the channel was only 5 m wide. Lawrance and Toole (1984) also identified an indistinct bifurcating buried "river course" some 40 km to the west of Jwaneng, near the village of Sedikane (approx 24°30'S, 24°20'E).

Aldiss *et al.* (1989) describe the morphology of the Dikhudu Valley, noting that the position of the main channel is controlled by northwest and north-northwest regional fractures. The valley shows an abrupt change in form, from a broad, flat "melapo" headwater area into a narrow incised valley, coinciding with the intersection of the channel and an east-northeasterly fault. Aldiss *et al.* (1989 p.122) also note the occurrence of silcrete and calcrete in the sides of the valley "...marking positions of successively older drainage channels."

Finally, the study of Kalahari geology by Mallick *et al.* (1981) notes the numerous small tributaries of the Mmone within the Khutse and Central Kalahari Game Reserves which have now been "beheaded" by a northwesterly-trending faultline.

(b) Field studies and aerial photography

The valleys of the Mmone system are shown in figure 5.6. Field surveys of these valleys were carried out during both 1989 and 1990, with work concentrating upon the Letlhakeng area. The results of field surveys are described below for the Letlhakeng-Meratswe, Naledi-Khwakhwe and Dikgonnyane valleys, with their associated tributary valleys.

(i) The Letlhakeng-Meratswe sub-system

The tributary valleys which converge in the vicinity of Letlhakeng village combine to form the Meratswe Valley. The Meratswe, together with the Kohiye Valley, forms the largest of the three sub-systems of the Mmone, with the Makiropetse and Dikhudu tributary valleys rising at the divide between the Makgadikgadi, Limpopo and Molopo drainage basins. Extensive outcrops of duricrusts occur in the valley flanks in the Letlhakeng area where the majority of field studies were undertaken. Duricrusts are briefly considered here but only in conjunction with valley form.

Detailed field studies in the Letlhakeng-Meratswe valleys included the following valley sections; 27 km of the Gaotlhobogwe valley (hereafter referred to as Letlhakeng Valley 1), 10 km and 11 km respectively of Letlhakeng Valley 2 and Valley 3 (the latter known as the Moshaweng Valley, not to be confused with the Kuruman tributary discussed in section 5.3.2 below) and the entirety of Valley 4 between Ditshegwane and its confluence with the Meratswe. Further reconnaissance level studies were made in the Meratswe between Letlhakeng and Kokosane villages and the Kohiye Valley at Marotswane.

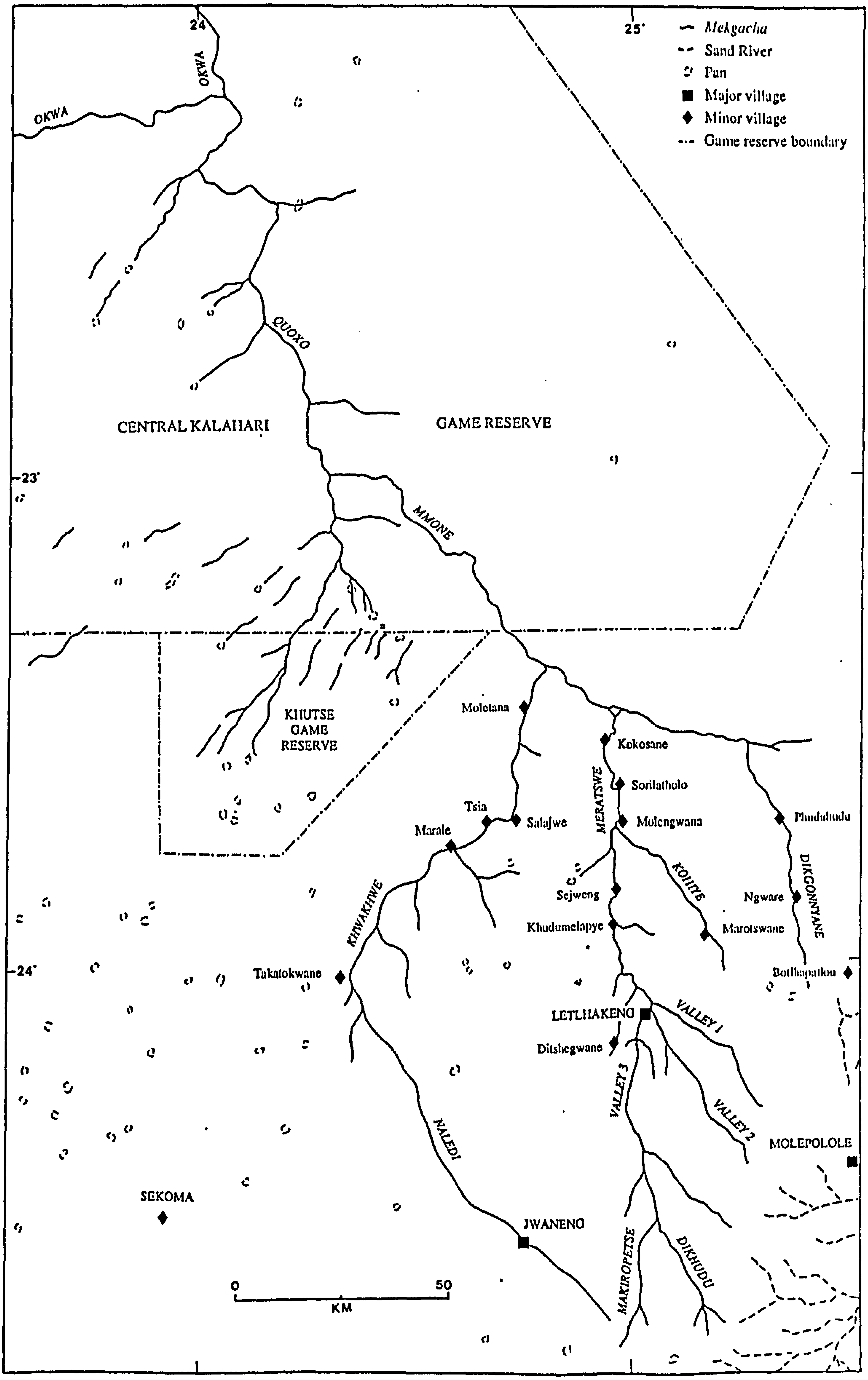


Figure 5.6: The valleys of the Mmone/Quoxo system.

The Gaotlhobogwe Valley (Letlhakeng Valley 1)

Letlhakeng Valley 1 is unique amongst the *mekgacha* investigated as it has an abrupt drop of over 10 m in its course at 24°09'35"S 25°12'00"E, exposing duricrusts in an amphitheatre-shaped valley head. Other valleys in the Letlhakeng area show a rapid change in form, typically with the development of steep valley flanks, but none have such an abrupt transition. Up-valley of the transition point Valley 1 has a broad shallow form with negligible relative relief (plate 5.8), a floor gradient of 0.21° and no clearly identifiable channel. Below the amphitheatre valley head, the valley deepens considerably, reaching a depth of incision of 35 m below the surrounding Kalahari Sand plateau and exhibiting a gorge-like form with steep flanks within 4.0 km of the nick-point. This deepening is accompanied by rapid widening to between 500 and 700 m, although widths of up to 1.5 km are attained further northwest. Valley floor gradients along the gorge section average 0.36° for the 8 km below the amphitheatre head. Again, no major channel occurs, although there are small semi-continuous gullies along the valley floor.

The amphitheatre valley head area is of interest as its headwalls are neither as abrupt or as sheer as the examples of valleys developed by groundwater sapping processes discussed in section 2.3.1 above. There is also indication of a channel having been cut through the headwall at its eastern end. The valley head is depicted in plan-form in figure 5.7 and in plate 5.9, which indicate the steepness of the duricrust cliffs and the abrupt drop from the surrounding plateau.

The upper duricrust surface (where not covered by Kalahari Sand) exhibits karst-like features, having a pavement appearance and crocodile-skin weathering textures. The maximum vertical exposure of massive, horizontally- and vertically-jointed silcrete was 5 m on the north side of the valley head. The upper sections of these cliffs were pitted by numerous holes and hollows, with sand-filled tube-like features up to 8 cm diameter occurring in the pavement surface and in the upper 0.5 m. These large, isolated tubes appear to be a form of pseudo-karstic surface weathering (Marker, 1976), but whilst being mentioned in the literature on silcrete (e.g. Shaw and De Vries, 1988) have not been satisfactorily explained.

A common feature of many silcrete exposures, particularly in the eastern end of the valley head area, was the presence of an overhang at the base of the duricrust cliff. The overhang had a maximum indentation of 0.5 m into the cliff face, and, as postulated by Shaw and De Vries (1988), probably represents a relict springline. This is suggested by the presence of numerous interconnected horizontal to sub-horizontal tube-like holes up to 2 cm diameter in a 20 cm zone at the base of the overhang. This narrow band of macroscopic tunnels may represent a seepage zone as observed in headwalls of networks developed by sapping processes in the Colorado Plateau (cf. Laity, 1983). The tunnels are of a different scale and form to those exposed in higher sections of the duricrust cliffs. If localised weakening of the rock by seeping water had caused the overhang, it might be expected that the silcrete in the overhang would be more friable than the rest of the exposure. However, the silcrete in this zone was only slightly weaker than the material exposed higher up the profile.

Plate 5.8: Letlhakeng Valley 1 at borehole 6479 up-valley of the amphitheatre valley head, looking northeast.

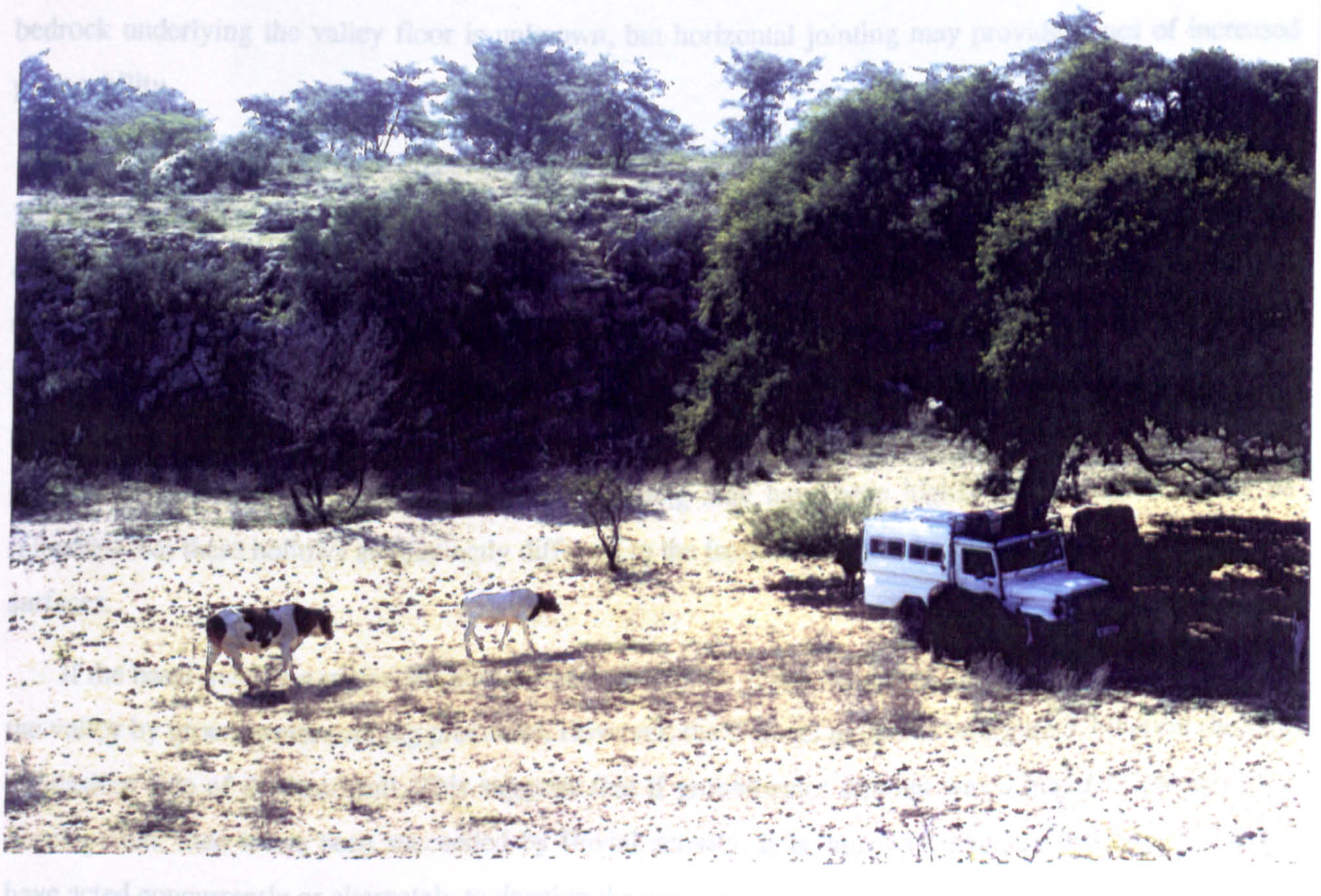


Plate 5.9: Silcrete cliffs in the amphitheatre valley head of Letlhakeng Valley 1, looking north.

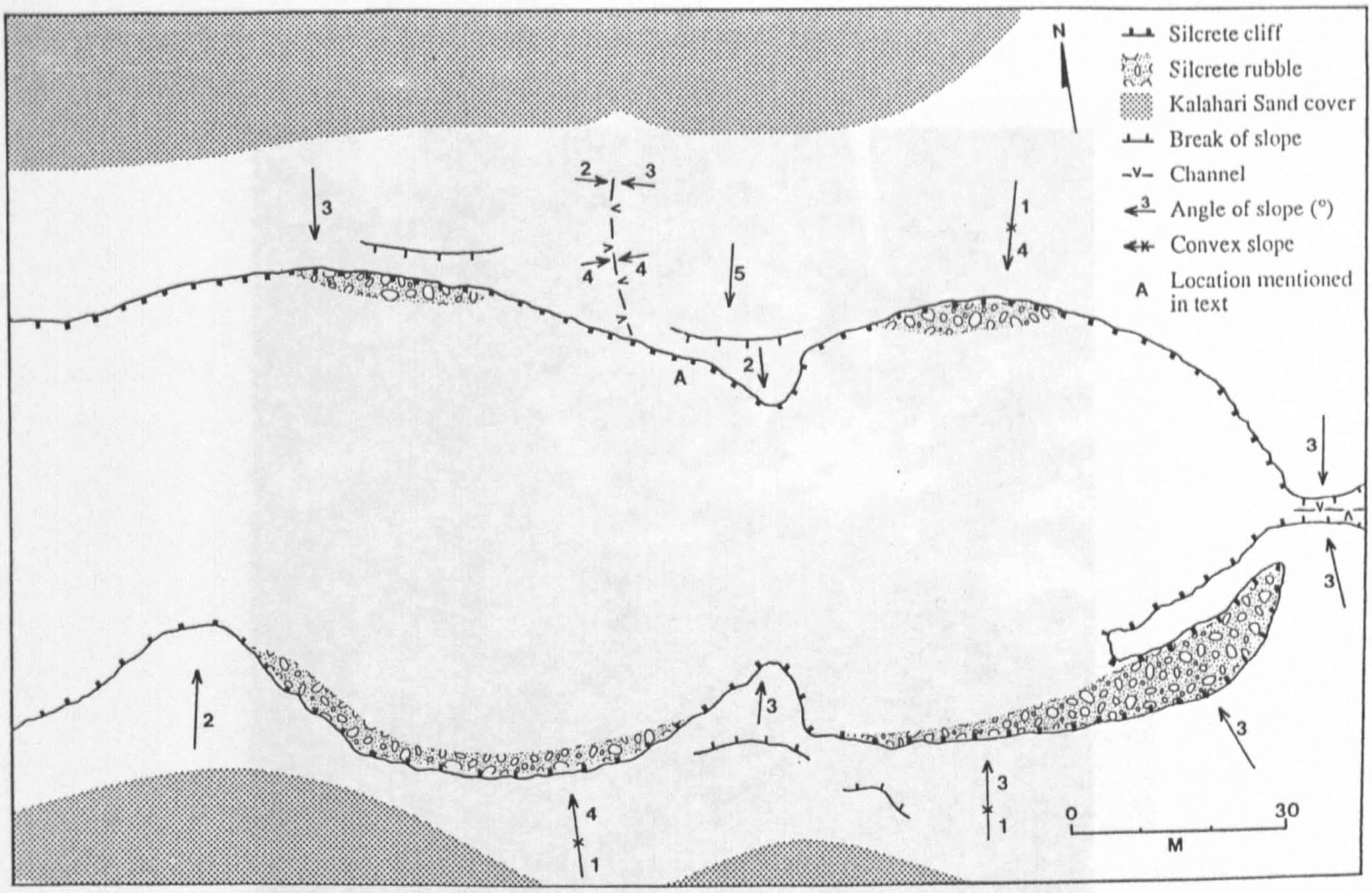


Figure 5.7: Morphology of the amphitheatre head in Letlhakeng Valley 1 (the Gaotlhobogwe Valley).

In Colorado Plateau examples, groundwater emergence was usually associated with a reduction in lithological permeability immediately beneath the seepage zone (cf. Baker, 1990). In Valley 1 the depth to bedrock underlying the valley floor is unknown, but horizontal jointing may provide zones of increased permeability.

Similar zones of solutional tubes were also seen higher up the silcrete cliffs. At point A on figure 5.7 a 30 cm thick horizontal zone of sub-horizontal tunnels (1 cm high and 3 cm long) showing no preferred orientation occurs 1.4 m up the cliff face (plate 5.10). The reason for this localised zone of tunnels is unclear, as unlike the material exposed in the basal overhang, the silcrete here shows no obvious lithological change. It may be that zones of hollows above ground level indicate seepage sites when water tables and/or the valley floor were higher. Thomas and Shaw (1991*a*) have cited lines of evidence which suggest that water tables in the vicinity of Letlhakeng may have been higher in the past. It should be reiterated that these hollows are distinctly different to the isolated macro-tubes developed on upper silcrete surfaces.

If the basal overhang represents a relict springline at the valley head, this may indicate development of the valley by groundwater sapping processes. However, the channel indicated on figure 5.7 has cut through the eastern side of the headwall. This suggests that if groundwater sapping was a major factor in valley development, then it has been succeeded by fluvial activity. It is, however, possible that both processes have acted concurrently or alternately to develop the present valley form.

Plate 5.10: Silcrete profile at site 2 in the amphitheatre valley head of Letlhakeng Valley 1, showing the basal overhang and a zone of tunnels at 1.4 m.

Soils in the valley head area are dominantly inblown Kalahari Sand, with an increase in clay content producing sandy vertisolic soils down-valley. Windblown sand at the amphitheatre head area appears to be associated with removal of vegetation from neighbouring valley slopes by cattle. Aerial photography of the valley (plate 5.11) indicates a change in vegetation in a circular zone of 2.5 km radius around a borehole situated 500 m from the valley head. A number of pans also occur in the valley floor containing more organic-rich soils.

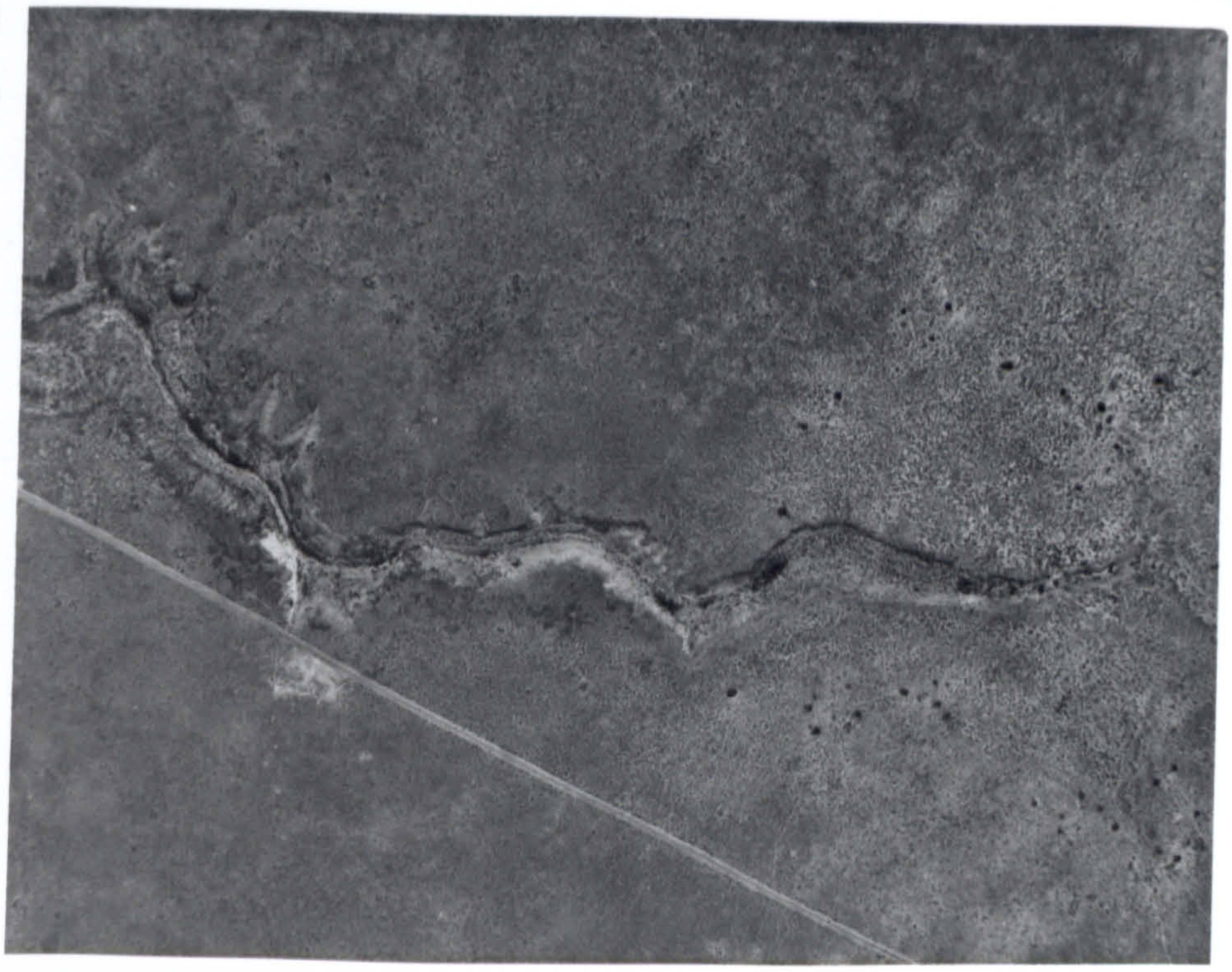


Plate 5.11: The amphitheatre valley head and gorge section of Letlhakeng Valley 1 (from aerial photography; scale 1:50,000). Botswana Department of Surveys and Lands photo number S.E.B. 15 (115) dated 22 May 1989.

Plate 5.12: The gorge section of Letlhakeng Valley 1 showing a duricrust terrace.

As noted above, the form of Valley 1 changes rapidly away from the valley head, reaching a maximum depth of incision of over 35 m. The typical form of the valley for 8 km northwest consists of a comparatively flat floor, with gently concave slopes of 3° to 7° leading up to steeper debris slopes beneath duricrust outcrops. The lower slope from the valley floor consists of fine sand of a probable aeolian origin. Where well-developed debris slopes occur, the sand usually covers the lower sections of the slope. Maximum debris slope angles of between 25° and 28° occur on the outside of bends in the valley. Further to the northwest as the valley approaches Letlhakeng village, slope angles decline and the valley widens considerably, reaching a maximum width of around 1.5 km to the east of the village. Duricrusts outcrop as low cliffs or bluffs, usually separated by a wide flat valley floor or partly buried by aeolian sand. Valley depth is maintained at around 30m below the surrounding Kalahari Sand plateau level, with the increase in width giving the impression of declining overall depth.

The extent of duricrusts are discussed in the following chapter, but some observations regarding the distribution of duricrust types can be made here. Firstly, it should be noted that the Kalahari Group sediments overlying Karoo bedrock are very thin in this region, with a maximum thickness of 12 m underlying the valley floor 4 km northwest of the valley head (section 6.2.2c). Karoo siltstone outcrops occur in the valley floor 5.5 km from the valley head, at 24°06'27"S 25°06'30"E. Secondly, both silcrete, calcrete and intermediate duricrust types occur within the valley flanks, most upper slopes being dominated by silcrete exposures. Hardpan calcretes also outcrop in the valley floor close to Letlhakeng village. The upper surfaces of duricrust cliffs form a terrace level, at a height of approximately 30 m above the valley floor (plate 5.12), with over 5 m of unconsolidated Kalahari Sand overlying the outcrops at slopes of around 5°.

The overall form of the Valley 1 gradually changes to the northeast of Letlhakeng where the valley joins with the combined Letlhakeng Valleys 2 and 3 to form the Meratswe Valley. Slope angles of only 3° to 4° occur with a maximum depth of incision of 18-20 m, with the two valleys separated by a low sandy spur. Slopes to the west and south of the confluence area are steeper than those to the east. Calcrete exposures in the vicinity of the confluence are limited and tend to consist of weakly indurated varieties. A well at the confluence indicates sandy, grey vertisolic soils to a depth of at least 1 m, with no evidence for calcification.

Letlhakeng Valley 2

Valley 2 is an unnamed tributary of the Meratswe valley immediately south of Letlhakeng village. As with Valley 1, a change in form occurs to the south of Letlhakeng, the transition point shown on the southern part of plate 5.13. South of this area, the valley has a poorly defined and shallow form, changing to a much more deeply incised, relatively steep sided valley over a distance of less than 500 m. No channel is present, although some localised gullying of the valley flanks has taken place. As can be seen from plate 5.13 two valleys join in the incised section separated by a low ridge, with both exhibiting this change in form.

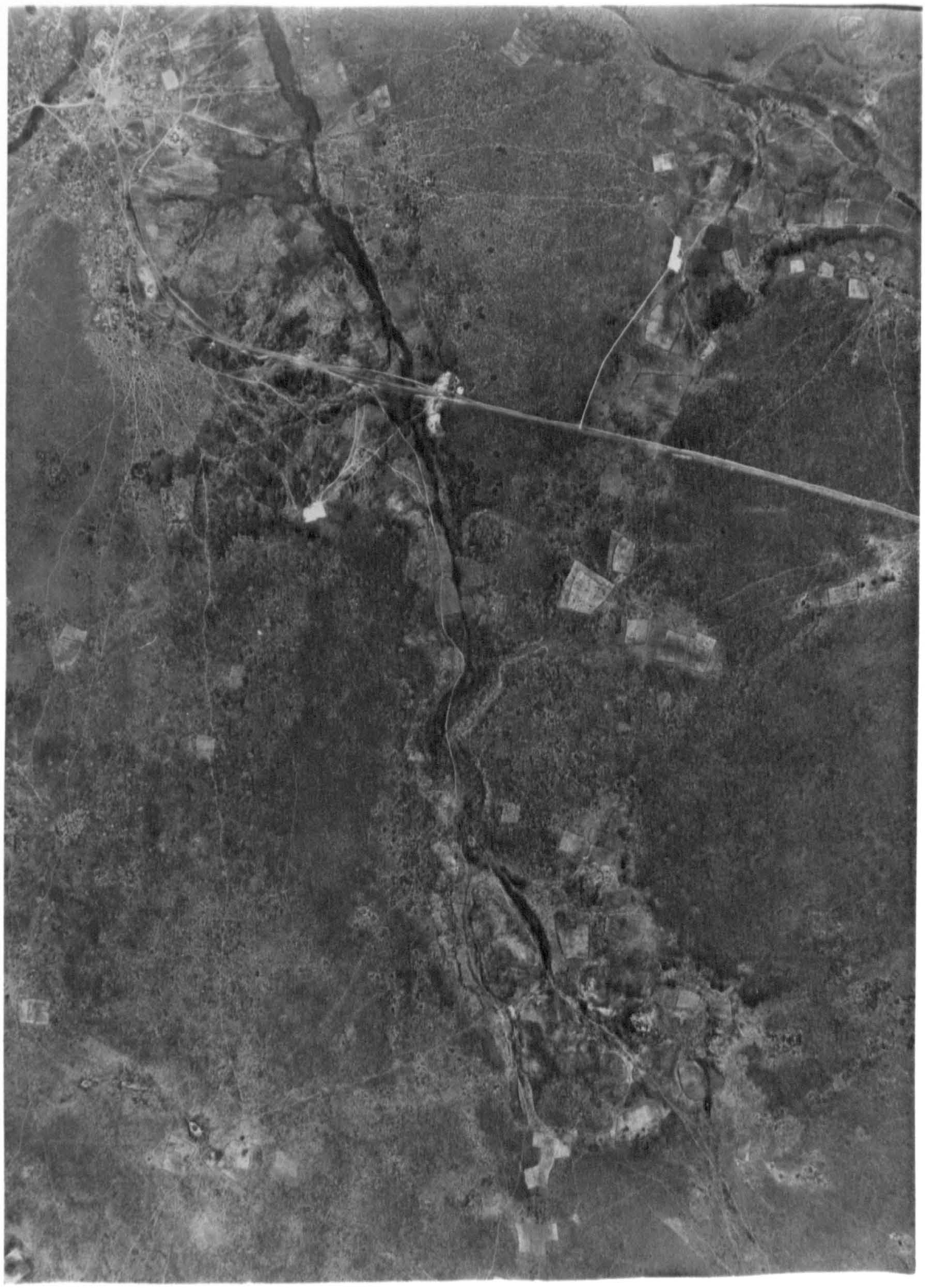


Plate 5.13: Letlhakeng Valley 2 (from aerial photography; scale 1:50,000). Botswana Department of Surveys and Lands photo number S.E.B. 15 (112) dated 22 May 1989.

Plate 5.14: The unconformity between the basal Kalahari Group conglomerates (background) and the underlying Kweneng Sandstone.

Unlike Valley 1, Letlhakeng Valley 2 does not have a clearly defined head but does exhibit similar steep flanks and extensive duricrust exposures. Cliff-like exposures of duricrusts are uncommon, most outcrops occurring as valley slopes at angles of up to 16° (but more commonly 7° to 8°). The maximum depth of incision of Valley 2 is also less than Valley 1, with a relative relief of less than 25 m for the deepest part of the valley, 4 km north of the transition point. Duricrusts tend to be dominated by conglomeratic cryptocrystalline silcretes in southern exposures and massive hardpan calcretes to the north, karst features being common on horizontal surfaces not covered by Kalahari Sand. Extensive outcrops of Karoo sandstone occur in the southern parts of the valley, immediately overlain by conglomeratic duricrusts which form the basal Kalahari Group sediments (plate 5.14).

With two valleys joining in the incised section of the valley, the valley form in southern sections is broad compared to further north of the confluence area. Here the valley narrows to around 400 to 500 m, with a narrow flat sandy vertisolic floor, and hardpan calcrete exposed in the valley flanks. Similar calcrete exposures occur in the tributary valley which enters on the east side 3.5 km down-valley of the transition point, although the calcrete in the tributary becomes less indurated away from the main valley.

Slope angles decline to less than 10° where the main road crosses the valley (see plate 5.13) with only limited rubbly calcrete outcrops occurring at this point. The overall depth of incision declines to around 15-20 m where Valley 2 joins with Valley 3 to the northeast of Letlhakeng village, with an associated decline in slope angles to between 5° and 7°.

The Moshaweng Valley (Letlhakeng Valley 3)

The Moshaweng valley (hereafter referred to as Valley 3) is the longest of the Meratswe tributaries, being the result of the merging of the Makiropetse and Dikhudu valleys (already both 25 km long) some 50 km to the south of Letlhakeng. The valley was studied over a distance of 11 km to the south of the village, this length representing the extent of major duricrust exposures.

As with the two Letlhakeng valleys described above, Valley 3 contains no obvious channel and shows a change in valley form in the southern part of the study section. The aerial photograph in plate 5.15 includes the main valley and two small tributaries to the south of Letlhakeng village. To the south of the confluence of the two tributaries and the main valley, the flanks have very gentle slopes averaging 3° to 4°. These slopes are comprised of Kalahari Sand overlying Karoo Kweneng sandstone, which intermittently outcrops as slabs and boulders on the valley flanks. Approximately 2 km south of the confluence the valley slopes steepen considerably, with extensive calcrete exposures occurring to the north of this point immediately overlying the sandstone (see section 6.2.1). Plate 5.16 shows the valley form at the confluence, indicating the flat valley floor and steeper flanks to the north of the confluence.

The transition is not as abrupt as in Letlhakeng Valley 1, but the contrast in valley form north and south of the confluence area is just as pronounced. To the north the valley is at most 500 m wide, but more typically 200-300 m, with flanks consisting of duricrust and debris slopes at angles of up to 23°. Near-vertical duricrust cliffs over 5 m high occur in places at the top of these slopes. Again, the horizontal uppermost surfaces of these duricrust exposures show karst-like features. The total depth of incision of the valley beneath the surrounding Kalahari Sand is at least 25 m at its deepest point.

The main pattern of duricrusts is of silcrete near the confluence area, but calcrete elsewhere. Localised outcrops of Kweneng sandstone occur, mainly in the centre of the main valley immediately north of the confluence area and at the distal end of the eastern tributary. The valley "thalweg" is diverted to either side of the former sandstone outcrop, the split in the thalweg indicated by the diversion in the track on plate 5.15.

Only one major pan occurs in the valley floor, 2 km north of the confluence area (plate 5.15). The pan is dammed down-valley to trap water for cattle during the wet season. This pan contains the only area of organic-rich vertisolic soil in the valley floor, the remaining soils consisting of clay-rich sandy vertisols with localised nodular calcrete development (as indicated by a series of 2 m deep hand-dug wells along the valley). In contrast, south of the confluence area the valley floor is mostly filled by inblown Kalahari Sand.

The tributaries to Valley 3 are of interest, mainly due to their perpendicularity to the main trunk valley. The tributaries have a similar form to the main trunk valley in the vicinity of the confluence, but become much less clearly defined towards their headward ends. Each has a maximum depth of incision of around 15 m and contains extensive duricrust exposures. The most probable suggestion for the perpendicularity of the confluence zone is some form of east-west structural control of the intersection. Interpretation of Landsat imagery by Mallick *et al.* (1981) indicates a structural lineament in this area.

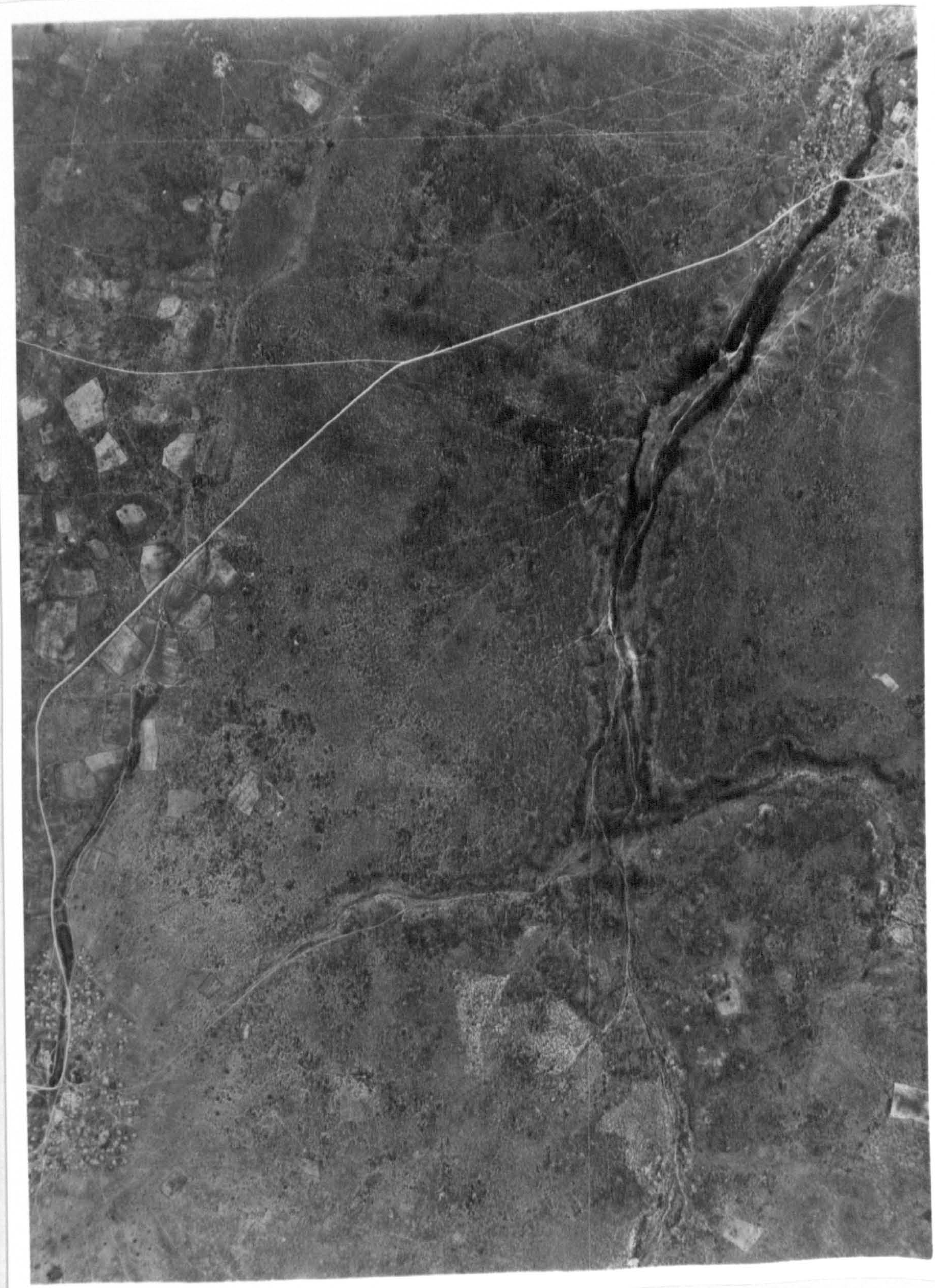


Plate 5.15: Letlhakeng Valley 3 (from aerial photography; scale 1:50,000). Botswana Department of Surveys and Lands photo number S.E.B. 15 (110) dated 22 May 1989.

Plate 5.16: Letlhakeng Valley 3, looking south across the confluence of the main valley and two perpendicular tributaries.

Plate 5.17: The northern section of Letlhakeng Valley 3, looking northeast.

In the northern sections of Valley 3, before it merges with Valley 2 near Letlhakeng, the overall valley form becomes much more subdued. Maximum slope angles decline to around 8°, the valley flanks being strewn with calcrete rubble, with an average width of 600-700 m (plate 5.17). Letlhakeng village itself is situated on the floor and flanks of Valley 3, with the large areas of hardpan calcrete exposed in the valley floor used as building foundations.

The confluence of Valleys 2 and 3 to the north of Letlhakeng has very gentle valley flanks, with slope angles of less than 2°. A 5 m high interfluve consisting of calcrete rubble separates the two valleys.

Letlhakeng Valley 4

The Tswaane Valley (Letlhakeng Valley 4) was studied at a reconnaissance level during 1989. In its headwater section near to Ditshegwane village, the valley has a very subdued dambo-like form (Shaw and De Vries, 1988). The valley floor consists primarily of unconsolidated red Kalahari Sand with no evidence of either duricrust outcrops or powdery calcrete in the valley floor soils. The form begins to change at a distance of 2.5 km from the Meratswe Valley confluence, with an overall narrowing and deepening. The valley reaches a maximum depth of incision of approximately 15 m, with a width of around 300-400 m and valley flank slopes of 8-10°. Duricrust outcrops are common on the valley flanks in this more incised section, with calcrete exposures up to 1 m high occurring at a height of 6-7 m above the valley floor.

At the confluence with the Meratswe, slope angles decrease to around 3-4° and duricrust outcrops are less common. Where calcrete does occur it is invariably either as rubble mantling slopes or as a hardpan to nodular variety forming the slope. The main calcrete exposures of this latter type are to the north of the confluence, with rubble more common elsewhere. Valley floor materials in the incised section of the valley consist of grey sandy vertisolic soils, with hardpan calcrete also outcropping in the vicinity of the confluence.

The Meratswe Valley between Letlhakeng and Kokosane

The form of the Meratswe Valley at Letlhakeng has already been described above. Field studies were undertaken in 1989 between this point and the village of Kokosane over 75 km due north of Letlhakeng.

The form of the Meratswe in the section immediately north of the confluence with Valley 4 is similar to that of the northwestern end of Valley 1, with a total depth of incision of between 15 and 20 m, a width of approximately 1.5 km and valley slopes of 7° to 8°. Duricrust exposures tend to be limited to rubbly outcrops on the upper valley flanks and weakly indurated hardpan calcretes in the valley floor. The floor is lined with a number of shallow pans and swings through a series of meanders in this section. Variations in slope angle are consistent with temperate fluvial valleys, with steeper slopes associated with outer banks. In the vicinity of Khudumalapye the valley depth has diminuished to between 5 and 6 m with a width approaching 500 m . There are extensive outcrops of hardpan calcrete in the valley floor and flanks within the village, overlain by red Kalahari Sand.

Plate 5.18: The Meratswe Valley south of Kokosane village.

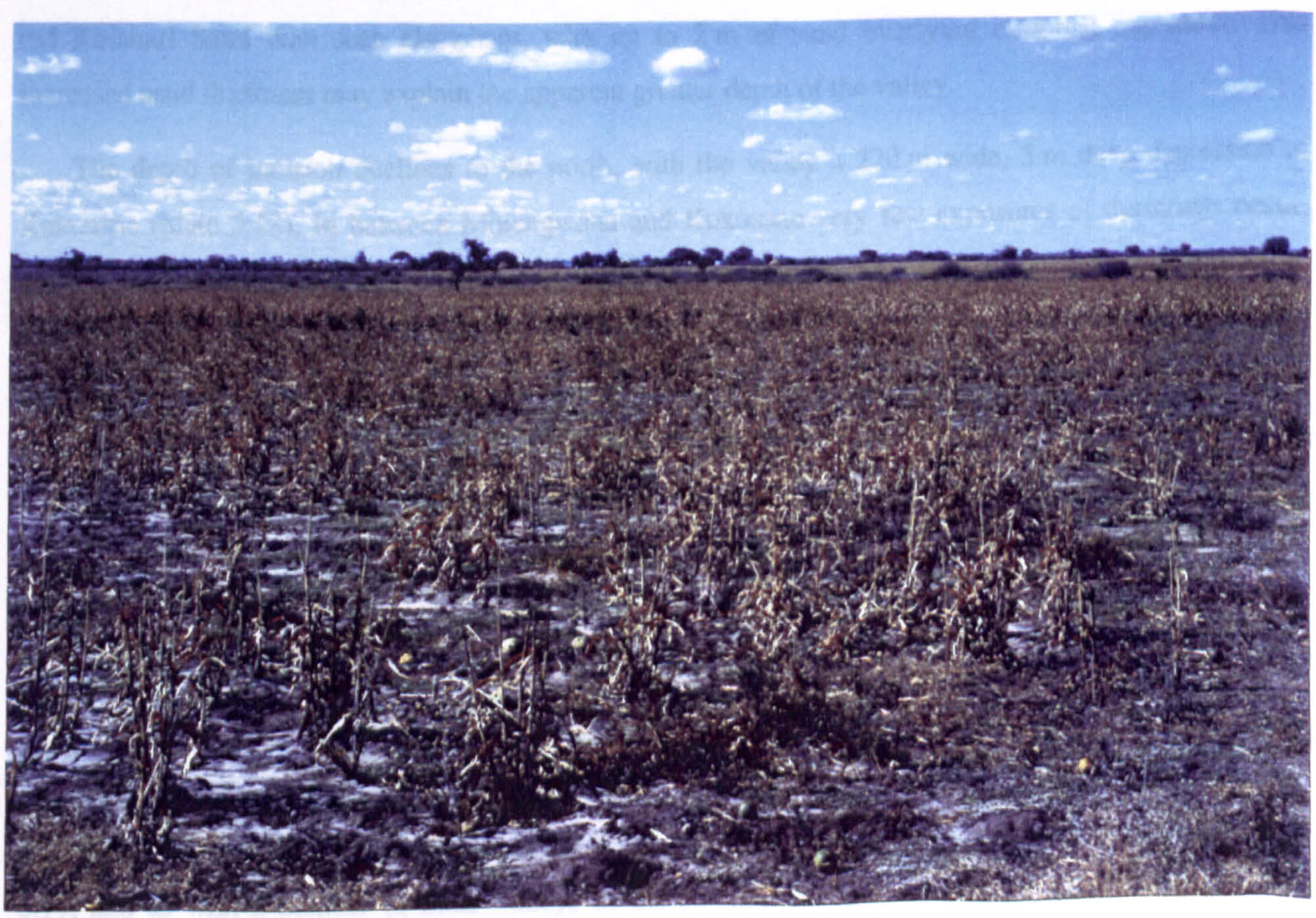


Plate 5.19: The Kohiye Valley south of Marotswane village.

Valley form undergoes another transition to the south of Sejweng borehole where red and pale pink Ecca Group sandstones outcrop. Fragments of sandstone can first be seen as spoil around a borehole 1.8 km north of the centre of Khudumalapye. The valley is incised through outcrops of this sandstone for 1.4 km south of Sejweng borehole, with outcrops disappearing for 300 m and then a further 800 m of outcrop. The southernmost sandstone outcrops were exposed on the west bank of the meander bend indicated immediately south of the borehole on figure 5.6. The valley shows definite variations in form due to the differing resistances of rock types in this area. In bedrock sections the valley sides are near-vertical, with bedrock cliffs reaching heights of over 6 m. The valley floor narrows to 100-150 m in incised sections and is commonly filled with sediment, suggesting that the valley may have been considerably deeper in the past. In contrast, in the 300 m section with no sandstone outcrop the valley flanks have maximum slope angles of only 8-10° and the valley floor is approximately 200 m wide. Outcrops of hardpan calcrete occur in the valley floor, whilst only powdery duricrust occurred within bedrock sections. To the north of Sejweng the valley widens to around 400 m but becomes less clearly defined. Outcrops of hardpan calcrete are less common, with the valley flanks consisting of red Kalahari Sand. The valley broadens considerably to over 1 km at Kokgole borehole where a poorly defined tributary valley enters from the west.

The Meratswe shows another change in form where it joins with the Kohiye Valley to the south of Molengwana village. The valley width remains constant at 500 m, but the depth of incision increases to 8 m, with slope angles increasing to a maximum of 7-8°. This increase in apparent depth may be due to increased incision down-valley of the confluence. However, the banks do consist of greater thicknesses of red Kalahari Sand than seen elsewhere, with up to 2 m of sand overlying duricrust exposures. This increased sand thickness may explain the apparent greater depth of the valley.

The depth of incision declines to the north, with the valley a 220 m wide, 5 m deep depression at Kokosane (plate 5.18). In between Molengwana and Kokosane very few exposures of duricrusts occur, notable exceptions being a localised outcrop of highly indurated cal-silcrete at Sorilatholo borehole, a low calcrete rubble terrace south of this village, and calcrete exposures around pans to the south of Kokosane. The terrace level only occurs over a distance of some 200 m and consists of a continuous bluff with a cover of aeolian sediment. The valley was followed no further north than Kokosane, but interpretation of aerial photography suggests that the Meratswe becomes even less clearly defined once it enters the Central Kalahari Game Reserve.

The Kohiye Valley at Marotswane

The headwaters of the Kohiye Valley were studied at Marotswane village during 1989. The valley can be accessed from west or east by taking the track from Letlhakeng village to Botlhapatlou via Marotswane. In the vicinity of Marotswane the valley is extremely broad and shallow with a very low relative relief (plate 5.19, and as with a number of other valleys of the Mmone system, can only be identified due to the presence of grey vertisolic soils marking the extent of the valley floor. The absence of any clearly identifiable valley flanks made estimation of valley width and depth impossible.

(ii) The Naledi-Khwakhwe sub-system

The Naledi Valley rises to the southeast of Jwaneng village, eventually joining with the Meratswe Valley south of the Central Kalahari Game Reserve boundary. The valley is known as the Naledi in its southern section and the Khwakhwe to the north of Takatokwane village. Field studies in 1989 concentrated upon the Khwakhwe Valley between Takatokwane and Moletana, with the headwater section of the valley at Jwaneng also briefly visited.

At Jwaneng, the Naledi is a broad, shallow, almost imperceptible feature, only visible on remotely-sensed imagery due to the slightly darker soil associated with the valley. This form is maintained as far north as Takatokwane beyond which the valley deepens slightly and becomes more clearly defined. Immediately north of Takatokwane the Khwakhwe is between 300 and 400 m wide with a total depth of only 3-4 m. Between this point and Tsia village duricrust outcrops occur only intermittently in the valley flanks, mainly in the vicinity of pans at Marale. However, hardpan calcretes are commonly exposed within the firm sandy vertisolic floor.

A number of pans occur between Tsia and Salajwe villages, with associated ferricrete outcrops exposed for over 800 m on the western flank 2 km south of Tsia. Hardpan calcrete outcrops around Salajwe, where the valley is still broad and has a depth of 5-6 m. The form of the Khwakhwe changes very little over the study section as far north as Moletana, the only variation being a gradual increase in depth of incision to a maximum of 7 m (plate 5.20).

(iii) The Dikgonnyane Valley

The Dikgonnyane Valley was studied during 1989, investigations taking place to the west of Botlhapatlou and between Ngware and Phuduhudu villages. The southernmost point at which the Dikgonnyane was encountered was 8.5 km along the east-west cut-line from Botlhapatlou to Marotswane (which can be reached by taking the north road from Botlhapatlou and turning west 4.5 km north of the village). Here the valley is broad and almost completely flat, only being distinguishable from the surrounding Kalahari Sand by virtue of the grey sandy vertisolic soil in the valley floor. There is also a change in vegetation with grass the dominant vegetation type in the valley whilst trees and shrubs occur away from the floor. Aerial photography indicates that the cut-line traverses the valley perpendicular to its course, and as grey soil occurs for over 800 m along the track, the actual valley is probably in excess of 1 km wide at this point.

At Ngware, 15 km to the north, the Dikgonnyane is between 8 and 10 m deep suggesting a relatively rapid incision (plate 5.21). The valley is approximately 800 m wide with sandy flanks and a grey, clay-rich vertisolic soil floor. No duricrust exposures were noted in the valley flanks at Ngware. Within 5 km north of the village, hardpan calcretes do outcrop in the valley, suggesting that they may occur beneath valley floor sediments. Such calcretes are typically poorly indurated porous varieties, possibly indicative of comparatively recent formation compared to harder partly silicified types seen in other valleys of the Mmone system.

Plate 5.20: The Khwakhwe Valley south of Moletana village.

Plate 5.21: The Dikgonnyane Valley at Ngware, looking east.

Within 10 km north of Ngware the form of the Dikgonnyane Valley has returned to a barely perceptible shallow depression, again only identifiable by the presence of grey vertisols. This form is maintained until the northern extremity of the study section at Phuduhudu, with the only change being an increase in the aeolian sand content of the valley floor. Studies of aerial photography to the north of this area, up until the confluence with the Meratswe, show little change in form. In places (e.g. around 25 km east of the confluence) the valley almost disappears to a string of interconnected pan-like depressions. This valley form continues into the Central Kalahari Game Reserve.

5.2.3 Central Kalahari valleys

(a) Previous studies

The Rooibrak/Passarge and Deception valleys are located in central Botswana, with their courses running in an easterly direction from barely perceptible headwaters (a series of small pans) to the east of Ghanzi village and ending on the fringes of the Makgadikgadi Depression (figures 5.8 and 5.9). On many maps the Rooibrak is referred to as the Rooibok Valley, whilst the western part of the Deception is termed the Letiahau. No major studies have been undertaken on either valley, although the general form of the Deception Valley is described in Owens and Owens (1984), and the valleys are mentioned in conjunction with the Makgadikgadi Depression by Helgren (1984) and Breyer (1982).

To date, the geomorphology of the Central Kalahari valley systems has only been discussed as part of other studies. These include the Kalatraverse project of the Botswana Department of Geological Survey (Coates *et al.*, 1979) and the uranium prospecting activities of Union Carbide (1979*a*, *b* and *d*, 1980*b*). The courses of both valleys are indistinct and marked in places only by a string of pans, with Silberbauer (1981 p.45) noting that the Deception Valley "...loses itself for some miles..." in the Piper Pan complex. Indeed, the more northerly valley is named the Rooibrak at its headwater end before disappearing into a disjunct line of pans and reappearing further east where it is named the Passarge.

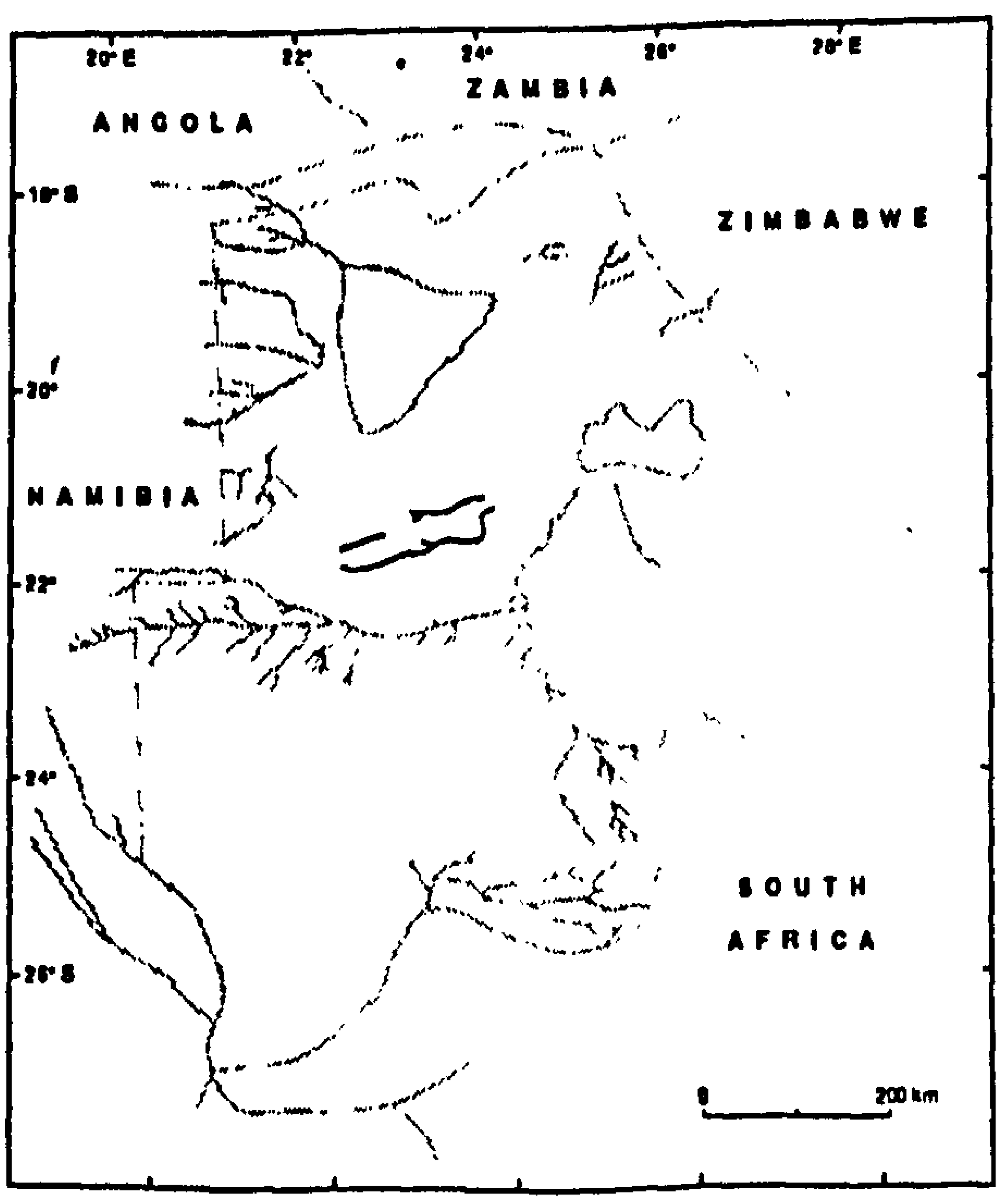


Figure 5.8: The location of the Central Kalahari valley systems.

Figure 5.9: The Rooibrak/Passarge and Deception valleys.

Plate 5.22: The Rooibrak Valley 31 km due east of De Graaf farm, looking east.

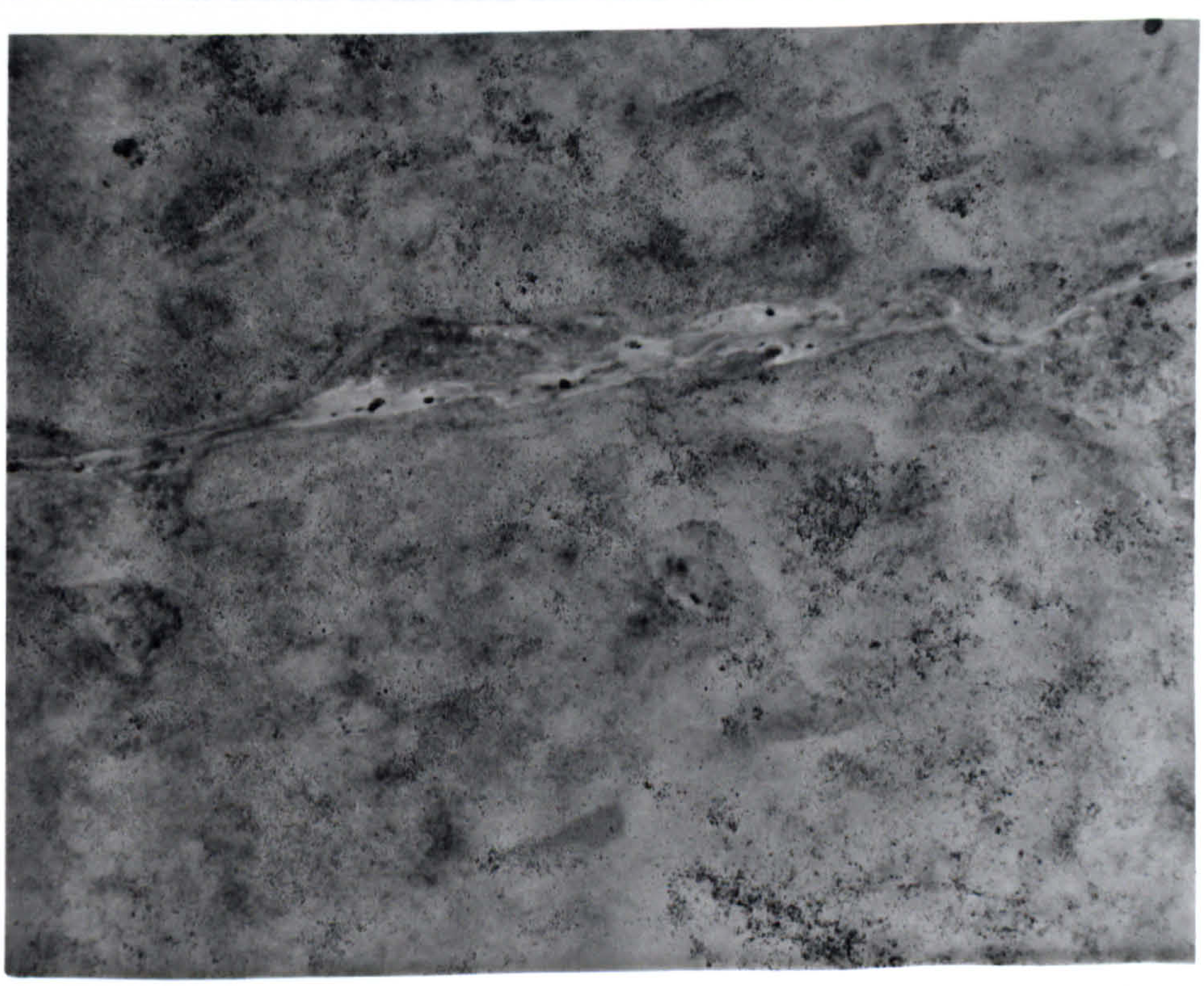


Plate 5.23: The Rooibrak Valley. Photo centred approximately 11 km west of the easternmost Ghanzi Farms boundary fence (from aerial photography; scale 1:50,000). Botswana Department of Surveys and Lands photo number Ghanzi 8 (039) dated 10 June 1987.

The Kalatraverse project studied the Deception Valley in the least distinct sections of its course, suggesting that it was an interdune *straat* (Coates *et al.*, 1979). This report also noted that a section of the Deception which shows a sharp northward deflection is associated with the presence of a major lineament (see section 7.2.2). Cooke and Verstappen (1984, p.8) indicate that both the Deception and Passarge valleys "cease to be distinct features" approximately 40 km west of the Gidikwe Ridge in an area of possible lagoonal deposits, with only "ghost traces" visible across the former lagoon.

The studies by Union Carbide concentrated mainly upon drilling sites in the Rooibrak Valley (although some drilling also took place in the Deception and Passarge valleys), investigating a uranium deposit concentrated in valley calcretes. These studies note that the Passarge consists of a series of curving lobes in a 1 km wide valley, with lobes at approximately 2 km intervals (Union Carbide, 1980*b*). The results of shallow drilling in the Rooibrak Valley are considered more fully in section 6.2.2 and in figure 6.6. It is, however, appropriate at this point to note that drilling proved calcrete deposits to be lenticular in cross-section, and thickest at the centre of the valley (Union Carbide, 1979*b*). The study also showed that lateral variations in valley width were associated with variations in the depth of the sub-Kalahari Basement, although no attempt was made to explain this observation. A wider valley was generally associated with greater depth to Basement, whilst Basement highs were associated with a narrow valley form (Union Carbide, 1979*a*).

(b) Field studies and aerial photography

Only the Rooibrak was studied in the field and followed for a distance of 35 km during the 1990 field season (figure 5.9). The valley is reached by taking the track south from D'Kar village (38 km northeast from Ghanzi) for 8 km to De Graaf farm, and then turning east.

The westernmost end of the Rooibrak was encountered approximately 42 km along the track from De Graaf farm, a distance of approximately 31 km due east from the farm. The valley consists of a series of small interconnected pans at this point with a low relative relief and localised patches of powdery calcrete. The Rooibrak becomes progressively more clearly defined to the east. Where the valley is crossed by the easternmost boundary fence of the Ghanzi Farms it is a very flat feature approximately 300-400 m wide with occasional small pans on the valley floor. The valley is bounded by low banks which reach a maximum height of 1 to 1.3 m and consist of poorly consolidated powdery calcrete (plate 5.22).

The colouration of the valley floor sediment is the only clear indicator of the presence of the valley, the sediment being composed of darker vertisolic soils which contrast markedly with the surrounding pale yellow to grey Kalahari Sands. This contrast can be clearly seen on aerial photographs (plate 5.23) where the photo-tone and texture of the valley differs from the surrounding landscape; the almost imperceptible relative relief of the Rooibrak makes field location more problematic. The old road from Xade to Ghanzi meets the valley approximately 3 km east of the Ghanzi Farms boundary. The Rooibrak was studied for 25 km east of this point, to within 20 km of the Central Kalahari Game Reserve boundary. Two boreholes were encountered along this section, 5804 at 5.8 km east of the Xade-Ghanzi road and 5290 at 18.8 km east. Chippings of spoil material around both boreholes consisted of purple coloured quartzite or sandstone; no traces of duricrust were found in the spoil.

Variations in the form of the Rooibrak confirm the observations of Union Carbide (1980*b*), with the valley width varying almost systematically. Valley floor width ranged between 450-500 m and 100 m wide, with narrow stretches dominating. This conforms to the suggestion of a lobate valley form, with the wider sections commonly containing pans. Very few exposures of indurated calcrete were encountered in this section of the Rooibrak. As noted further west, any exposures that did occur were of rubbly and/or powdery calcrete in the valley flanks. Powdery calcrete was seen in the valley floor where exhumed around animal burrows. Although the Rooibrak was not followed into the Central Kalahari Game Reserve, discussion with local farmers confirmed that, whilst relatively well-defined in the study section, the valley was almost untraceable within the western parts of the reserve. The valley becomes more clearly defined further east where, as the Passarge Valley, it can be followed almost to the Makgadikgadi Depression. The form of the Deception was considered by these farmers to be very similar; almost imperceptible in its western extremities but more clearly recognisable to the east, particularly in the vicinity of Deception Pan.

5.2.4 Northern valley systems

(a) Previous studies

The Kalahari valley systems termed "Northern" in this study include all the possible former tributaries to the Okavango River and Delta region (Moore, 1988) from northeastern Namibia and Ngamiland in Botswana (figures 5.10 and 5.11). Considerable confusion exists in the nomenclature of these valleys, compounded by the fact that valleys cross an international boundary and that many minor interdune tributaries (termed *straats* or *melapo*) are given local names which are often duplicated. The names of the major valley systems referred to by Thomas and Shaw (1991*a*) are taken as standard, the most northerly valley being termed the Ncamasere, with the Xaudum, Qangwadum, Eiseb, Epukiro and Groot Laagte to the south of this.

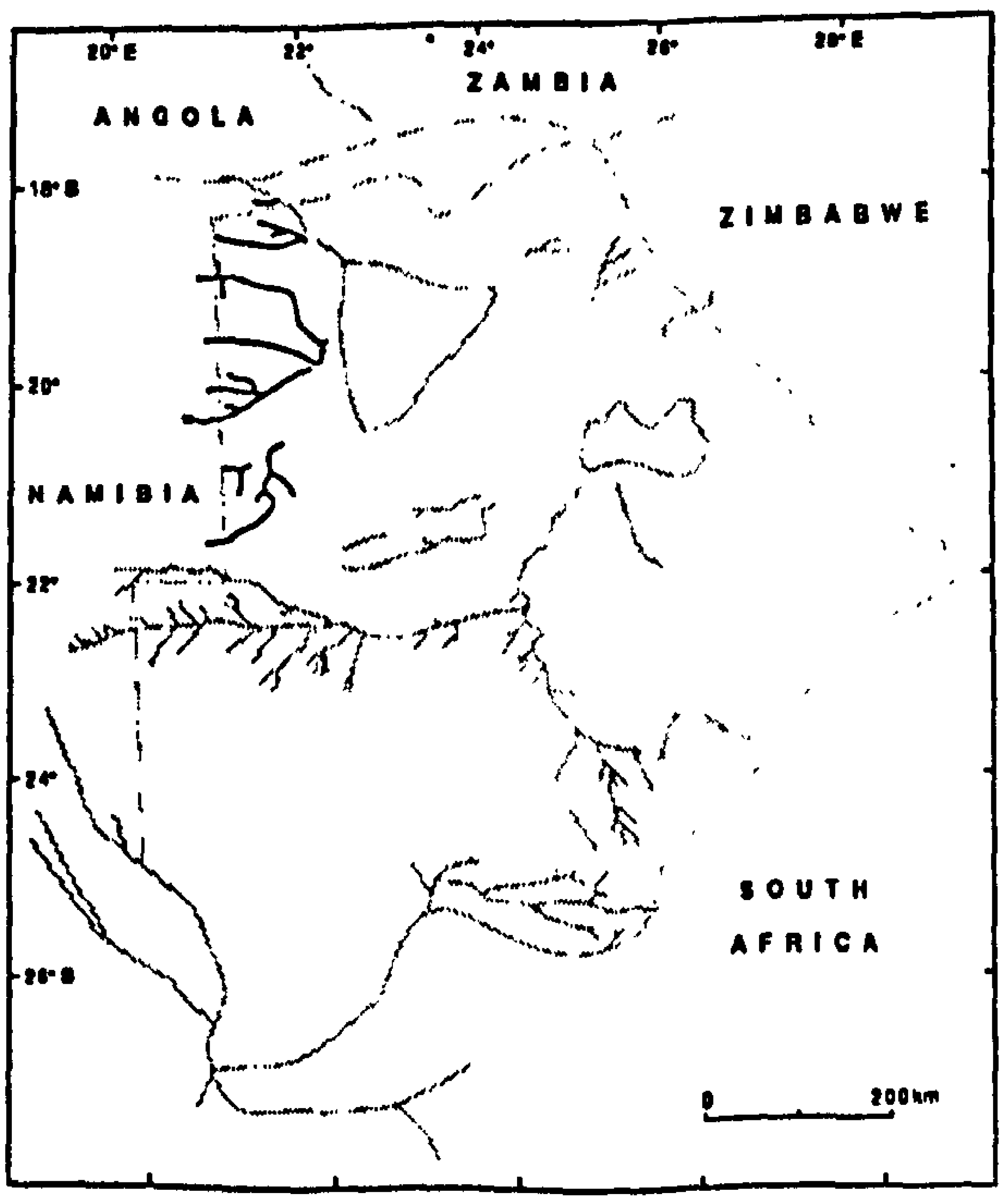


Figure 5.10: Location of the Northern valley systems.

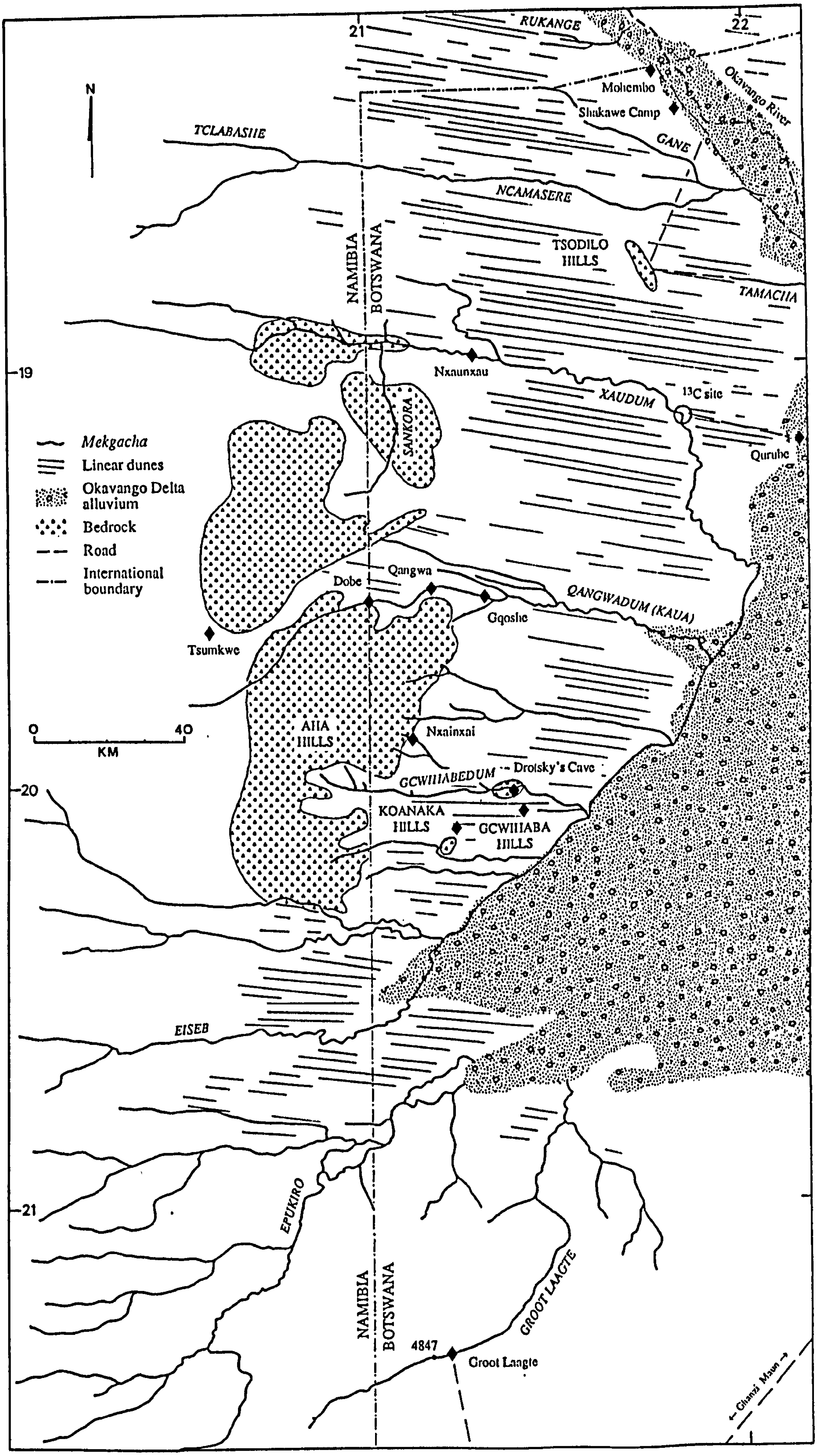


Figure 5.11: The valleys of the Northern valley system.

The earliest reference to any of these valleys was by Passarge (1899 p.312), who crossed the Groot Laagte at a point where its valley was 3 km wide and its channel 400 m wide in places, noting that it "...must once have carried a body of water comparable to that of the Okavango of the present day". Schwarz (1920 p.139) also refers to the combined Epukiro and Eiseb valley as "the great river from the west".

The valleys of the northern Kalahari have been discussed in a geomorphological context by a number of authors, and also in association with geological studies (e.g. Albat, 1978; Union Carbide, 1980d; Hegenberger, 1982; Lüdtke, 1986). The form of the Nxaunxau (Xaudum) is described by Jones (1962), who noted that valley tributaries are usually interdune *straats*. Jones (1962), Grove (1969) and Wright (1978) note that the Xaudum is also joined by tributaries which cut across the local linear dunefields, indicating that either the valley is younger than the dunefield or that flow has occurred since a period of dune-building activity. The Eiseb Valley similarly cuts through linear dunes (Yellen and Lee, 1976). Wright (1978) further suggests the possibility that the Xaudum has incised its course along an older valley trend, and also cites evidence for the effects of regional downwarping and uplift associated with the Gomare Fault. This suggestion arises from anomalous reverse drainage patterns present in parts of the Groot Laagte and variations in valley form exhibited by the Qangwadum. As a result of shallow drilling in three valley systems, Union Carbide (1980*d*) suggest that the Groot Laagte is the oldest of the Northern valleys, citing the fact that the Xaudum and Ncamasere both cut through the linear dunefield, whilst the Groot Laagte apparently does not. This may, however, merely be due to the absence of a well-defined linear dunefield in the vicinity of the Groot Laagte.

Other studies have been undertaken in the Gcwihabe Valley by Cooke (1975) and in the Dobe Valley by Helgren (1978) and Helgren and Brooks (1983). Cooke (1975) uses the evidence of cave deposits and the displacement of a series of valley calcretes by faulting to construct a sequence of environmental changes in the vicinity of the Gcwihabe Hills. Helgren and Brooks (1983) studied sequences of sediments and calcretes in the valleys draining from the Aha Hills on the Namibia/Botswana border. They, and Yellen and Lee (1976), note that most of these drainages emanate from springlines at an unconformity between schists and limestones in the Aha Hills, and also from springs due to groundwater ponding behind Karoo dolerite dykes. In addition to studies in the Dobe Valley, alluvial terraces and fills are also recorded in the Qangwadum.

Finally, a number of references to the form of the Qangwadum have been made during the course of anthropological studies of San Bushmen by Yellen and Lee (1976) and Lee (1979, 1984). All of these studies note that the Qangwadum is incised along parts of its course between the areas of Qangwa and Gqoshe, with a depth of incision of 2m at Qangwa, 4m at Qangwametse and 7-8 m at Gqoshe. In addition, Yellen and Lee (1984) note that the Eiseb, Qangwadum and XaiXai (Gcwihabedum) all have sources of shallow groundwater, although the main water sources in the area are large pans within the valleys.

(b) Field studies and aerial photography

Reconnaissance level surveys were undertaken in the Ncamasere, Xaudum and Groot Laagte systems during 1990. Studies were concentrated in more accessible parts of the systems, with mechanical difficulties precluding work in more remote areas; the results of these surveys will be dealt with in turn.

(i) The Ncamasere Valley

The Ncamasere was studied near its easternmost end; where it joins the Okavango floodplain and where the road from Shakawe Fishing Camp to Tsodilo Hills crosses the valley (see figure 5.11). The Ncamasere rises in Namibia, where it is termed the Kaudom and has the Tclabashe as its main tributary, and can be traced for over 65 km west of the Namibia/Botswana border. The valley continues a further 110 km from the border to the Okavango floodplain. Aerial photography of the valley within Botswana was analysed in Gaborone, although photos of the headwater sections of the valley within Namibia (held in Windhoek) were not studied.

Where the Ncamasere enters Botswana it has a well-defined "floodplain" approximately 45-50 m wide, which gently meanders within a 0.5 to 0.8 km wide west-east trending valley. The terrain surrounding the valley shows no evidence of any surface lineations, suggesting that the Ncamasere Valley has developed within the Kalahari Group sediments at this point. This overall form is maintained until 25 km east of the Namibia/Botswana border (18°33'S 21°16'E), when the channel form abruptly changes to a gently anastomosing pattern (Nash *et al.*, 1993). This consists of a 1 km wide zone of multiple channels located within a broad valley, continuing for 19 km along the valley (plate 5.24).

The reason for this abrupt change in channel pattern is difficult to assess without field evidence, particularly in the absence of information on any changes in valley slope. Studies of channel pattern (e.g. Schumm and Khan, 1972; Knighton, 1984) suggest that the main factors likely to influence pattern are slope and sediment characteristics, along with the cohesivity of the valley flanks. A shift from a meandering to a braided (or anastomosing, in the case of finer sediments) pattern is likely in an area of locally steeper valley slope and/or where there is an increase in sediment load. Such sedimentary input could have been provided by flash-flooding as no tributary valleys join the Ncamasere immediately up-valley of this point. Supplies of unconsolidated sediment may have been available for transport as drainage was re-established at arid to humid transitions (after the ideas of Cooke and Verstappen, 1984). However, without field evidence this suggestion can only be hypothetical. Variations in bank cohesivity may be important, removing lateral constraints to channel location, although no evidence of changes in phototone occur which would indicate locally outcropping bedrock or indurated duricrusts.

Immediately east of this area the definition of the valley decreases on aerial photography, with the disappearance of a clearly identifiable floodplain. It would appear that any flow which may periodically occupy the Ncamasere only extends as far as the easternmost end of the anastomosing channel area (i.e. approximately 38 km east of the Namibia/Botswana border). Further east the valley does not appear to increase in width or depth significantly, but the apparent lack of increase in depth may be due to infill by windblown material.

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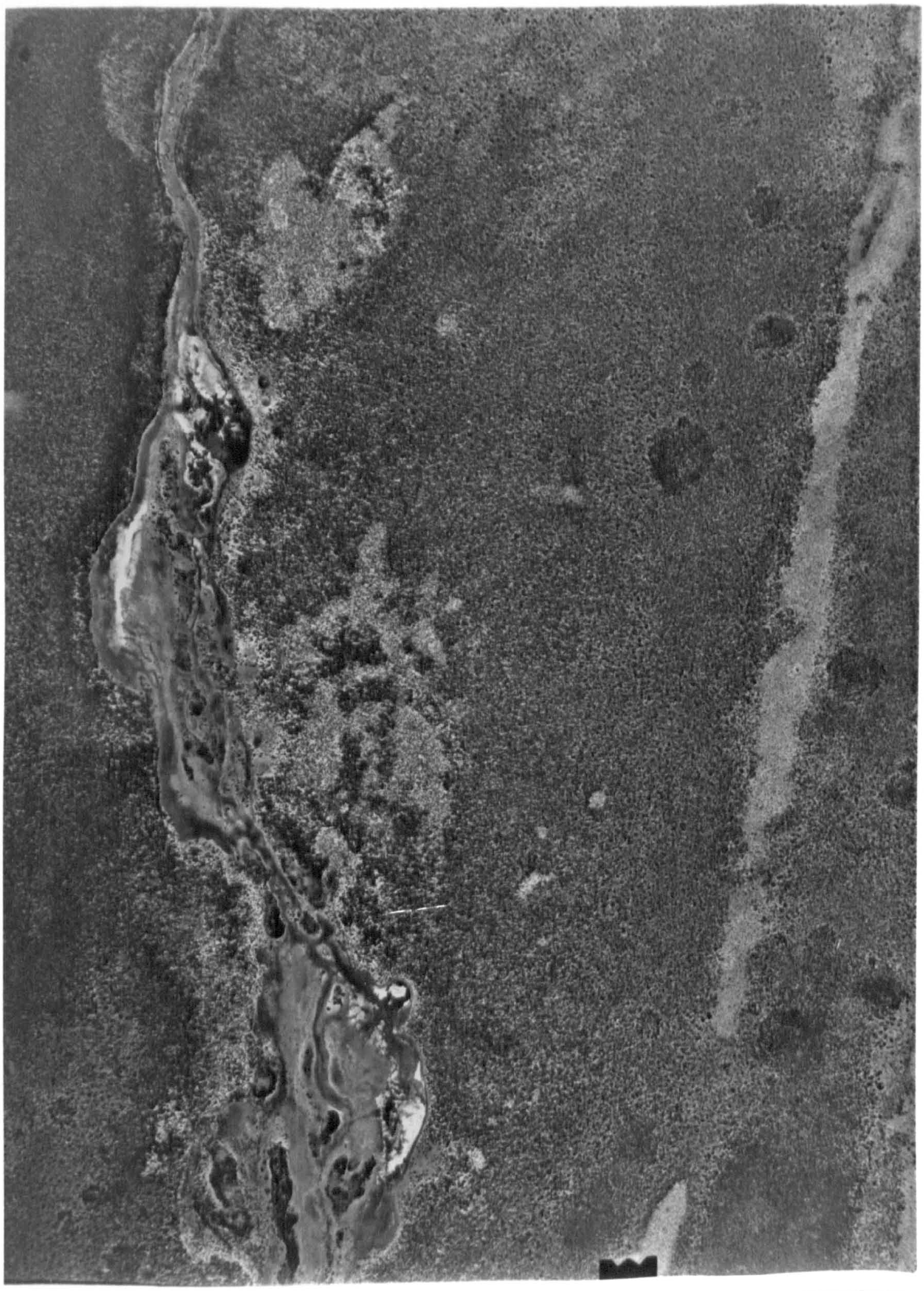

Plate 5.24: Anastomosing channels in the Ncamasere Valley (from aerial photography; scale 1:40,000). Botswana Department of Surveys and Lands photo number 1/80 Ng.W. 4 (128) dated 1 June 1980.

Plate 5.25: The Ncamasere Valley where it is crossed by the track to Tsodilo Hills, looking southwest.

The photographic evidence for sand infill within the main and tributary valleys of the Ncamasere was confirmed by field studies. Three valleys were crossed on the track from Shakawe Fishing Camp to Tsodilo Hills, at distances of 8.1 km (the Gane Valley), 13.5 km (a tributary to the Gane) and 19.5 km (the Ncamasere) from the main Mohembo-Sehithwa road (see figure 5.11). The track cuts the long axis of the extensive linear dune field in this area (Thomas, 1984*a,b*) almost at right angles, and therefore crosses both interdune *straats* and *mekgacha*. Dune crests typically support a higher tree growth than the grassy interdune hollows and *mekgacha*. The valleys are, in the case of the Ncamasere (plate 5.25) and Gane, barely distinguishable from interdune hollows, both having a total relief of approximately 1.5 metres, widths of less than 100m and being infilled by dune sand. The Gane tributary was slightly deeper, with a relative relief of 3.5m, and was 75m wide. The floors of these valleys are typically composed of light buff coloured sand, which contrasts with the surrounding red dune sand. No duricrusts were exposed in the valley floor or flanks. The Gane tributary is the least extensive of the three valleys investigated (figure 5.11) and yet has the greatest total depth. This is because it follows an interdune *straat* with higher sand dunes flanking the valley.

The area where the Ncamasere Valley meets the Okavango floodplain (plate 5.26) was most extensively investigated. Where the Mohembo-Sehithwa road crosses the valley (approximately 18°34'45"S 21°59'15"E) it is 850 m wide with a floor consisting of relatively coarse white sand. The valley at this point contains two distinct channels at the northern and southern sides of the valley floor. These locally developed channels, are separated by an area of chaotic hummocky terrain with a relative relief of less than 2.0 m. There is a large area of partially vegetated sand within the valley due to the selective removal of vegetation by grazing animals (plate 5.26). To the west of the main road the two channels disappear, with no distinct channel present and the valley fill consisting of dune sand.

Plate 5.26: The Ncamasere Valley in the vicinity of the Mohembo-Sehithwa road (from aerial photography; scale 1:40,000). Botswana Department of Surveys and Lands photo number 1/80 Ng.W.4 (148) dated 1 June 1980.

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To the east of the Mohembo-Sehithwa road the two channels merge within 1 km of the road causeway to form a single flat-floored valley. This section of the Ncamasere is affected by the annual flood in the Okavango River, and can be considered as sedimentologically continuous with the Okavango floodplain which it joins some 7.5 km east of the causeway. The Okavango floodwaters regularly backflood to the west of the road (personal communication, owner of Shakawe Fishing Camp, 3rd July 1990), resulting in a transition zone between the sandy floor of the Ncamasere *mekgacha* and the floodplain area. In this transition zone the valley floor is relatively flat, supports a tree and shrub vegetation and contains a number of pan depressions. Further east the valley floor is around 250 m wide with few trees, a cover of grass and rushes and a dark grey organic-rich soil. At the time of field study (July 1990) the Okavango floodwaters had largely receded, but a continuous shallow water body was still present along the valley axis. The presence of abandoned *mokoro* canoes in isolated pools indicates that at the time of maximum flood the valley must be fully connected with the Okavango.

The flanks of the Ncamasere in the floodplain section show an abrupt change from sand to the alluvial valley floor. The sequence from the valley floor consists initially of white sand which forms banks at slopes of up to 15° but more typically 7° or 8°. These white sand areas gradually give way to redder dune sands typical of the sand comprising the linear dunefield further west. The relationship between the valley flanks and sedimentary infill of the Ncamasere in this region indicates that since the cessation of flow within the valley it has been increasingly dominated by Okavango backflooding. The sedimentary infill has backed-up to within 1 km of the road causeway, and may encroach further as the Okavango region subsides and back-flooding causes the lagoonal environment created by the valley to fill with sediment.

(ii) The Xaudum

The Xaudum rises in Namibia, where it is termed the Nhoma on topographic maps, and can be traced for over 80 km before reaching the Namibia/Botswana border. The valley has numerous parallel headwater tributaries developed in interdune *straats*. As with the Ncamasere, only aerial photography of the valley system within Botswana was studied. The Xaudum at the international boundary has a well defined 25 m wide channel (plate 5.27). The valley in this area cuts across the regional NNW-SSE structural "grain" of out- and sub-cropping Nosib Group quartzites of the Damara Sequence (Hegenberger, 1982; Balfour *et al.*, 1985). The main valley runs west-east on plate 5.27, with a southern tributary (the Sankora) entering at 5 km east of the border. The orientation of this tributary valley appears to be controlled by structures within the Nosib quartzites. The Sankora can first be seen on air photos around 9 km south of the Xaudum and is of potential palaeoenvironmental interest as its course cuts through the extensive linear dune field in this area at almost 90°. Tributaries to the valley are developed within interdune *straats* and join the valley from both west and east. This suggests that there has been flow within the Sankora since the formation of the now vegetated linear dunes. Alternatively, if the valley were incised into the quartzites prior to the last major period of dune activity this may explain its persistence. However, without detailed field investigation this suggestion cannot be substantiated.

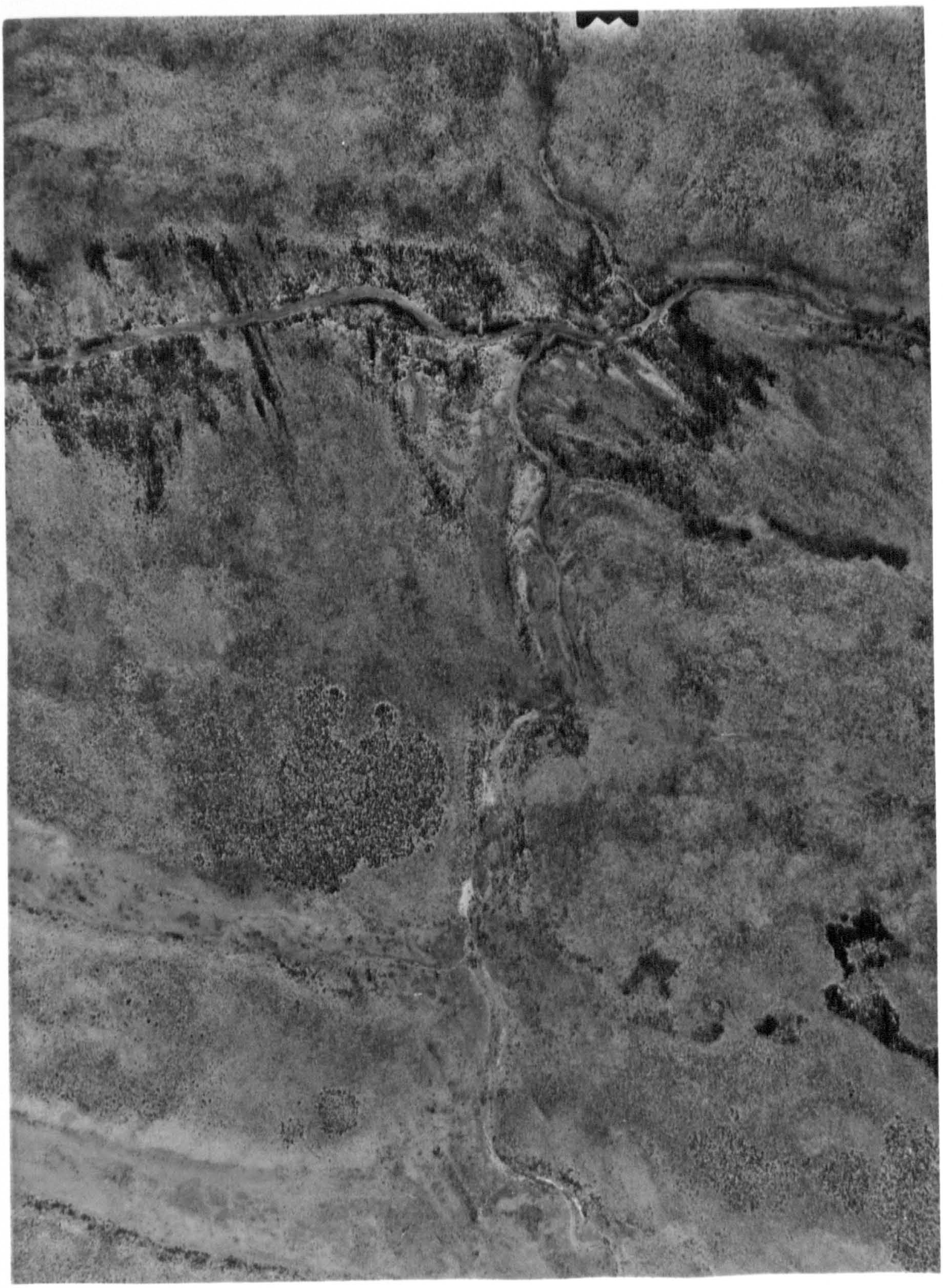


Plate 5.27: The Xaudum near the Namibia/Botswana border (from aerial photography; scale 1:40,000). Botswana Department of Surveys and Lands photo number 1/80 Ng.W. 36 (147 dated 6 June 1980.

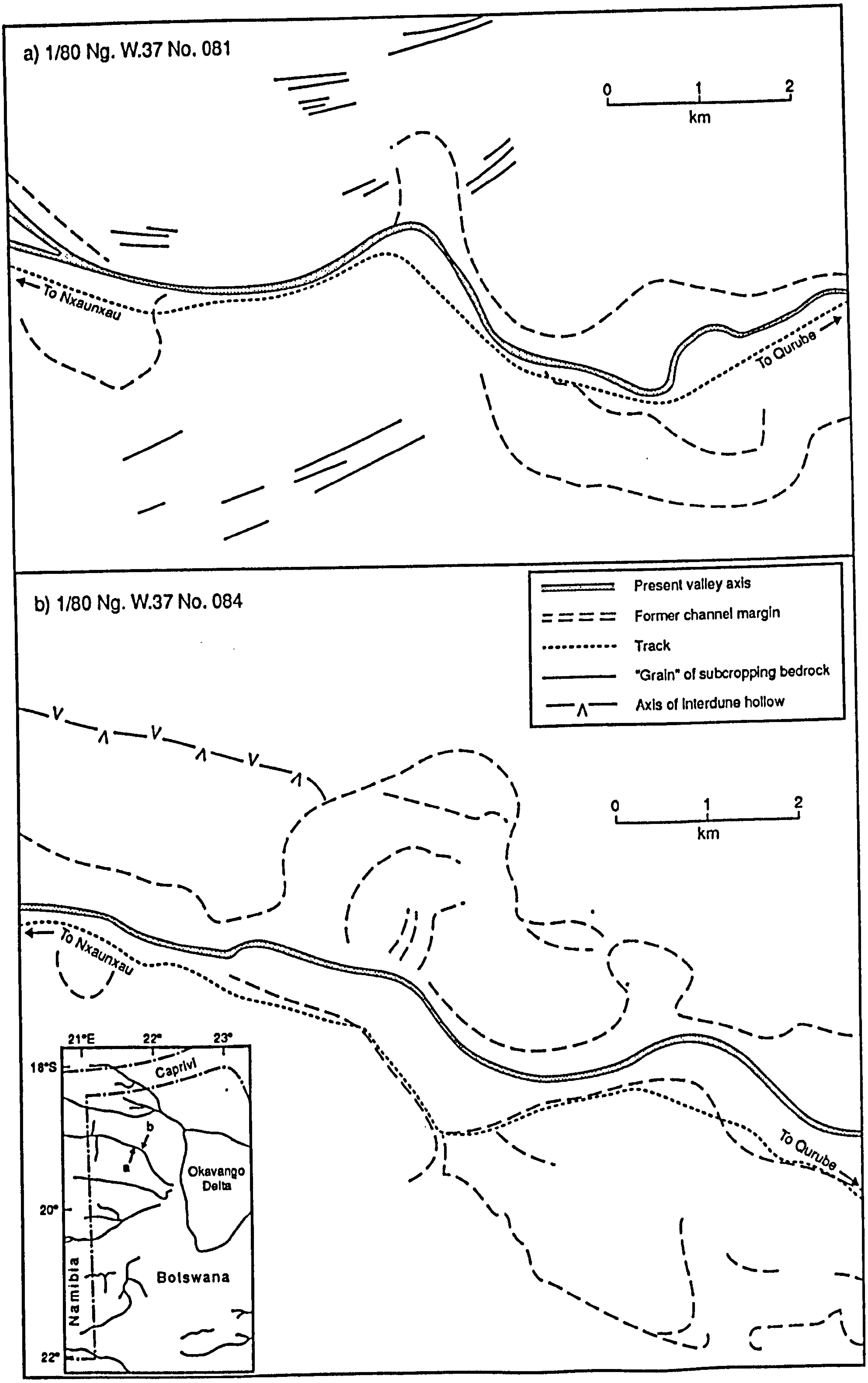


Figure 5.12: Abandoned meander channels in the Xaudum; a) at 19°03'S 21°38'E and b) at 19°07'S 21°44'E.

As in the Ncamasere, the course of the Xaudum becomes less clearly defined to the east, and commonly consists of a number of intermittent pans within the valley. There are parts which contain abandoned meander channels at 19°03'S 21°38'E and 19°07'S 21°44'E (respectively, 71 km and 79 km down-valley from the Namibia/Botswana border; figure 5.12). In these areas, the present "channel" is usually distinguished on aerial photography by a lighter tone, with abandoned meanders exhibiting a more chaotic texture and denser vegetation cover. These meander channels appear to be developed within a much wider valley at this point (Nash *et al.*, 1993). However, as with the Ncamasere, it is difficult to assess whether the meandering has eroded the valley or has simply developed within a wider pre-existing valley.

No abandoned meanders can be identified from aerial photography further to the east. Instead, the entire valley meanders, with the channel apparently restricted to the base of this incised valley. At around 19°04'S 21°45'E, the course swings southeastwards and cuts across the direction of the linear dunefield. West of this point the Xaudum is aligned approximately parallel to the long axis direction of the dunefield i.e. at approximately N105°E. It was in this section of the valley that field studies were carried out during 1990, with an aim to investigate the potential palaeoenvironmental significance of the valley/dune relationships.

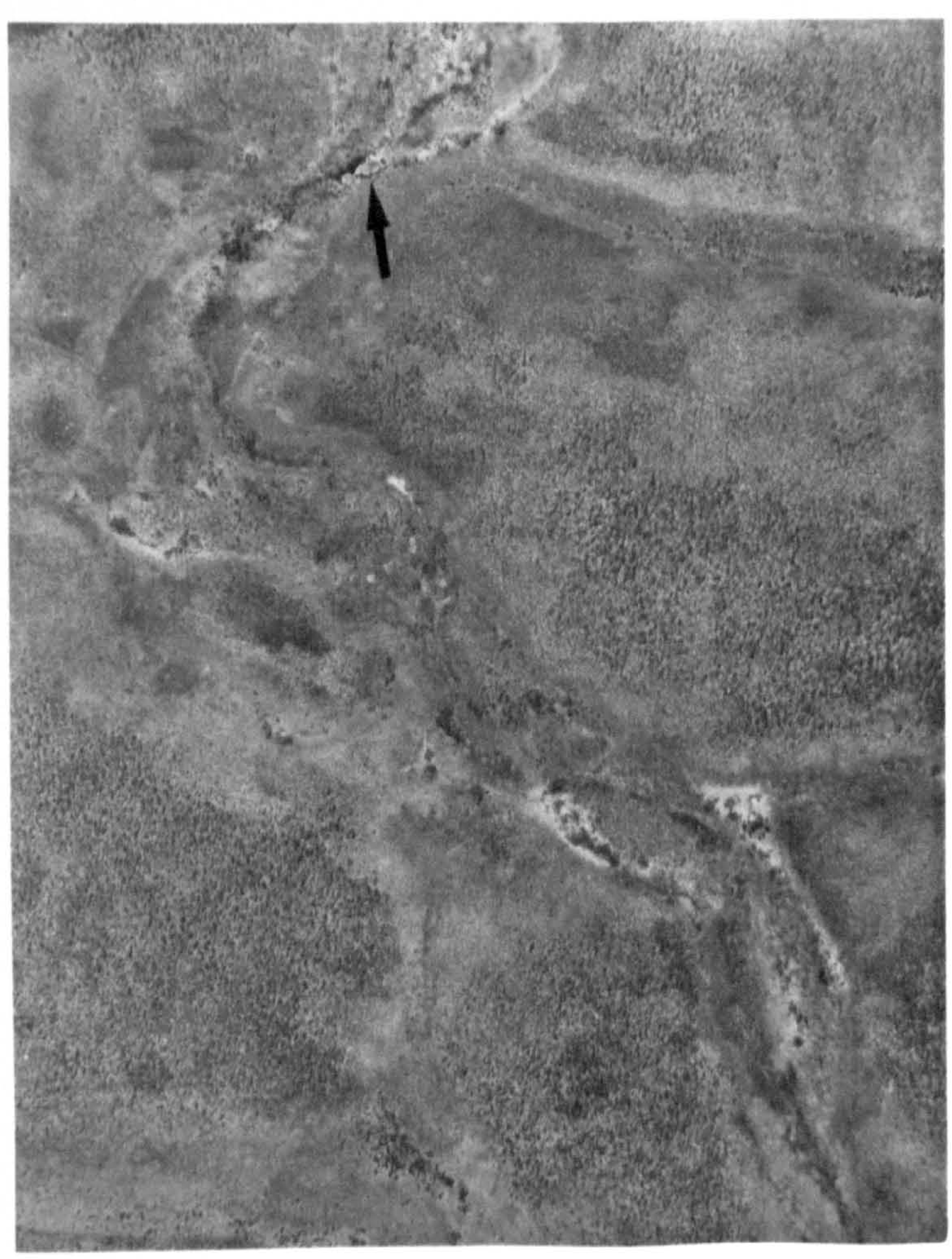


Plate 5.28: The Xaudum, indicating the radiocarbon dated calcrete sample location (from aerial photography; scale 1:40,000). Botswana Department of Surveys and Lands photo number 1/80 Ng.W. 38 (200) dated 7 June 1980.

Owing to technical problems, the Xaudum was only studied in the field where the road from Qurube to Nxaunxau first intersects the valley (figure 5.11; plate 5.28). This site is located in a meandering section, where the course of the valley cuts through the dune field and tributary valleys are commonly developed in interdune *straats*. The Xaudum has a total depth of incision of approximately 8-10 m, a valley width of around 500 m and slopes of 4° on average. Slopes can be zonated into a concave lower section, a rectilinear slope at angles of up to 8° with a convex upper zone. The valley floor is flat and contains dark sand-rich soil with a relatively high organic matter content; this contrasts with the flanks which typically consist of buff to red Kalahari Sand. Pans also line the valley bed, which shows no morphological evidence for former flow. Although the Xaudum cuts through the well vegetated linear dunefield in this area there is no evidence for any extension of the dunes across the valley, as seen elsewhere (e.g. the Auob Valley; section 5.3.1). This suggests that any flow which may have occurred in the valley post-dates the last period of linear dune extension.

Calcrete exposures are common on the valley flanks, with a continuous low terrace level on both flanks at a height of 0.8 to 1.0 m above the valley floor (Shaw, Thomas and Nash, 1993). The terrace was identifiable along the entire valley section studied, and comprises $CaCO_3$ cemented sand, alluvial silt and shell material. The calcrete appears to have developed within the confines of the valley and sedimentologically represents a cemented former valley floor. The shell material is dominated by well preserved *Lymnaea natalensis*, a species indicative of near-permanent still water habitats (Brown, 1980). This habitat may have been provided by pan depressions within the valley. A radiocarbon date (GrN 18071) for a combined calcrete and shell sample (at 19°08'06"S 21°49'47"E from the west side of the Xaudum adjacent to where the Qurube to Nxaunxau road crosses the valley, at a height of 0.5 m above the valley floor), yielded an age of 14,570 ± 160 years BP (Shaw, Thomas and Nash, 1993). As with the radiometrically dated samples from the Okwa Valley (discussed in section 5.2.1 above) problems of potential cross-contamination of inorganic and organic carbonates are likely to occur. The date should therefore be regarded only as a minimum age for the shell deposit which is probably 15-25% older than the combined shell/matrix date would suggest (Shaw and Thomas, 1992). This would make the age of the shell material largely consistent with other dates from the Middle Kalahari.

Studies of aerial photography indicate that the Xaudum becomes increasingly poorly defined down its course. This may be partly due to the absence of a major track along the valley to the south of the Qurube to Nxaunxau road, and hence disturbance around settlements along such a track (which assists identification of valley floor areas). Where the valley approaches its confluence with the Gcwihabedum it has a valley floor width approaching 400 m. East of this point the valley diverges into a number of distributaries, possibly indicative of a former deltaic zone, and subsequently disappears into an area of chaotic terrain, ultimately merging with the Okavango sediments at around 19°31'S 22°11'E.

(iii) The Groot Laagte

The Groot Laagte was studied over a 5 km section in the vicinity of the Groot Laagte BaSarwa village (figure 5.11) during 1990. This village can be reached by taking the track signposted "Groot Laagte" 42 km west of Ghanzi; the valley is crossed 51 km north of the Ghanzi-Mamuno road.

The main valley rises within Namibia, approximately 20 km west of the international boundary, although the width of the valley at this point is unknown since aerial photography was unavailable. Where first studied, the valley was comparatively broad and of low relative relief, with a width of 2.5 to 3.0 km, flank slope angles of less than 1° and a maximum depth of no more than 10 m (although the great width made relative relief assessments unreliable).

The main headwater of the Groot Laagte is unusual amongst other northern valley systems as it does not appear to possess any tributaries. On Namibian topographic maps it rises at 21°33'S 20°53'E, and Landsat MSS images held at the UNDP/FAO Soil Survey Unit in Gaborone show no evidence for tributaries extending further southwest than this point. This contrasts markedly with the extensive tributary networks of the Epukiro system to the northwest and the Hanchai to the south and southwest. This apparent difference in the density of tributary networks could be due to a number of reasons. Firstly, the Epukiro and Hanchai rise on areas of sub- or outcropping bedrock whereas the Groot Laagte is in an area of more extensive Kalahari Sand cover; this cover may have buried any tributaries originally present or may not promote low level tributary flows. A second possibility is that the development of the Epukiro and Hanchai systems limited the potential catchment area of the Groot Laagte, which rises in between these other systems.

Field studies in the Groot Laagte covered the section of valley between borehole 4847, located 4.0 km west of the track into Groot Laagte village, and around 1.0 km east of borehole 4846 which is situated in the valley base adjacent to the track. Along this section the main distinguishing feature of the valley base was a light grey vertisolic soil which is visible on aerial photography (plate 5.29). Some poorly exposed hardpan calcrete outcrops in the valley base, but generally duricrust only occurs as powdery calcrete within the valley soils. No clearly defined channel was present within the valley, and no obvious banks were seen (plate 5.30).

Very little variation in valley form was seen along the study section. In the vicinity of Groot Laagte village (see plate 5.30) average slopes into the valley are less than 1°, with the valley floor over 500m wide. One 2.5 m deep well close to borehole 4846 contained soft powdery calcretised valley floor sediments.

To the east of the study section, aerial photographs indicate that, as with many other *mekgacha*, the Groot Laagte becomes less clearly defined. The valley becomes very indistinct at 21°33'S 21°45'E when the last traces of an individual channel can be identified before eventually merging with the Okavango Delta sediments in a 4,600 km^2 area of chaotic terrain, identified by Thomas and Shaw (1991*a*) as deltaic deposits, at 21°33'S 21°37'E.

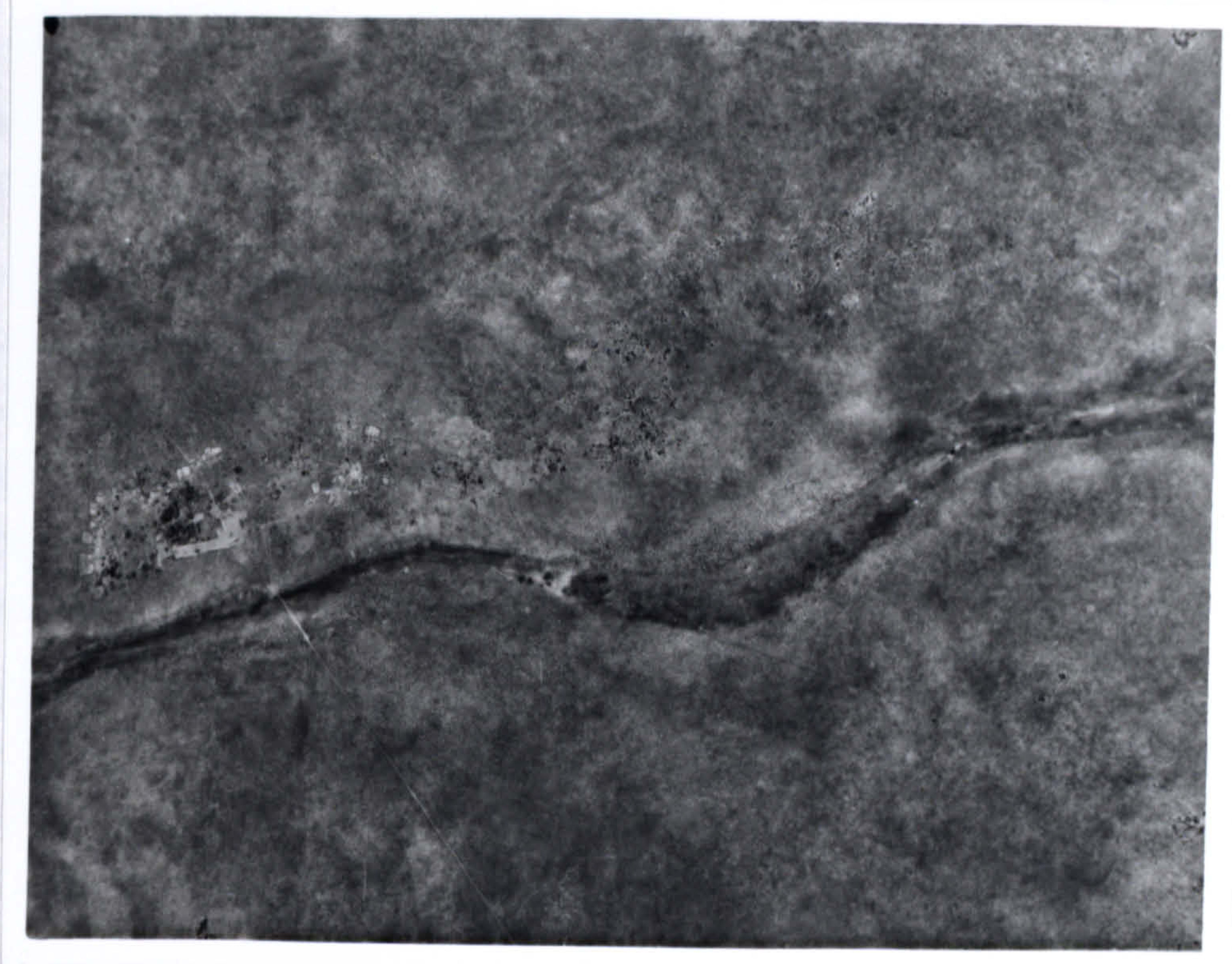


Plate 5.29: The Groot Laagte at Groot Laagte BaSarwa village (from aerial photography; scale 1:50,000). Botswana Department of Surveys and Lands photo number Ghanzi 6 (095) dated 9 June 1987.

Plate 5.30: The Groot Laagte immediately west of Groot Laagte village.

(iv) Other northern valley systems

In addition to the Ncamasere, Xaudum and Groot Laagte, reconnaissance level field work was undertaken in two other valleys; the Rukange in northeastern Namibia and the Tamacha which can be traced due east from Tsodilo Hills (figure 5.11).

The Rukange Valley was only studied in the vicinity of the Mohembo to Bagani road in the western Caprivi Strip. The valley, which forms a tributary to the Okavango River, is the product of the combination of a number of interdune valleys. The valley is a barely noticeable depression with no channel present and a total depth of incision of no more than 2 m. It is only distinguishable by a zone of very dark organic-rich vertisolic soil and reed-like grasses. The width of this zone of darker soils is approximately 300 m, and this is taken as indicative of the valley floor width.

The Tamacha Valley was briefly studied at its eastern end where it is crossed by the main Mohembo-Schithwa road. The valley is barely identifiable, with negligible relative relief, but slightly darker soils occur and numerous fragments of terrazzo silcrete were seen in the vicinity of the road. The southern track to Tsodilo Hills partly follows the valley before crossing sections of the linear dunefield.

The Gcwihabedum was not investigated in the field, but aerial photographs of its headwater sections were studied. These indicate that the valley is very indistinct where it crosses the Namibia/Botswana border, with a channel around 100 m wide. Short tributaries to the Gcwihabedum often end in pans. The Nxainxai Valley, a tributary to the Gcwihabedum, rises to the east of the border at Nxainxai village, with farmlands identifiable within a dambo-like lobate valley head area.

5.3 Exoreic drainage systems

In this section the Molopo, Kuruman, Auob, Nossop and Moselebe valley systems, formerly connected to the Orange River, and the Serorome Valley linked to the Limpopo via the Bonwapitse River are considered. Detailed studies were undertaken in the Kuruman, Auob and Serorome, the remainder being studied at a reconnaissance level.

5.3.1 Auob and Nossop valleys

(a) Previous studies

Both the Elephants (the main tributary to the Auob) and Black and White Nossop Rivers (which join to form the Nossop) rise in the Auas Highlands to the east and northeast of Windhoek in central Namibia. The Black and White Nossop join some 80 km to the south of Gobabis, with the Nossop River crossing the Namibian border 240 km downstream. The valley then forms the Botswana/South Africa border for over 290 km within the Kalahari Gemsbok National Park, merges with the Auob River 4.5 km north of Twee Rivieren Rest Camp and then joins the Molopo at Bokspits, 60 km south of this point. The Auob River runs for 115 km through the Kalahari Gemsbok National Park having entered South Africa 70 km "downstream" of its confluence with the Elephants River (figure 5.13).

The geomorphology of the Auob and Nossop has been comparatively well documented, mainly due to the fact that both valleys pass through a National Park and are therefore relatively accessible; Du Toit (1926*b*) and Weinberg (1975) note that the Auob was a major motoring route between Upington and Windhoek. References to the Nossop date back to the time of Alexander (1838, Vol 2, pp 159-160) who stopped to shoot elephants in the valley. He noted bushes in the stony bed, with numerous holes made by elephants to gain access to shallow groundwater. Similarly, Hodson (1912 p.58) describes the Nossop as being "...very broad and quite dry; in fact, there were a number of trees growing in the middle of it."

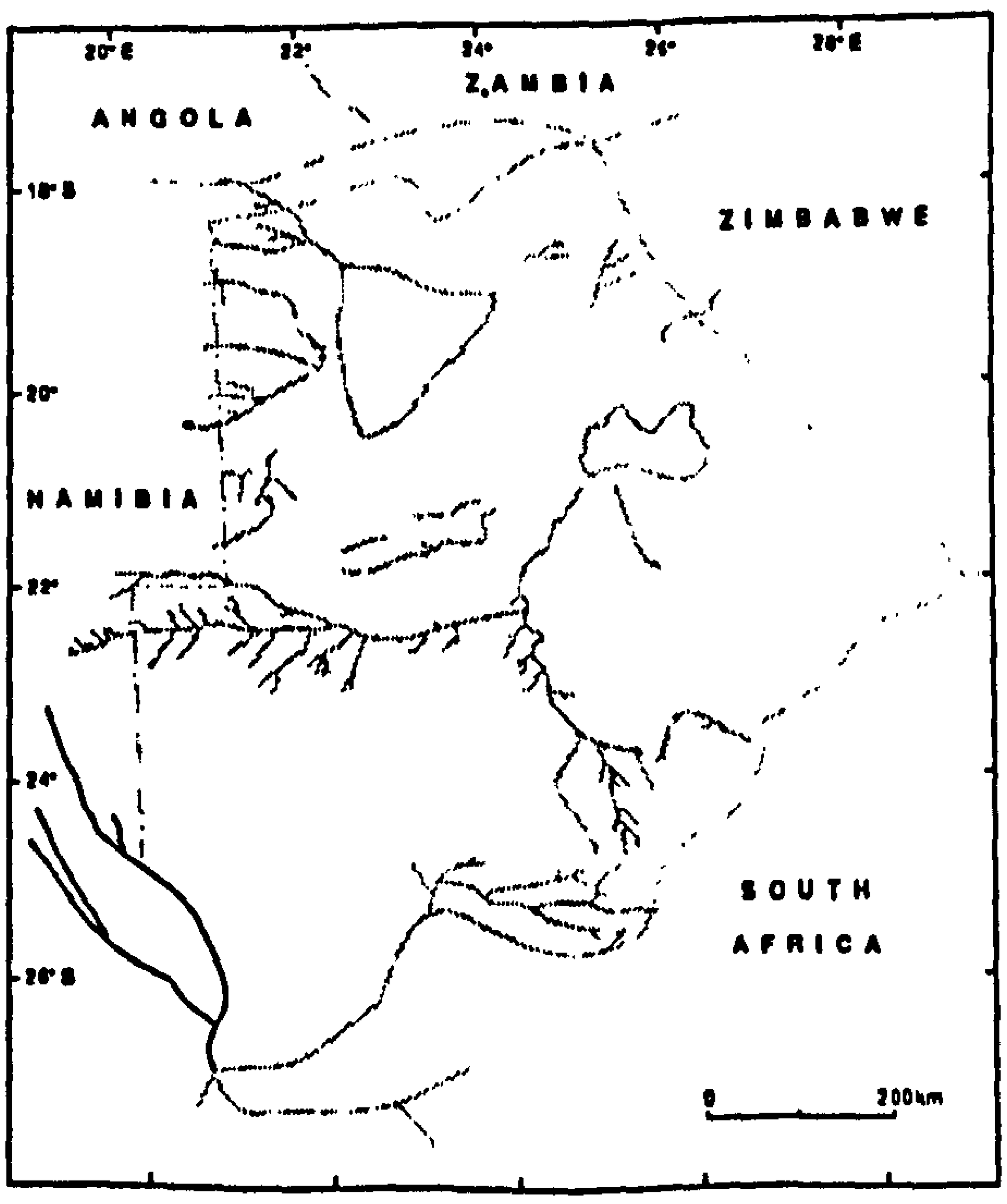


Figure 5.13: The location of the Auob, Nossop and Elephants valley systems.

Andersson (1856) also refers to the Nossop, suggesting that the river was linked to the Orange using evidence of common fish species. Both Baines (1864) and Chapman (1886) crossed the Elephant and Nossop valleys, with Baines (1864 p.69) describing the Elephant as "...three or four times as broad as a turnpike road".

One of the most detailed early references is that of Herbst (1908 pp 207-208) who notes from studies of wells in the floor of the Nossop and Molopo that;

> "Their beds have... been raised considerably, for in those wells that have been sunk therein the white shingle, beautifully smooth and rounded by running water in time past, is found fifty feet below the present surface. Below this shingle... water is found..."

Both the Auob and Nossop are, in places, greatly incised into the surrounding plateau, a fact described by a number of studies including Range (1910, 1912, 1914). The Nossop is noted by Leistner (1967) as being deeply incised between Leonardville and Aranos in Namibia, and alternately wide and shallow from Union's End to Kameelsleep and incised between Kijkij and the Molopo confluence in South Africa. The course of the Nossop also meanders considerably within the Kalahari Gemsbok National Park (Leistner, 1967) with evidence of abandoned channels and an oxbow lake near Grootbrak waterhole (Thomas *et al.*, 1988; Nash *et al.*, 1993). Dry tributary riverbeds also join the Nossop, for example, near to Grootbrak waterhole (Leistner, 1967).

The relationship between the Auob and Nossop valleys and the surrounding topography is discussed by a number of authors. Whilst many studies (e.g. McConnell, 1956; Thomas *et al.*, 1988) assume that the valleys simply incised into a pre-existing plateau, Mabbutt (1955) suggests that both must have existed during at least the latter stages of the formation of the Kalahari Group sediments. He cites the fact that calcrete is commonly thicker along the line of valleys, sometimes merely capping a pre-existing valleyside slope, and that terraces flanking the valleys contain fluvial gravels (also noted by Brain, 1957). This would suggest a period of entrenchment after an earlier period of calcification. The valleys also cut an extensive area of linear dunes; Mabbutt (1955) infers that the valleys must have existed prior to the formation of these dunes, although the presence of small linear dunes on the floor of the Nossop (Lewis, 1936) suggests a comparatively recent minor incursion of dune sand. Whilst Lewis (1936) views valley floor dune development as a past process, dune patterns suggest that sand has crossed the valley in the past and may be able to do so at the present day. Initial sand deposition may also have pre-dated valley formation. Clearly, valleys and dunes have a more complex interrelationship than either Mabbutt or Lewis suggest.

Evidence for the antiquity of headwater sections of the Nossop is provided by Hegenberger and Seeger (1980) whose geological map of the Gobabis area shows Dwyka tillite erratics within the Black Nossop over a distance of some 50 km south of Gobabis. This suggests that a valley has existed along this headwater section of the Nossop course since the Permo-Carboniferous glaciation. Changes in the drainage pattern of the Nossop headwaters are also mentioned by Hegenberger and Seeger (1980) and McDaid (1985). Hegenberger and Seeger (1980) also record three gravel terrace levels in the Black and White Nossop rivers.

Water resources in the Auob and Nossop form a prominent part of the literature, including studies of groundwater nitrate levels in the Stampriet recharge area near the Auob headwaters (Heaton *et al.*, 1983). Possible periods when the Auob and Nossop may have permanently flowed have been suggested by several authors (e.g. Brain, 1957; Heine, 1978*b*). Of these, only Heine (1978*b*) suggests absolute dates, inferring flow during the period 18,000-11,000 BP based on radiocarbon dates from freshwater and terrestrial mollusca. Dates of historical flows are, however, much better documented. Under normal rainfall conditions, flow in the Nossop reaches Aranos (Leistner, 1967) and as far as Gochas in the Auob (Range, 1912). However, years when floodwaters extended beyond these locations in the Auob and Nossop (and also the Kuruman and Molopo rivers) are recorded by a number of sources (e.g. Du Toit, 1926*a*; Kokot, 1948; Clement, 1967; Weinberg, 1975; Verhagen, 1983; National Parks Board, 1986; Molosiwa, 1989). Such floods were relatively infrequent and rarely "torrential", as can be discerned from the memoirs of the period 1926-1939 by Weinberg (1975 p.52). He notes that, having become stuck in mud whilst crossing the bed of the Auob during the rainy season;

> "There was no fear of the river coming down as the Auob never really flows strongly even on those rare occasions when it flows at all".

Floods are known to have occurred in the Nossop in 1806, 1933-34, 1963 and 1987, in the Auob in 1933-34, in the Kuruman in 1891-92, 1894, 1896, 1915, 1917-18, 1920, 1933-34, 1974-77 and 1988-89, and in the Molopo in 1891-92, 1896, 1933-34, 1915, 1917-18 and 1988. The 1933-34 flood was unique in including all four rivers, with the flood in the Nossop described by Clement as being 450 feet wide and travelling at 6 mph. Somewhat contradicting his earlier comments, Weinberg (1975 p.100) graphically depicts the effects of the floods in Namibia;

> "...after a long drought, the rains came suddenly with fury. The dry rivers were converted into raging torrents overnight. The hard, dry, flat sun-baked pans that could be, and actually were, used for car-racing in some places became huge lakes...".

The effect of the flooding is reflected in sedimentary deposits in a number of locations. The National Parks Board (1986), in particular, note long linear ridges of silt deposited by the 1933-34 flood in the bed of the Nossop at Grootbrak and Kaspersdraai waterholes.

(b) Field studies and aerial photography

(i) The Auob Valley

The Auob was studied during both 1989 and 1990, with investigations concentrating upon the valley between Stampriet in Namibia and the confluence with the Nossop Valley in the Kalahari Gemsbok National Park. Duricrust exposures are described in section 6.2.1. This section considers variations in valley form in two parts; firstly, the form of the 250 km of valley between Stampriet and Mata Mata Camp (at the Namibia/South Africa international boundary), followed by a brief consideration of the section within the Gemsbok National Park.

The Auob between Stampriet and Mata Mata Camp

The Auob rises to the north of Stampriet (24°20'S 18°24'E) in an area of shallow subcropping bedrock identifiable from remotely-sensed imagery. A number of small tributaries combine to form the main valley in the vicinity of the town. The valley becomes most clearly defined beyond where it swings southeastwards 2 km east of Stampriet.

The most notable feature of the morphology of the Auob in this 250 km section is its consistent depth. The valley maintains a relative relief of between 22 and 25 m beneath the plateau of Kalahari Group duricrusts. The valley width, however, varies considerably, giving a false impression of increased incision in places. The form varies between gorge-like sections and gently sloping convex/concave valley flanks. Immediately down-valley of Stampriet the floor of the Auob consists of dark, organic-rich (due to cultivation) alluvial soils and is approximately 500 m wide in a 1.8 km wide valley. The valley flanks are gently sloping, with average slope angles of between 2° and 5° and a "well-rounded" valley landscape which contrasts markedly with the valley seen further south. No major duricrust outcrops are present, the valley flanks consisting of sand and duricrust rubble.

The valley gradually narrows to the south of this point, with an associated increase in the extent of duricrust cliffs at the top of the valley flanks. Rectilinear duricrust debris slopes occur below the vertical cliffs, with an average angle of between 25° and 28° tailing off to a basal concavity; at Witkrans, this debris slope is 8 m high vertically. The appearance of duricrust cliffs at Witkrans is concomitant with extensive dissection of the edge of the plateau surface by small dendritic tributary valleys. This dissected landscape has the appearance of a "badlands" topography, which is initially best developed on the western valley flank. However, within 2 km of the onset of the dendritic tributary drainage pattern, both flanks have a similar appearance (plate 5.31). The typical valley form, as seen at Kalkheuval (24°45'45"S 18°44'25"E) where extensive duricrust sampling was undertaken, is shown in plate 5.32.

This "badlands" landscape appears to be associated with extremely indurated upper surface horizons of the duricrusts forming the plateau. These act as a cap-rock, which, once eroded allow extensive removal of softer underlying material and the development of the dendritic gully and channel pattern. The more indurated duricrusts remain as a low cliff with a rectilinear transport slope beneath the free face covering the softer material.

The best examples of the "badlands" topography occur to the north of Gochas town (24°51'30"S 18°48'30"E) where extensive dissection of the valley flanks has led to the development of inselberg features (plate 5.33). In places (e.g. Simon Koper Farm, 24°50'S 18°47'E) the valley flank slope form is complicated by a localised lower outcrop of indurated sil-calcrete at a height of approximately 7 m above the valley floor which creates a false terrace level within the valley. This lower indurated siliceous duricrust horizon only occurs over 4 km of the eastern valley side. The "badlands" occur as far south as Minneplaats Farm (25°10'S 18°59'E), with the valley flanks less extensively dissected to the south. The valley also narrows to around 500 m at Minneplaats, with a 175 m wide valley floor.

Plate 5.31: The Auob Valley at Kalkheuval (from aerial photography; scale 1:50,000). Namibia Department of Surveys and Lands photo number 8213, run 10 (747).

Plate 5.32: The Auob Valley at Kalkheuval, looking south.

Plate 5.33: Inselbergs on the eastern flank of the Auob, north of Simon Koper farm (Gochas).

Plate 5.34: The Auob Valley north of Eindpaal Farm, looking south.

To the south of this point, until the Namibia/South Africa border at Mata Mata, the valley form is much simpler. Duricrusts still outcrop extensively at the top of the valley flanks, but they form low bluffs above largely undissected rectilinear debris slopes. The valley is approximately 400 m wide in this area (plate 5.34) and is more entrenched, with incised meanders occurring at Duikerloop (25°15'30"S 19°04'00"E). The valley is clearly incised through an extensive plateau surface, now covered by a Kalahari Sand dune field. For example, the flat duricrust plane can be seen extending up to 800 m away from the valley axis in the vicinity of the Auob-Elephants confluence. Analysis of spot heights immediately adjacent to the Auob and Nossop valleys and in interdune hollows within the dune field confirms this observation. These heights were taken from Namibian 1:50,000 topographic maps and results are not included here. Whilst there is a slight rise in relief between the valleys, probably associated with the thickness of dune sand, the overall plateau surface slopes at a gradient of between 1 and 2 m per kilometre towards the Auob-Nossop confluence.

There is evidence for surface flow within the more northerly parts of the Auob. During the 1990 field season continuous surface water occurred up to 35 km south of Stampriet, with intermittent pools of water observed up until 42 km south of the town. A definite channel occurs in the upper part of the valley, although this has virtually disappeared by as far south as Gochas. As with other valleys, the floor of the Auob is characterised by a greyer sediment compared to the valley flanks.

The relationship between the adjacent dune field and the Auob is of interest both in this section and for the valley within the Kalahari Gemsbok National Park. In the more northerly parts of the Auob, the orientation of the valley and the trend of the linear dunefield are approximately parallel north of the incised meanders at Duikerloop Farm (25°15'30"S 19°04'00"E). This results in little interaction between the dune and valley systems, with a distinct corridor containing no linear dunes extending for 2 km west and east of the Auob (plate 5.31). The reason for this dune-free corridor is difficult to explain. It may be due to a lack of sand, the removal of sand which may have been present in the past or the unsuitability of the plateau surface for sand deposition. Analysis of aerial photography shows that thin sand does cover the plateau adjacent to the valley in places, eventually leading up to a dune ridge, but in most areas the upper duricrust surface is free of sediment. To the south of Duikerloop, the course of the valley swings gradually towards the east before the confluence with the Elephants River. The valley and dune orientations become progressively less aligned until they are almost perpendicular at the confluence. Linear dune sand extends beyond the plateau edge onto the valley flank to the north of Vergenoeg and Eindpaal farms (at 25°20'S 19°08'E and 25°21'S 19°09'E respectively) where linear dunes meet the valley. A serrated valley edge is indicated on aerial photography where dunes extend onto the northern flanks. Where these dunes abut the valley they reach heights of up to 7 m above the plateau surface.

Finally, before considering the form of the Auob within the Kalahari Gemsbok National Park, mention should be made of the form of the Elephants River, the major tributary to the Auob which joins the valley at Twee Rivier Farm (25°27'30"S 19°26'30"E). The valley was only studied at a reconnaissance level in three places; where it is crossed by the Stampriet-Aranos and Gochas-Nossop roads and at the Auob confluence. At the first of these three locations, the valley was little different to an inter-dune corridor, with no channel present. The only distinguishing features between the valley and adjacent inter-dune corridors were slightly darker soil in the valley floor and the presence of some larger shrubs. To the east of Gochas the valley form had changed only slightly, with some duricrust rubble in the valley flanks. At Twee Rivier the form was little different to that of the Auob at the same point, with low duricrust bluffs at the top of rectilinear debris slopes, a depth of incision of approximately 22 m below the plateau surface and a width of around 250 m.

The Auob between Mata Mata Camp and the confluence with the Nossop

The depth of the Auob at Mata Mata (figure 5.14) is little different to that described above i.e. the valley has a relative relief of approximately 23 m below the adjacent plateau surface. The valley floor, however, is significantly wider for a distance of approximately 2 km south of the international boundary, with the channel meandering across a 1.1 km wide "floodplain". Generally the valley maintains a relatively constant width of around 300-400 m. The main variation in the form of the Auob is a gradual decrease in relative relief along the valley. Slopes are maintained at a relatively constant 18° to 25° and the valley floor is flat and sandy with sporadic *Acacia erioloba* spp. However, the depth of the valley floor beneath the duricrust plateau surface declines to between 12 and 15 m near to the Nossop confluence. The valley often appears deeper where dune sands abut the valley flank, increasing the relative relief to over 20 m in places.

Linear dunes are juxtaposed with the valley flanks more frequently within this section of the valley than seen further north. This appears to be primarily due to the valley cutting diagonally across the dune trend. There are locations, however, where the duricrust plateau surface is devoid of any sand veneer. These locations occur primarily on the western valley flank, with dune sands more commonly abutting the eastern flank. Where the Auob joins with the Nossop to the north of Twee Rivieren it is the smaller of the two valleys at approximately 300 m, its width being around 75% of that of the Nossop.

(ii) The Nossop Valley

The Nossop Valley was studied during 1989 and 1990, mainly where it forms the boundary between the Botswana Gemsbok National Park and South African Kalahari Gemsbok National Park. Brief visits were also made to the White and Black Nossop valleys near Gobabis and the Nossop at Aranos in Namibia.

The Nossop near Gobabis and Aranos

The Black and White Nossop rivers have been mentioned in part a of this section with regard to changes in drainage pattern (Hegenberger and Seeger, 1980). Figure 5.15 shows the present-day courses of both rivers in the vicinity of Gobabis, with a probable former course of the White Nossop superimposed. This would have joined the Black Nossop some 20 km further north than the present confluence.

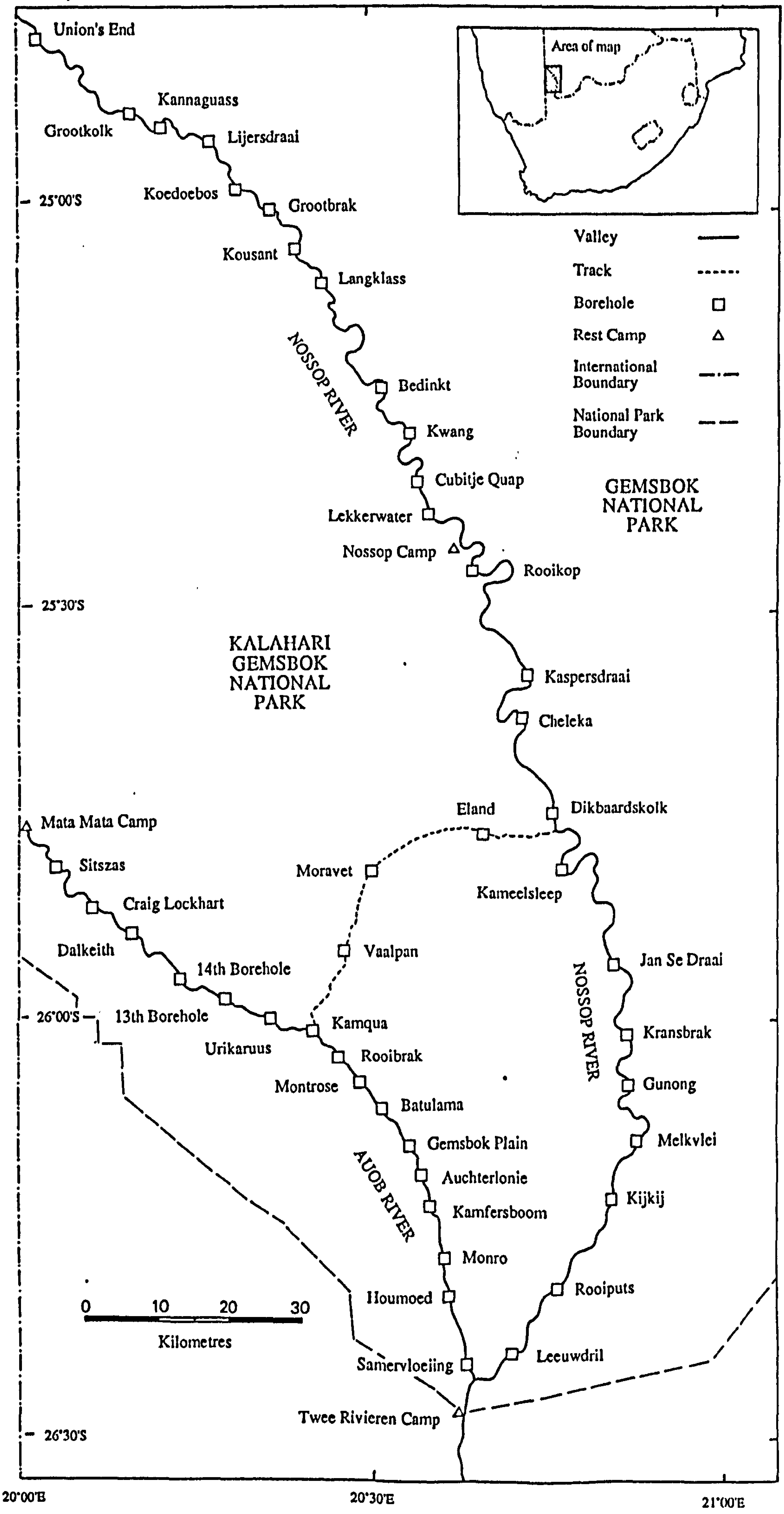


Figure 5.14: The location of wildlife boreholes along the Auob and Nossop valleys within the Kalahari Gemsbok National Park.

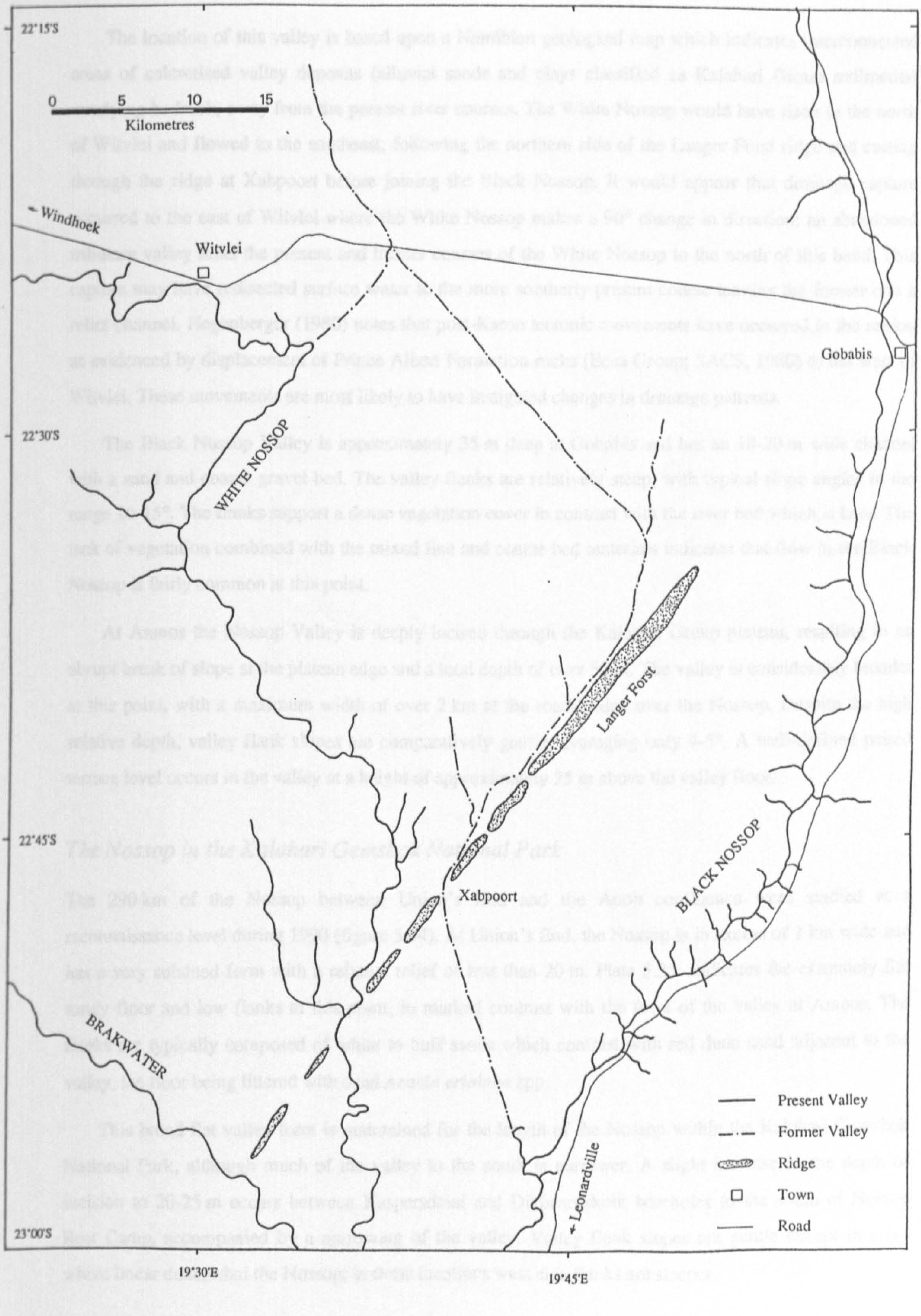


Figure 5.15: Drainage diversion in the White Nossop River to the west of Gobabis (Namibia).

The location of this valley is based upon a Namibian geological map which indicates interconnected areas of calcretised valley deposits (alluvial sands and clays classified as Kalahari Group sediments) overlying bedrock, away from the present river courses. The White Nossop would have risen to the north of Witvlei and flowed to the southeast, following the northern side of the Langer Forst ridge and cutting through the ridge at Xabpoort before joining the Black Nossop. It would appear that drainage capture occurred to the east of Witvlei where the White Nossop makes a 90° change in direction; an abandoned tributary valley links the present and former courses of the White Nossop to the north of this bend. This capture may have redirected surface water to the more southerly present course leaving the former one a relict channel. Hegenberger (1980) notes that post-Karoo tectonic movements have occurred in the region as evidenced by displacement of Prince Albert Formation rocks (Ecca Group; SACS, 1980) to the west of Witvlei. These movements are most likely to have instigated changes in drainage patterns.

The Black Nossop Valley is approximately 35 m deep at Gobabis and has an 18-20 m wide channel with a sand and coarse gravel bed. The valley flanks are relatively steep, with typical slope angles in the range 10-15°. The flanks support a dense vegetation cover in contrast with the river bed which is bare. The lack of vegetation combined with the mixed fine and coarse bed materials indicates that flow in the Black Nossop is fairly common at this point.

At Aranos the Nossop Valley is deeply incised through the Kalahari Group plateau, resulting in an abrupt break of slope at the plateau edge and a total depth of over 55 m. The valley is considerably broader at this point, with a maximum width of over 2 km at the road-bridge over the Nossop. Despite the high relative depth, valley flank slopes are comparatively gentle, averaging only 4-5°. A well-defined paired terrace level occurs in the valley at a height of approximately 25 m above the valley floor.

The Nossop in the Kalahari Gemsbok National Park

The 290 km of the Nossop between Union's End and the Auob confluence were studied at a reconnaissance level during 1990 (figure 5.14). At Union's End, the Nossop is in excess of 1 km wide and has a very subdued form with a relative relief of less than 20 m. Plate 5.35 indicates the extremely flat sandy floor and low flanks at this point, in marked contrast with the form of the valley at Aranos. The flanks are typically composed of white to buff sands which contrast with red dune sand adjacent to the valley, the floor being littered with dead *Acacia erioloba* spp.

This broad flat valley form is maintained for the length of the Nossop within the Kalahari Gemsbok National Park, although much of the valley to the south is narrower. A slight increase in the depth of incision to 20-25 m occurs between Kaspersdraai and Dikbaardskolk boreholes to the south of Nossop Rest Camp, accompanied by a narrowing of the valley. Valley flank slopes are gentle except in areas where linear dunes abut the Nossop; in these locations west side flanks are steeper.

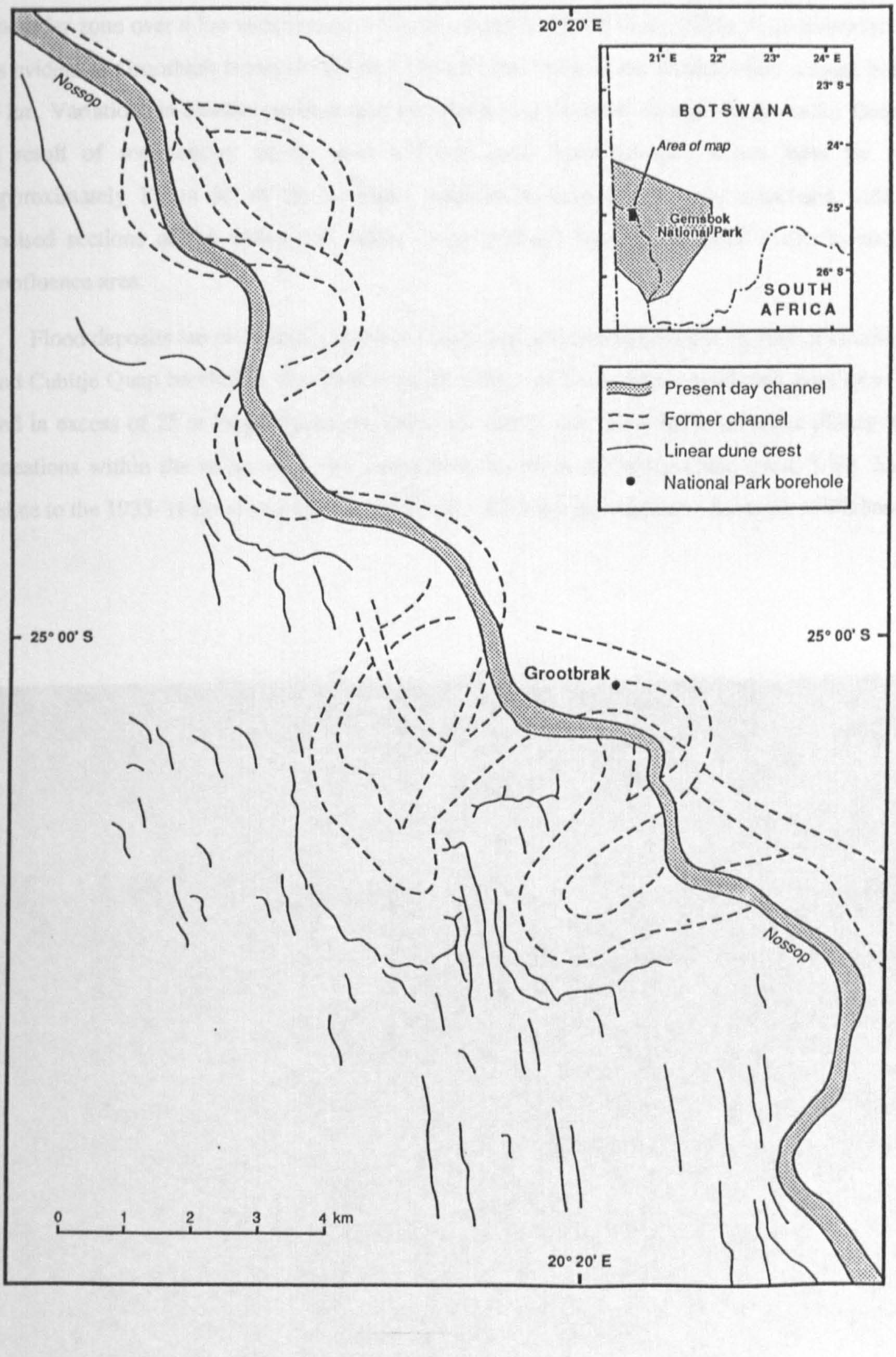


Figure 5.16: Abandoned meander channels in the Nossop Valley at Grootbrak wildlife borehole.

VARIATIONS IN VALLEY MORPHOLOGY

Field studies and analysis of aerial photography indicate variations in the planimetric form and sedimentary deposits within the valley. There is evidence to suggest that the current course of the Nossop has meandered over a wide floodplain set within a broader valley. Whilst the course of the present channel meanders extensively, this older valley is much less sinuous. It appears to have been infilled with sediment, so that only ghost traces of former meanders can be identified. For example, an abandoned meander zone over 4 km wide occurs at Cheleka borehole (Nash *et al.*, 1993). More extensive meandering is evident at Grootbrak borehole (figure 5.16) with the width of the meander belt varying between 3 and 5 km. Variations in channel position near Grootbrak may be due to changes in the valley floor gradient as a result of sedimentary inputs from a minor north bank tributary which joins the main valley approximately 13 km SE of the borehole. Abandoned meanders are only associated with less deeply incised sections of the valley, the valley being confined between duricrust flanks towards the Auob confluence area.

Flood deposits are evident in a number of locations, most notably in the vicinity of Grootbrak, Kwang and Cubitje Quap boreholes. The flood deposits consist of linear ridges of silt and sand up to 75 cm high and in excess of 25 m long aligned parallel to the valley axis. They occur on aerial photography as pale lineations within the valley floor, for example to the north of Cubitje Quap (plate 5.36). Such deposits relate to the 1933-34 flood (National Parks Board, 1986) and are evident as far south as Dikbaardskolk.

Plate 5.35: The Nossop Valley at Union's End, looking northeast.

Plate 5.36: Flood deposits in the Nossop Valley north of Cubitje Quap borehole (from aerial photography; scale 1:50,000). Botswana Department of Surveys and Lands photo number S.W.K. 36 (062) dated 25 July 1987.

Plate 5.37: The confluence of the Auob and Nossop valleys (from aerial photography; scale 1:50,000). Botswana Department of Surveys and Lands photo number S.W.K. 50 (038) dated 22 July 1987.

As noted above, the main change in the valley form of the Nossop occurs between Kaspersdraai and Dikbaardskolk boreholes. South of this point the valley becomes more incised and outcrops of duricrusts become more apparent within the valley flanks. These duricrusts are mostly altered and partly silicified calcretes, and occur intermittently as low 1 to 2 m high cliffs. In this section duricrusts are less apparent is between Kijkij and Melkvlei boreholes where the flanks are predominantly sandy. Towards the Auob confluence dune sands increasingly encroach upon the western side of the valley where the course of the valley swings to a SSW orientation, perpendicular to the dune alignment. As in the Auob valley, dune sands actually spill over the duricrust cliffs and cover the valley flanks. Plate 5.37 shows the the Auob-Nossop confluence, the encroaching dunes creating a serrated edge to the valley.

Few observations were made to the south of Twee Rivieren, but the most notable feature of the Nossop is that it becomes increasingly incised towards the confluence with the Molopo Valley. Red dune sands cover the valley flanks over much of this section, mainly due to the increased aeolian activity associated with vegetation removal by grazing to the south of the National Park. An additional point to note is the presence of low linear dune ridges on the floor of the Nossop in a number of locations. These ridges are almost perpendicular to the valley axis and reach a maximum height of around 1 m above the valley floor. In most cases they appear to be direct continuations of linear dunes adjacent to the valley, suggesting the transportation of aeolian sediments off the plateau surface onto the valley floor.

5.3.2 Molopo and Kuruman valleys

(a) Previous studies

The Molopo and Kuruman valleys (figure 5.17) form the southern part of the Kalahari exoreic drainage system, the former delimiting the Botswana/South Africa border and the latter cutting across the Northern Cape Province. They are considered separately in this chapter from the Auob and Nossop valleys which they join in the southwestern Kalahari. This is in part due to differences in morphology, but also because detailed fieldwork has been undertaken in the Kuruman as opposed to the reconnaissance level work in the three other systems.

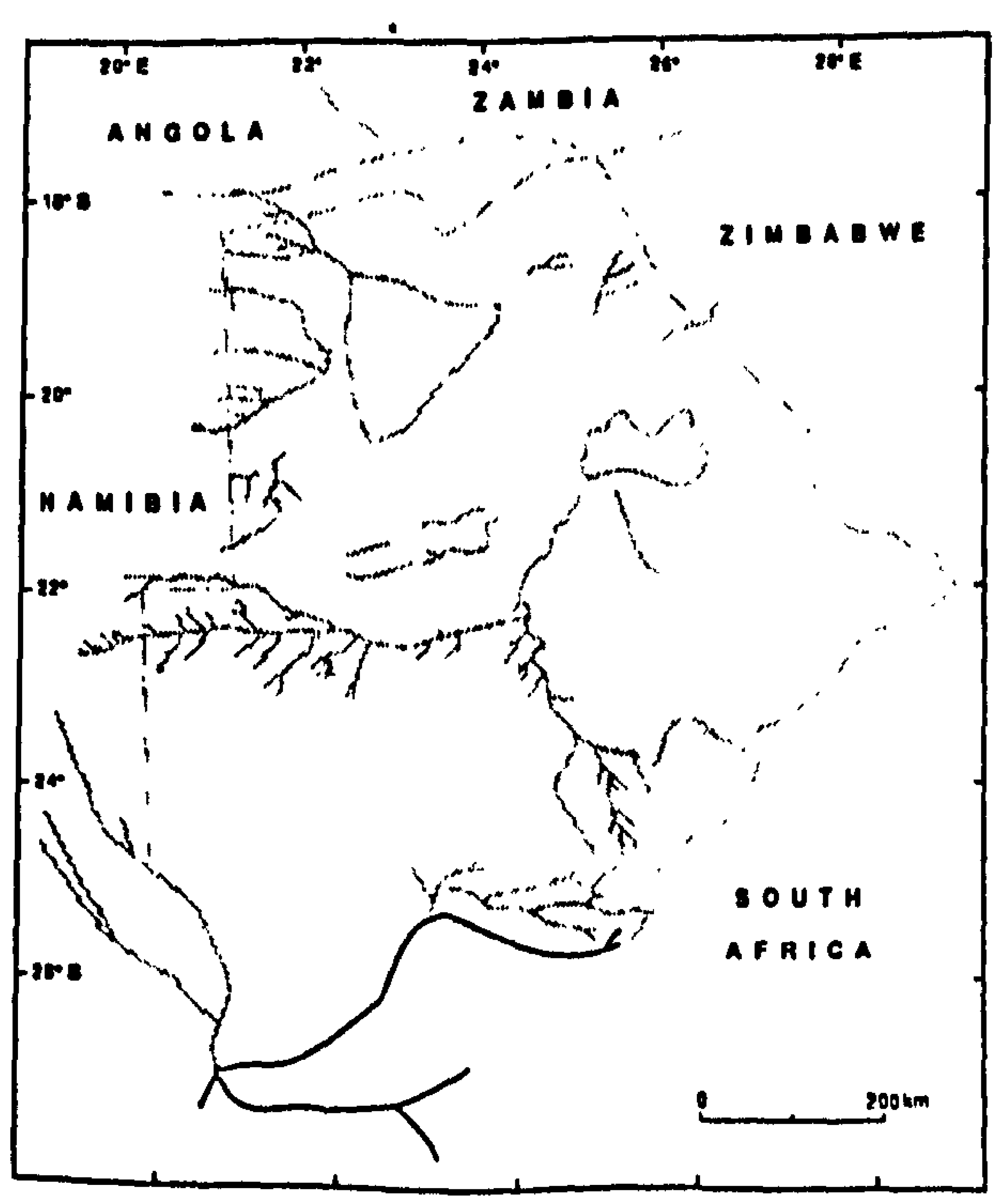


Figure 5.17: The Molopo and Kuruman valleys.

References to the Molopo and Kuruman valleys have been made since the early nineteenth century following the establishment of the London Missionary Society mission at Kuruman in 1820 (Thomas and Shaw, 1991*a*). Despite the importance of the town of Kuruman, very little reference has been made to the river which rises from springs in its vicinity, most writers being concerned with the Molopo. Earliest references to the Molopo are provided by Bain who crossed the river in 1826, describing it as a "fine running stream" (Lister, 1949). Bain also made observations of the environment near the Molopo, describing travelling through grassland for several hours on either side of the valley (Campbell and Childs, 1971) in an area now dominated by *Acacia* and *Terminalia* scrub savanna (Thomas and Shaw, 1991*a*).

The form of the Molopo has been mentioned by a variety of authors, including Rogers (1907 p.12) who described its course as "...often very slightly defined; at places [it] may be crossed without being noticed by the traveller". Records by European travellers indicate that the Molopo may have carried more water in the historic past than at present, the majority of references indicating diminution of surface water supplies (Thomas and Shaw, 1991*a*). Cornwallis-Harris crossed the river in 1836, describing it as having;

> "a broad shallow bed, covered with turf, traversed by a deep stream about ten yards wide, completely overgrown with high reeds... [with] an abundance... of hippo" (Cornwallis-Harris, 1852 p.66).

Gordon Cumming (1850) also indicates flowing water in the Molopo when he crossed it in 1843. He reports that;

> "...this darling little river is here completely concealed by lofty reeds and long grass which clothe its margins to a distance of at least a hundred yards".

There is considerable evidence for former flow within the Molopo, discussed by Du Toit (1927) and Wellington (1955). The valley cuts through banded ironstone formations near Phitsane (Aldiss, 1985) and has incised a 30 m deep channel through quartzite for over 10 km near Khuis (Wellington, 1955). Rogers (1936) also notes terrace levels within the valley. Reasons for the cessation of flow are provided by Wellington (1955 p.56). Both the Kuruman and Molopo are fed by perennial springs (Kokot, 1948), and Wellington proposes that headward erosion by more aggressive rivers draining the Bushveld complex may have captured Molopo headwaters and tapped underground aquifer sources. The only attempt to assign absolute dates for periods of former flow in the Molopo has been made by Heine (1978*b*, 1979, 1981, 1982), who, on the basis of radiocarbon dating shell material invokes perennial drainage in the Molopo during the period 16,500 to 15,500 years BP.

Other references to the Molopo have concerned a postulated ancient link with other valley and river systems to the north. This dates to the suggestion of Barber (1895, cited in Kokot, 1948) that the Molopo and Mashoweng (the Moshaweng tributary to the Kuruman) must once have been strong, permanent rivers. He further suggested that the drying up of these valleys could not be solely attributed to a decrease in rainfall and that an old Boer story that the Okavango had at one time "watered the Kalahari" might be true. This proposition culminated in the suggestion by Schwarz (1920) of a "Proto-Orange" River joining the Molopo near Makopong, as discussed in section 5.2.1*a* above. Changes in drainage patterns were also proposed by Rogers (1934) on the basis of similarities between fossil mollusca found in the Molopo and more northerly river and *mekgacha* systems. Wellington (1955) considered such a link with a drainage line from the northeast to be a possible explanation for the vigorous erosion of the gorge at Khuis and for the width of the Orange River below the Aughrabies Gorge.

There is much evidence of an extensive network of buried channels (figure 5.18) beneath the Molopo (Smit, 1977; Gould and Rathbone, 1985; Levin *et al.*, 1985; Gould *et al.*, 1987, 1989). Smit (1977) identified northward draining buried channels cut into the sub-Kalahari surface. However, southerly draining channels have also been identified (e.g. Gould and Rathbone, 1985; Levin *et al.*, 1985), leading to the suggestion that the Molopo incised its course along an alignment predetermined by a buried channel (Smit, 1977; Rathbone and Gould, 1982; Bruno, 1985). The present Molopo Valley may have developed by headward erosion, which would explain why its course cuts across buried north-trending channels. The buried channels beneath the Molopo and Kuruman have been extensively prospected due to their uranium bearing potential, leading to the discovery by Levin *et al.* (1985) of low quality uraniferous organic-rich diatomaceous earth on a farm to the west of Vanzylsrus.

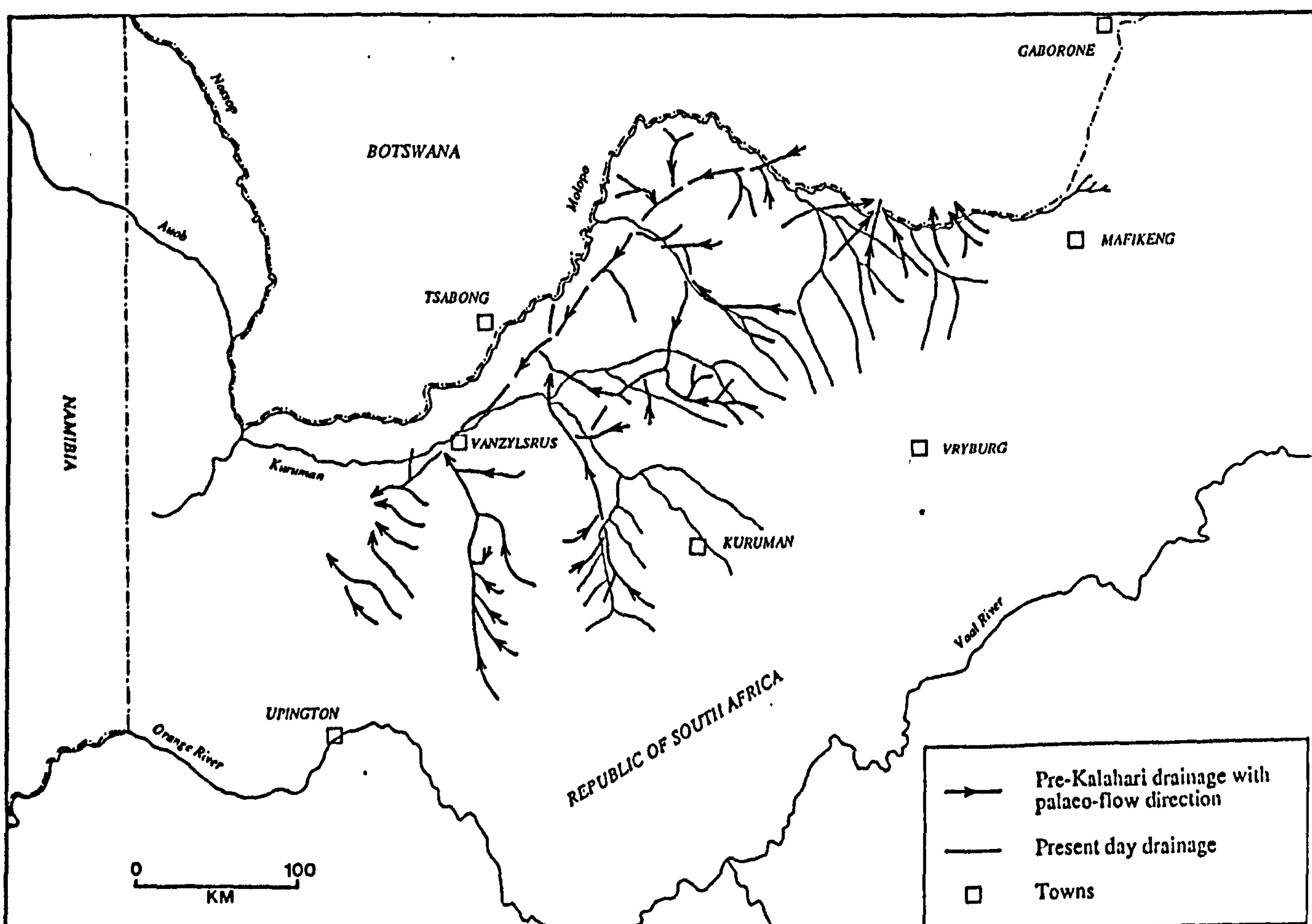


Figure 5.18: Pre-Kalahari drainages of the northern Cape Province (after Levin *et al.*, 1985).

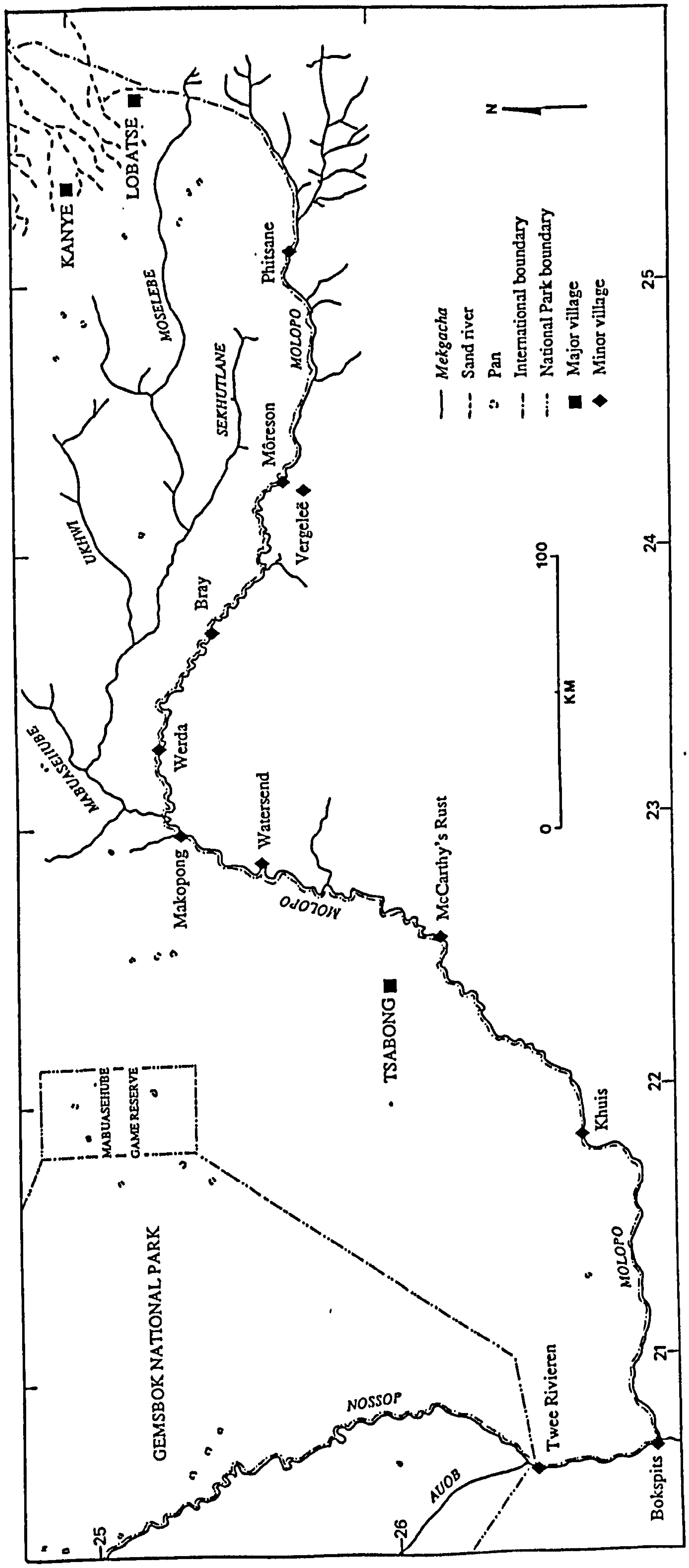


Figure 5.19: Locations along the Molopo Valley.

Plate 5.38: The Molopo Valley at Môreson Farm, Northern Cape Province.

Plate 5.39: The Molopo Valley at Watersend Farm, Northern Cape Province.

The majority of more recent studies have been made in a geological context. Boocock and Van Straten (1962) and Mallick *et al.* (1981) note a variation in duricrust types from the Molopo, with calcrete dominating the Upper Molopo and silcrete, calcareous sandstone and conglomeratic calcrete in lower sections. Duricrust studies along the Molopo have also been made by Summerfield (1978, 1982, 1983*c*), and terrace gravels and conglomerates have been discussed by Aldiss (1985). In addition, a number of references to the Molopo have been made in conjunction with the Botswana Department of Geological Survey "Molopo Farms Project". These studies have generated information on geological formations beneath and to the north of the Molopo from borehole and gravimetric surveys (Kimbell *et al.*, 1984; Gould *et al.*, 1987, 1989).

In addition to the details of historical and contemporary flows within the exoreic valley systems of the southern Kalahari contained in section 5.3.1*a*, other references have been made to flows within the Kuruman and Molopo valleys. As mentioned above, both valleys are fed by perennial springs and contain flow within the headwater sections of their courses on an annual basis. This also means that they are highly responsive to precipitation in their aquifer recharge areas (Shaw, Thomas and Nash, 1993). The Kuruman is fed by a number of dolomitic springs in the vicinity of Kuruman town, most notably the "Eye of Kuruman" which has maintained a constant output of 750 m^3 $hour^{-1}$ since at least 1820 (Thomas and Shaw, 1991*a*). As a result the river is perennial for the first 10 km below the "Eye" (Shaw, Thomas and Nash, 1993). The Molopo, however, is used extensively for irrigation (Kokot, 1948) and usually only contains flow as far west as Bray during average rainfall years (Grove, 1969). The furthest that flow has been known to extend along the Molopo is to Watersend farm (Shaw, Thomas and Nash, 1993). Dates of floods have been given in section 5.3.1*a*, but Gould *et al.* (1987) also mention less extensive flooding as far west as Makopong on the Molopo in 1976. It should be noted that whilst the 1933-34 flood affected all four of the main exoreic valleys, the Kuruman only contained limited flow (Kokot, 1948). In contrast, Kokot reports that the 1920 flood was extensive, and washed debris from Kuruman to Vanzylsrus.

(b) Field studies and aerial photography

(i) The Molopo Valley

The Molopo was studied between Môreson farm (10 km NNE of Vergeleë) and Watersend in northern Cape Province in 1990, and between McCarthy's Rust and Khuis in Botswana during the 1989 field season, a total distance of over 330 km (figure 5.19).

The Molopo between Môreson Farm and Watersend

The 200 km of the Molopo between Môreson Farm and Watersend represents a contemporary transition in the availability of surface water, with the presence of recent water only evident at the eastern end of the study section. Near Môreson the valley is over 150 m wide with a relative relief of less than 6 m (plate 5.38). The valley floor is comprised of a rich brown to black soil and contains a number of interconnected pans. The soil appears to indicate the presence of long-standing water, being much richer in organic material than the ephemeral pans seen within *mekgacha* in other locations. To the west of this point there was evidence for only limited standing water during the wet season. Indeed, no obvious channel within the valley was seen west of Môreson Farm.

At Bray a series of pans occur within the valley, which is around 100 m wide and is incised to a depth of between 8 and 10 m below the surrounding terrain. Observing the Molopo from the southern bank (i.e. looking towards Botswana), the absence of vegetation due to grazing pressures on the Botswana side of the border reveals a distinct difference between the valley floor and flank sediments. The valley flanks at this point clearly consist of red Kalahari Sand, contrasting markedly with the grey-white sediment of the valley floor. No exposures of duricrust were seen in this section of the valley, presumably due to the Kalahari Sand cover. The Molopo narrows to the west of Bray, although the depth of incision of the valley is still no more than 10 m, with slopes of up to 15°. A seasonal pan also occurs at Werda, where the valley attains a depth of incision of 10-12 m. The contrast between valley floor and flanks is more acute here due to almost total vegetation removal on the Botswana side of the valley.

The valley maintains this relatively narrow form until just east of Makopong, downstream of the Mabuaschube-Molopo confluence. To the west of the confluence the valley broadens considerably to around 250 m, although this broadening occurs over a distance of 2-3 km. The valley maintains this overall form as far as Watersend, the westernmost part of this study section, with seasonal pans in the valley base and red Kalahari Sand flanks. Additional traces of powdery calcrete were also evident on the uppermost valley flanks to the east of Watersend. The form of the Molopo at Watersend Farm is shown in plate 5.39.

The Molopo from McCarthy's Rust to Khuis

The Molopo south of McCarthy's Rust shows a similar form to that described above at Watersend, having a width of around 250-300 m and a total depth of incision of between 10-12 m. The valley flanks are characteristically gentle, with slopes of between 7° and 8° maximum. The contrast between the red Kalahari Sand valley flanks and the buff-white valley floor sediment is again apparent.

The major difference between the Molopo in this section and those locations observed up-valley is the presence of extensive outcrops of grey crystalline terrazzo silcrete. Outcrops of up to 1 m thick occur within the first 20 km south of McCarthy's Rust, many exhibiting solution hollows and occasionally crocodile-skin weathering textures on their upper horizontal surfaces. Vegetation types also differ between the two study sections, with thorn scrub dominating up-valley whilst grasses are prevalent in this section. Specimens of deep-rooting *Acacia erioloba* also occur sporadically, but the vegetation is otherwise dominated by grass. This apparent change in vegetation is as likely to be due to climatic variations as it is due to changes in water availability in the Molopo; the Molopo joins the Nossop in one of the most arid parts of the Kalahari (Thomas and Shaw, 1991*a*).

Indications of extensions of the southwestern Kalahari dunefield into the Molopo Valley occur near Middleputs where a series of small barchanoid dunes up to 3 m high can be seen in the valley bed. The final section of the Molopo observed during 1989 was at Khuis where a major outcrop of Khuis Quartzite occurs (Mallick *et al.*, 1981), cutting across the valley. Whilst observations were not made downstream of this bar, the valley does not appear to deepen considerably, maintaining its broad shallow cross-profile immediately west of the rock bar.

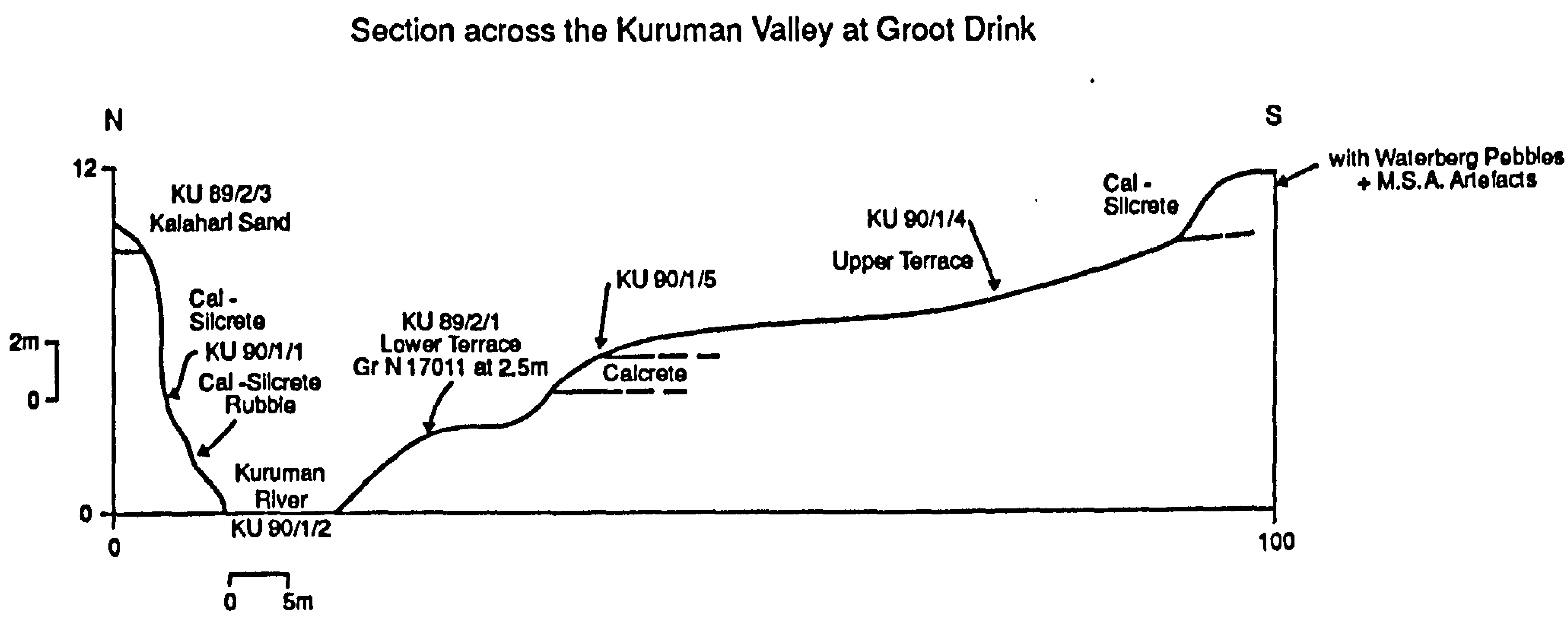
(ii) The Kuruman Valley

The Kuruman Valley was studied during both 1989 and 1990 field seasons, with further investigations by D.S.G. Thomas and P.A. Shaw at Aansluit Farm (see below) in 1991. The results of these investigations are documented in Shaw, Thomas and Nash (1993). Studies concentrated upon six sites along the valley between Groot Drink farm, 38 km upstream of the Kuruman-Moshaweng confluence, and Uitkyk farm, 15 km downstream from the confluence (figure 5.20*b*). The results from these sites will be considered systematically in a downstream direction, with an overview of information from the Kuruman at the end of this section. Sedimentary data for samples from each study site are given in table 5.2.

Groot Drink (26°55'16"S 22°44'33"E)

The Kuruman is incised to an average depth of 10 to 15 m beneath the surrounding Kalahari Sand surface at Groot Drink. It is characterised by relatively steep-sided duricrust flanks with two terrace levels clearly identifiable, one at a mean height of 2.5 to 3.0 m above the valley floor and another at 7.5 to 8.0 m. The typical valley form is shown in figure 5.20*a* for the study site at Groot Drink, and shows the relative height of both terrace levels. The terraces are generally best preserved on inner bends along the valley (plate 5.40). The uppermost outcrops exposed on the valley flanks are highly indurated silcretes and cal-silcretes, which are conglomeratic in places, containing jasper and Waterberg pebbles. The duricrusts in the steep northern flank show a change in morphology down the exposed profile, associated with an increased calcite content. The horizontal duricrust surface has a karst-like appearance with a honeycomb texture. Uppermost sections are typically more silicified, with weakly indurated cal-silcrete containing opalline silica veining in lower sections. Some Middle Stone Age artefacts were found on the southern valley flank lying upon cal-silcrete at the top of the upper terrace level; this indicates that the terrace predates these artefacts.

Mean grain sizes for samples from the upper and lower terraces are shown in table 5.2, along with samples of local Kalahari Sand and river bed sediment. The terraces are sedimentologically similar, being composed of fine sand with a minor medium sand and silt/clay component, although the lower terrace (samples KU 89/2/1 and KU 90/1/1) has a higher silt/clay content. Neither terrace showed well developed bedding in the limited exposures at Groot Drink, both being relatively homogenous. Both terraces do, however, contain calcrete nodules, with the upper terrace weakly calcretised in its upper sections. Molluscs are common within the sediments of both terraces, the upper level containing specimens of the terrestrial gastropod *Xerocerastus* and the lower terrace containing a variety of aquatic genera. The terrestrial species are unsuitable for absolute dating; contemporary species are highly mobile and any fossil specimens present are likely to have accumulated since the formation of the upper terrace. The lower terrace, sampled at a height of 2.5 m above the river bed in an exposure on the side of an old wagon cutting, contains a 15 cm concentrated layer of whole and comminuted *Ceratophallus* (75% of species present), *Lymnaea* (20%) and *Burnupia* (5%). A sample of these freshwater species were radiocarbon dated to 320 ± 150 years BP (GrN 17011), a date which indicates recent formation but may also suggest considerable sample contamination or a mixed age of shell material.



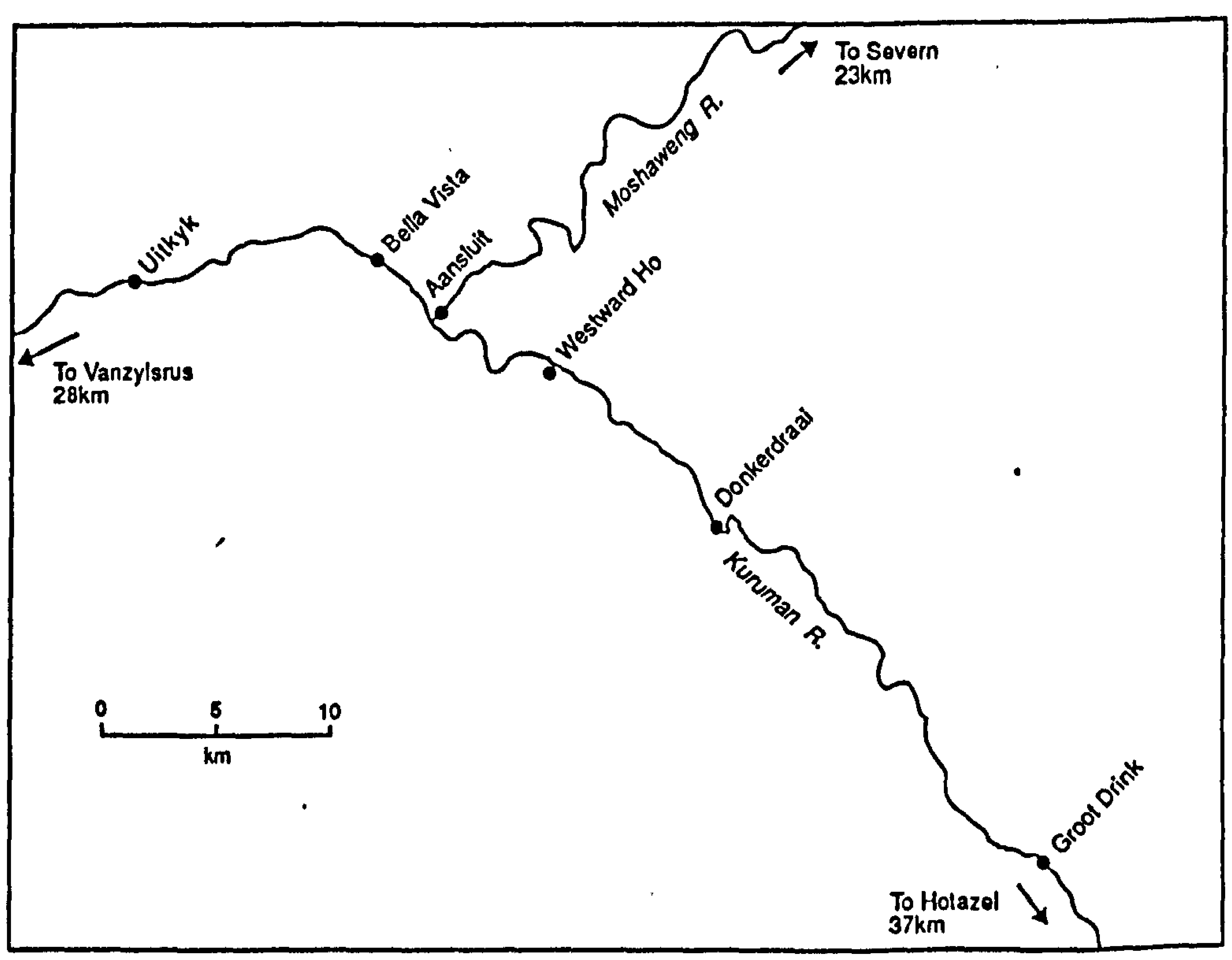


Figure 5.20: a) Cross-section through the Kuruman at Groot Drink, b) Locations along the Kuruman Valley mentioned in section 5.3.2.

Donkerdraai Farm (26°48' 14"S 22°36' 40"E)

At Donkerdraai, a major meander bend occurs within the Kuruman, the inner bank consisting of the upper terrace surface seen at Groot Drink at a consistent height of 8.0 to 9.0 m above the valley floor. The surface is relatively horizontal but slopes slightly upwards where it merges with the valley flanks, suggesting incursion of slope material onto the back of the terrace. The lower terrace was only poorly developed at this location, but could be clearly identified within 500 m down-valley of the meander.

Westward Ho Farm (26°44' 27"S 22°31' 40"E)

Both upper and lower terrace levels were present at Westward Ho, distinguished by their differing gastropod species and, to a lesser extent, colour. Sedimentary analysis of samples from each terrace on the northeastern side of the valley are given in table 5.2. As no vertical exposures of terrace sediment occur, samples were collected from a depth of 30 cm below the surface. These indicate that whilst variations in grain size occur between Westward Ho and Groot Drink, the general characteristics of each terrace are largely consistent.

Aansluit Farm (Kuruman-Moshaweng confluence, 26°43' 31"S 22°28' 31"E)

Extensive exposures of terrace sediments occur within the Moshaweng Valley at Aansluit, immediately up-valley of the Kuruman-Moshaweng confluence. This exposure occurs on the west side of the valley, extending for over 100 m with a maximum height of 4 m, and may possibly be equivalent to the lower terrace seen at Groot Drink and Westward Ho (but see discussion below). A series of discontinuous sub-horizontal beds of fine sand are exposed, with occasional silt- and clay-rich bands. Material suitable for absolute dating (including charcoal, bone and Xerocerastus Molluscs) occurred in a number of places within the sedimentary sequence, and sections of the exposure containing such material are indicated schematically on figure 5.21 (logged in 1990 and 1991). A charcoal layer (KU 90/5/7) within fine sand in profile KU 90/5 gave a radiocarbon age of 1,780 ± 60 years BP (GrN 18070).

The sedimentary sequence indicated in figure 5.21 consists of a series of depositional cycles, with a general "fining upward" sequence of sand followed by laminated silt with a final clay-rich deposit. Profile KU 90/3 contains at least six such cycles, which are indicative of periodic flooding, possibly including back-flooding from the river confluence (Shaw *et al.*, 1993). There is a marked lack of lateral continuity between laminated silt or clay bands along the exposure, suggesting erosion of previous flood deposits prior to further deposition. Such erosion may be associated with the onset of a flood-event, with the falling stage of the flood leading to sediment deposition, possibly including dune forms which would produce the dipping strata present in a number of locations. Reworking of sediment is suggested by the presence of rounded calcrete pebbles and comminuted shell and bone material within the depositional sequence.

Table 5.2: Sedimentary data from Holocene flood deposits in the Kuruman Valley. Sample locations are shown in figure 5.20.

Sample	Coarse	% Sand Med.	Fine	% Silt/ Clay	Folk and Ward Statistics Mean	Sort.	Skew.	Kurt.	Munsell colour
GROOT DRINK Farm									(26°55'16"S 22°44'33"E)
Lower Terrace (S) KU 89/2/1	1.9	17.0	75.6	5.5	2.65	0.73	0.01	1.26	10YR 6/2
Kalahari Sand KU 89/2/3	0.9	10.6	86.3	2.2	2.67	0.58	-0.01	1.13	7.5YR 5/6
Lower Terrace (N) KU 90/1/1	1.8	1.8	87.2	9.2	3.13	0.60	0.14	1.21	10YR 6/2
River Bed KU 90/1/2	18.5	51.0	30.3	0.2	1.71	0.73	-0.36	2.52	10YR 7/2
Upper Terrace (S) KU 90/1/4	2.7	22.3	72.2	2.8	2.38	0.66	-0.14	1.05	10YR 6/3
U. Terrace calcrete KU 90/1/5	42.0	5.5	31.5	21.0	2.16	1.95	-0.08	0.68	2.5YR 7/1
WESTWARD HO Farm									(26°44'27"S 22°31'40"E)
Lower Terrace KU 90/7/1	15.3	1.3	64.6	18.8	3.13	1.45	0.16	2.09	10YR 6/2
Upper Terrace KU 90/7/2	0.6	10.9	84.8	3.7	2.71	0.59	0.10	1.19	10YR 6/3
AANSLUIT Farm									(26°43'31"S 22°28'31"E)
KU 90/3/1	5.6	43.5	48.7	2.2	2.13	0.71	0.11	1.15	10YR 5/3
KU 90/3/3	1.9	9.1	85.6	3.4	2.74	0.57	0.09	1.23	10YR 6/3
KU 90/3/5	4.4	74.5	11.5	9.6	2.90	0.90	-0.15	1.24	10YR 6/3
KU 90/3/7	3.0	22.4	72.4	2.2	2.45	0.66	-0.10	1.04	10YR 6/3
KU 90/3/8	0.5	6.0	93.4	0.1	2.58	0.39	0.14	1.34	10YR 7/3
KU 90/5/1	2.6	16.6	77.6	3.2	2.64	0.71	-0.02	1.02	10YR 5/3
KU 90/5/2	9.4	11.4	73.8	5.4	2.64	0.98	-0.16	1.29	10YR 4/2
KU 90/5/3	0.1	1.5	95.8	2.6	2.98	0.42	0.11	1.07	10YR 6/3
KU 90/5/4	27.2	11.3	56.0	5.5	1.90	1.37	-0.44	0.76	10YR 4/2
KU 90/5/5	3.3	6.2	80.1	10.4	3.23	0.74	0.19	1.56	10YR 6/4
KU 90/5/6	1.1	1.7	88.6	8.6	3.33	0.46	-0.04	1.38	10YR 7/2
BELLA VISTA Farm									(26°42'26"S 22°26'49"E)
KU 89/1/1	0.5	4.3	91.8	3.4	2.84	0.52	-0.14	1.07	10YR 5/3
KU 89/1/3	0.6	22.4	76.5	0.5	2.38	0.47	-0.02	0.98	10YR 6/2
KU 89/1/4	6.4	69.6	20.7	3.3	1.65	0.54	0.11	1.31	10YR 6/3
KU 89/1/5	0.2	5.2	91.9	2.7	2.62	0.50	0.10	1.22	10YR 7/3
KU 89/1/7	0.7	6.8	88.9	3.6	2.62	0.54	0.15	1.25	10YR 7/2
KU 89/1/8	0.2	2.3	97.1	0.4	2.82	0.21	-0.04	1.38	10YR 8/2
KU 89/1/9	3.5	9.5	85.4	1.6	2.58	0.59	-0.13	1.32	10YR 7/2

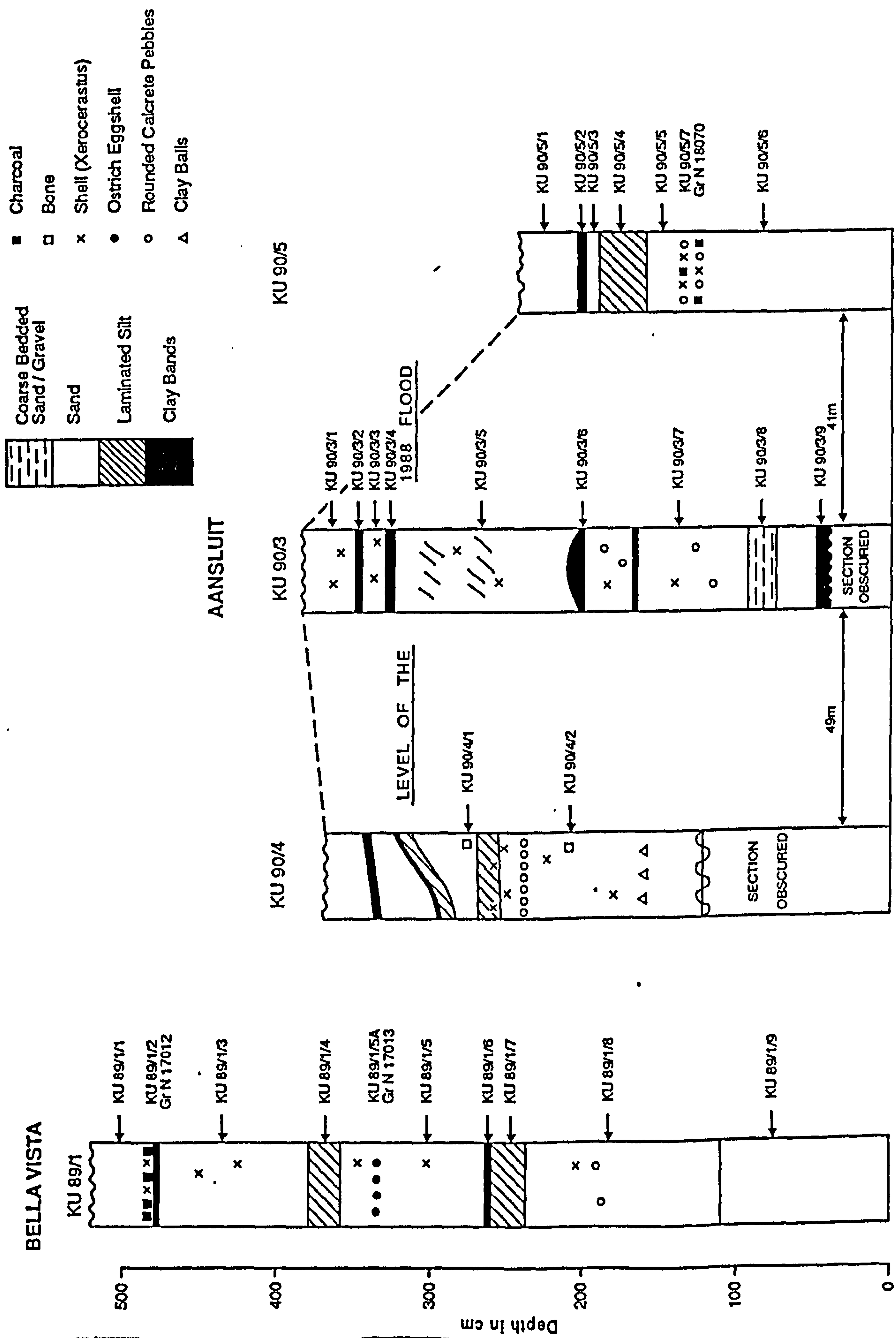


Figure 5.21: Schematic sedimentary logs of flood deposits in the Moshaweng and Kuruman valleys at Bella Vista and Aansluit farms, near the Kuruman-Moshaweng confluence.

Plate 5.40: The Kuruman Valley at Groot Drink Farm, looking northwest.

Plate 5.41: Flood deposits exposed in the Kuruman River at Bella Vista Farm.

As mentioned in part *a* of this section, the Kuruman experienced an extensive flood along its entire length during 1988. The maximum level of floodwater at Aansluit is indicated by a "trash line" at a height of 3.1 m above the river bed. As the Moshaweng was not known to have flooded at the same time as the Kuruman it can be assumed that this level is associated with back-flooding from the confluence.

Bella Vista Farm (26°42'26"S 22°26'49"E)

The 1988 flood caused extensive damage along the Kuruman, including the destruction of the road bridge at Bella Vista Farm, 4 km downstream of the Kuruman-Moshaweng confluence (plate 5.41). The Kuruman has a 30-38 m wide channel at this point, a depth of incision of approximately 10 m, a near vertical south bank and slopes of 25° to 30° to the north. The flood caused extensive scouring around the end of the damaged bridge, resulting in the formation of an exposure of flood deposits in excess of 5 m high. These deposits were systematically logged in 1989 and can be seen in the background of plate 5.41 and in figure 5.21. This exposure, located immediately west of the bridge on the south side of the river bed, was destroyed during the reconstruction of the road bridge.

The sedimentary sequence at Bella Vista includes four depositional cycles of the type identified at Aansluit, and includes material suitable for absolute dating. Radiocarbon dating of ostrich eggshell at a height of 3.4 m above the river bed yielded an age of 2,840 ± 110 years BP (GrN 17013), whilst charcoal at 4.8 m was dated at 540 ± 30 years BP (GrN 17012). A major difference between this site and Aansluit is a lack of comminuted or rolled material, suggesting a different type of depositional environment. The fact that the ostrich shell was found as a localised deposit in a single layer indicates sediment deposition under comparatively gentle conditions (e.g. by bank overtopping) as opposed to higher energy dune movement (Shaw *et al.*, 1993). Lower energy conditions are not, however, indicated by sediment grain sizes which are largely consistent with those at Aansluit, although this is partly dependent upon local source materials.

The valley floor at Kuruman additionally contained laminated cross-bedded recent sediments when investigated in 1989. Scour of up to 2 m had occurred, with erosion particularly extensive where the 1988 flood had been diverted around the remains of the bridge.

Uitkyk Farm (26°43'06"S 22°20'36"E)

A track leads northwards from the Hotazel to Vanzylsrus road at Uitkyk Farm. This track reveals a section through the lower terrace level described at earlier sites, and indicates that the terrace has a levéed form on both north and south river banks. The presence of levées may support the evidence for gentle depositional environments suggested above, although the time period over which such bank overtopping has occurred is unknown.

General implications

The presence of two clearly defined terrace levels identified along this section of the Kuruman is possibly indicative of two periods of valley incision. There do, however, appear to be differences in the characteristics of the exposures above and below the Kuruman-Moshaweng confluence. These are firstly that only one terrace is identifiable below the confluence, and secondly that the terraces seen up-valley of the confluence show no evidence of bedding structures. It is not, therefore, advisable to relate the terrace down-valley of the confluence with the lower terrace seen at Westward Ho and Groot Drink without further evidence for similarity of age. However, despite the dangers of interpolation due to these differences, the sequence of flood deposits below the confluence probably provide an analogue for the formation of those up-valley, with terrace development proceeding by intermittent flood events. Shaw *et al.* (1993) note that the 1988 flood was associated with an average rainfall in the Kuruman area of 236 mm in the fourteen days following 9th February 1988. This represents a 1 in 50 to 1 in 500 year magnitude storm event. Whilst no data on discharge within the Kuruman at the onset of the 1988 flood is available, it is known that the bridge at Groot Drink was overtopped (pers. comm., owner of Groot Drink Farm, 30th July 1990) and that back-flooding up the Moshaweng reached a height of over 3 m. These flood levels would exceed the height of the lower terrace seen along the Kuruman, and may be analogous to those indicated by the palaeo-flood deposits.

The context of the material selected for radiocarbon dating also needs to be considered. The high standard deviation about the mean date of 320 years BP for the lower terrace shell material at Groot Drink has already been commented upon. This indicates that the activity of the sample has a high degree of uncertainty, probably due to contamination (Lowe and Walker, 1984). However, a further three points need to be considered in the interpretation of the absolute dates. Firstly, it should be noted that all dates are from partly reworked material and indicate the maximum possible age for each flood deposit. Secondly, the four dates indicate a Late Holocene age for the flood deposits, with no consistent pattern emerging. This probably reflects the ephemeral nature of floods within the Kuruman. Thirdly, of the four available dates for the Kuruman, three are from within homogenous sedimentary units and one lies on top of a clay layer (sample KU 89/1/2, GrN 17012 at Bella Vista). If a sand-silt-clay "fining upwards" sequence exists, then the three dates from within homogenous units represent maximum ages for when sandy material was being deposited. The date from charcoal on top of a clay band could indicate any time during the hiatus between the end of a sedimentary sequence and the commencement of the next flood event.

Shaw *et al.* (1993) note that the grain sizes encountered within terraces are consistent with the range of sedimentary characteristics for Kalahari Sand identified by Thomas (1987*a*). As such, despite the variability of terrace sediment grain sizes, they are unlikely to have any significance for palaeohydrological reconstruction, merely reflecting the range of material sizes available within the source sediment.

Finally, the relationship between the overall valley form and the terrace sediments needs to be considered. Both strongly and partially indurated duricrusts are present within the Kuruman, highly siliceous types exposed in the valley flanks and powdery calcretes within terrace sediments. Whilst borehole information is not available to indicate the extent of sedimentary fill within the valley, it appears that the upper terrace was deposited within a deeper valley. The deposits of the lower terrace exposed at Groot Drink probably represent more recent flood deposits deposited against these older terrace materials within the valley.

5.3.3 Moselebe Valley system

(a) Previous studies

The valleys of the Moselebe system (figure 5.22) drain the southeastern portion of Botswana, and include the Moselebe, Ukhwi, Sekhutlane and Selokolela valleys, which join with the Molopo via the Mabuasehube Valley (figure 5.23). The main Moselebe Valley has its headwaters just inside South Africa, rising on the hardveld to the south of Lobatse.

The Moselebe has received comparatively little attention in either the geomorphological or geological literature. Early references are limited, with only the Selokolela and Mabuasehube (Malatswana) valleys briefly mentioned by Schwarz (1920 p.139) as part of his "Thirstland Redemption" scheme. The lack of more recent study is due in part to the relatively thick cover of Kalahari Group sediments in the area which made mineral prospecting comparatively difficult. However, the few available studies are of particular interest in terms of establishing the possible modes of origin of Kalahari valley systems.

Geophysical studies have identified the presence of large buried channels beneath a number of valley courses (Peart, 1979). Masson-Smith and Evans (1966) located a buried 150 m wide, 30 m deep bedrock depression of rectangular cross-section 300m south of the Moselebe Valley, although details of the precise location of the study site are not included. Gould and Rathbone (1985) and Gould *et al.* (1987, 1989) note the presence of a major pre-Kalahari channel extending from 24°30'S 23°40'E to the Ukhwi-Moselebe confluence (25°10'S 23°40'E).

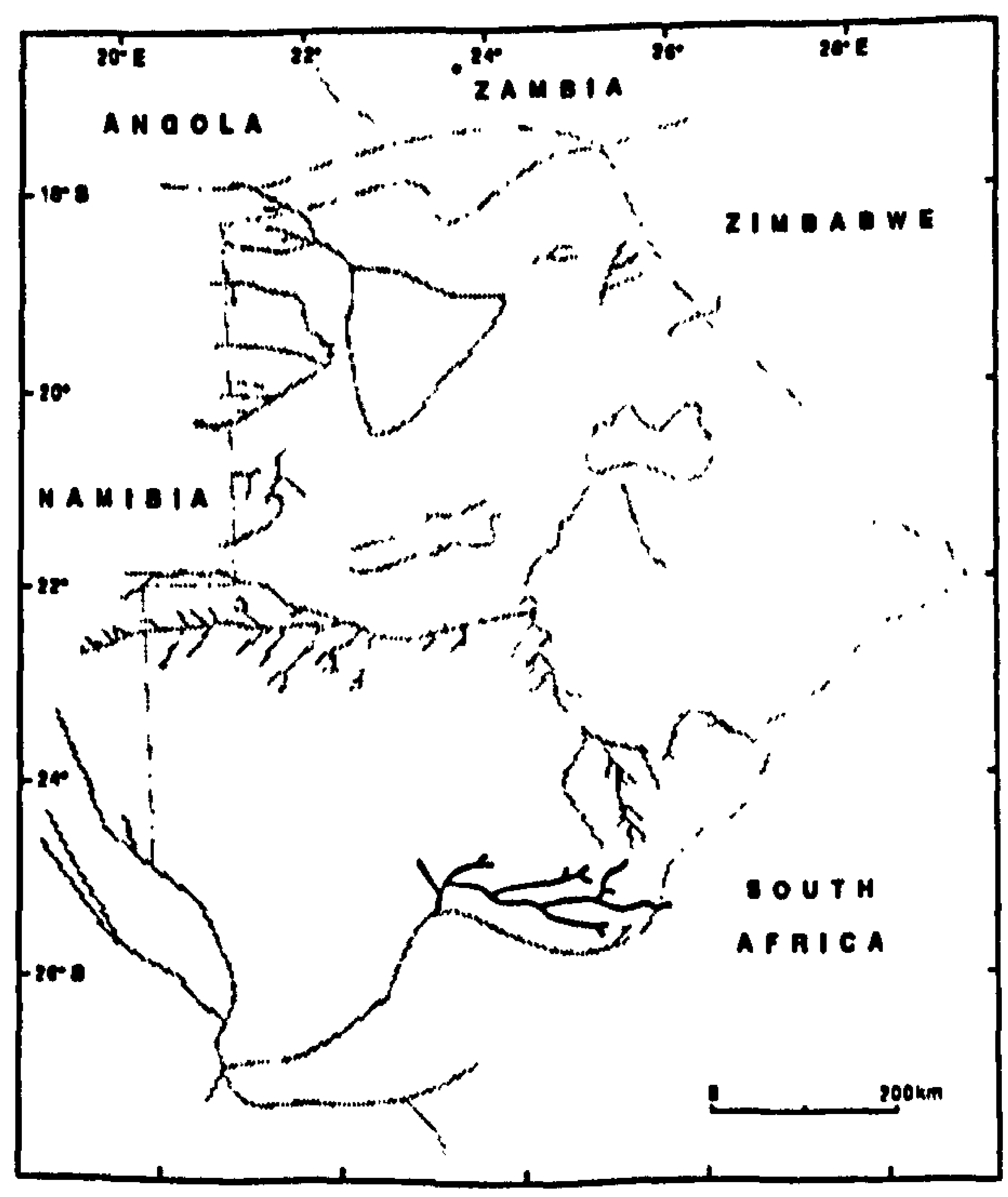


Figure 5.22: The Moselebe valley system.

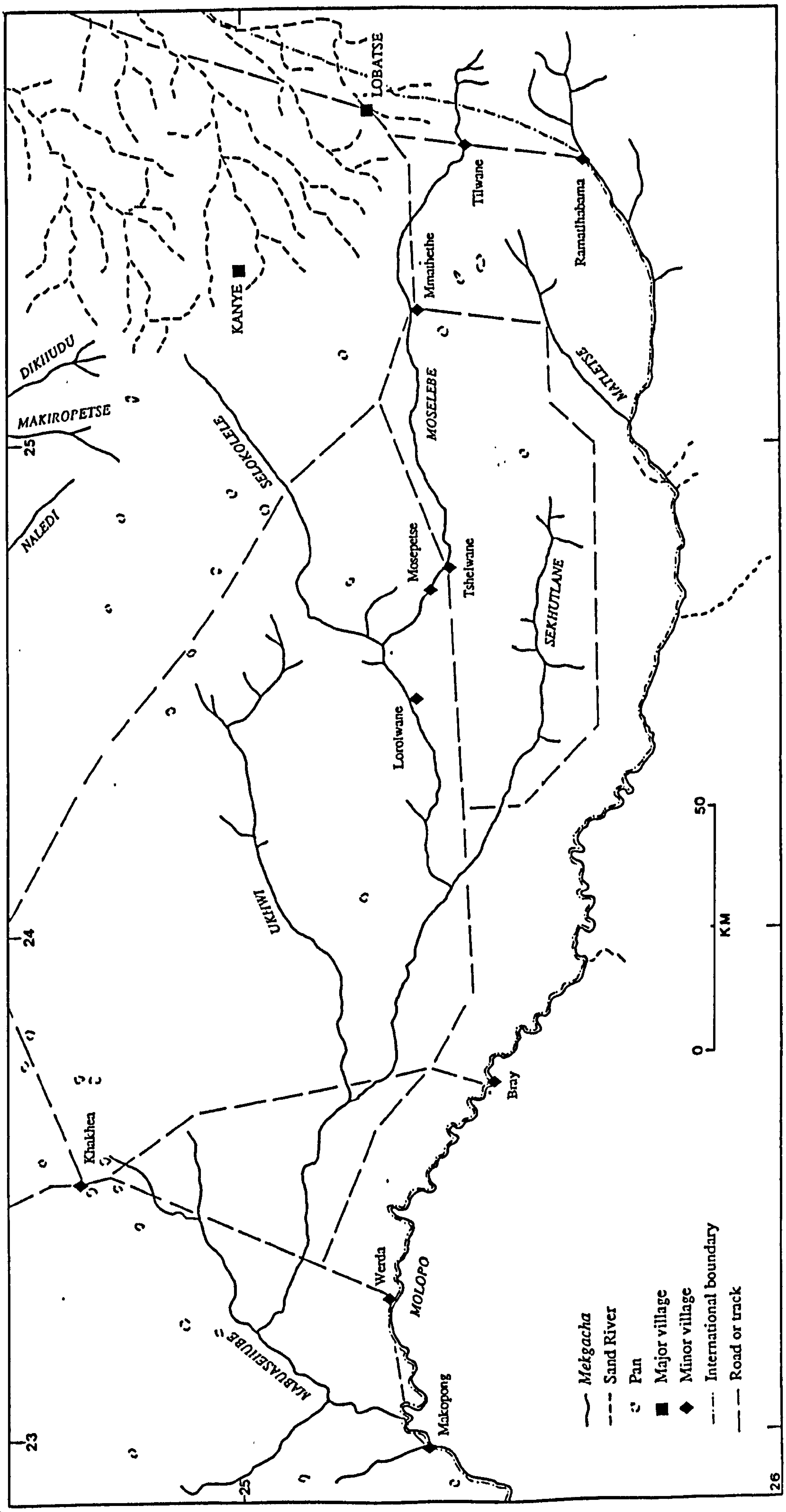


Figure 5.23: The valleys of the Moselebe system.

Peart (1979) has located a similar channel 200 m to the north of the Sekhutlane Valley, suggesting this may indicate upward migration of the valley course as the Kalahari Group sediments were deposited. The fact that the buried channel is not located directly beneath the present course is possibly indicative of lateral as well as upward migration of the valley. However, this may be an isolated buried channel which was coincidentally intersected during the study, and more detailed evidence is needed before an upward migration hypothesis can be supported.

A study by Aldiss (1985) of the Matletse Valley (a tributary of the Molopo) has revealed terrace levels and evidence for structural control. The line of the valley also cuts through the banded ironstone Mosi Ridge, suggesting that the course of the Matletse may have existed at least since the basal Kalahari Group sediments were deposited, thus predating the present course of the Molopo Valley. Gwosdz and Modisi (1983) also discuss the Matletse Valley in the context of calcrete resource evaluation.

(b) Field studies and aerial photography

The valleys of the Moselebe system were studied during 1989 and 1990 (figure 5.23) with investigations concentrating upon the main Moselebe Valley.

The Moselebe rises near Tilwane in southeastern Botswana (25°23'S 25°42'E) at an altitude of 1360 m asl (figure 5.23). It exhibits many characteristics typical of other *mekgacha* having only negligible relief. Where the main Ramatlhabama to Lobatse road crosses the valley it is a shallow trough of approximately 5 m depth with a poorly defined channel area less than 10 m across. However, in the vicinity of Mmathethe the valley is 700-800 m wide, reaching a width of 1.5 km and depth of over 10 m near Tshelwane.

The relief of the Moselebe in hardveld areas is generally subdued, with evidence of valley incision only being identifiable within the Kalahari sandveld. The most prominent feature of the valley within the sandveld is a low terrace level which is clearly identifiable from immediately east of Tshelwane to beyond the Moselebe-Ukhwi confluence (plate 5.42). Near Tshelwane the terrace level is at a height of approximately 2.5 m above the valley base, with a "channel" approximately 90 m wide between the terrace banks. The channel maintains this width until the Selokolela confluence. The terrace material contrasts markedly with the valley floor, the flanks generally covered by red Kalahari Sand and the base consisting of grey sediment with a high clay content.

Duricrust exposures (mainly crystalline conglomeratic silcretes) are common at the base of the terrace in this area. Spoil around the borehole in the base of the Moselebe at Tshelwane consisted of grey terrazzo silcrete. Silicified calcrete was also exposed in the terrace flank. More extensive outcrops of duricrust occur to the north of Mosepetse, particularly between 8 and 10 km northwest of the village. In this section, the duricrust-supported terrace level is up to 4.5 m above the valley floor, with the "channel" possibly incised through the silcrete. This is suggested by the extensively developed flat terrace level and steep slopes off the terrace to the valley base. The "channel" within the valley also exhibits steeper outer banks on the outside of bends, as expected if fluvial activity had been important in channel development.

Plate 5.42: The terrace level in the Moselebe Valley near Tshelwane.

Plate 5.43: The Sekhutlane Valley where it is intersected by the Mmathethe-Werda road.

Down-valley of the Moselebe-Selokolela confluence, the incised section of the Moselebe broadens to around 150-200 m, with exposures of silcrete still present. At the confluence, the Moselebe is the more deeply incised of the two valleys, with an overall depth of 6-7 m (as opposed to 3-4 m for the Selokolela). The two valleys are separated by a low sandy spur at the confluence with a relatively steep gradient from the Selokolela into the Moselebe Valley. This may indicate that the main valley has contained flow more recently than its tributary, although the widening of the channel section downstream of the confluence suggests that the Selokolela may have contributed flow at one time. The overall valley form is maintained to the west, although the incised central channel is less distinct in the vicinity of Lorolwane, with a width of 120 m and a depth of 2.5-3.0 m.

The Moselebe was also studied where it is intersected by the Bray to Khakhea road, having an overall depth of 7-8 m and a width in excess of 1 km. The terrace level is still evident in this area at a height of 3.5 to 4.0 m above the valley base with a 200 m wide channel present. Valley asymmetry is evident, with the terrace level on the southern valley flank being more pronounced than that to the north.

Three other valleys of the Moselebe system were studied in locations where they were intersected by main tracks. The northern Ukhwi Valley was crossed on the Bray to Khakhea road, where it was deeply incised and contained a terrace level at approximately the same height above the valley base as seen in the Moselebe (i.e. at 3.5 to 4.0 m). A similar channel section was also present with a width of 225 m, with some powdery calcrete exposed in the valley floor. Silcrete exposures also occur intermittently at the top of the terrace level.

The Sekhutlane Valley was crossed on the Mmathethe to Werda road, where it has a width of 1.7 km and a total depth of 6-8 m (plate 5.43). No terrace level was present although the eastern flank does contain two barely perceptible "steps", with the valley base being distinguished by the presence of a 150 m wide zone of grey clay-rich sediment. The presence of tall *Acacia erioloba* within the valley floor suggests the presence of groundwater reserves beneath the Sekhutlane. The final part of the Moselebe system considered by field studies was the Mabuasehube Valley which was crossed on the Makopong to Werda road. The valley was typically broad (approximately 1.2 km wide) and shallow (7-8 m deep) and contained powdery calcrete in its valley floor. No terrace level was evident, although a series of partly silicified calcrete exposures (up to 2 m high) at a height of 2.5 m above the valley base may be terrace remnants.

5.3.4 Serorome Valley

(a) Previous studies

In comparison with other Kalahari *mekgacha*, the Serorome Valley (figure 5.24) has received very little mention in the literature, excepting cursory mentions by travellers crossing the valley (e.g. Chapman, 1886, at Boatlaname). The only consideration of the geomorphology of the valley has been made by Boocock and Van Straten (1962), who suggest that the reason for cessation of flow in the Serorome is due to mantling of the headwater regions by Kalahari Sand.

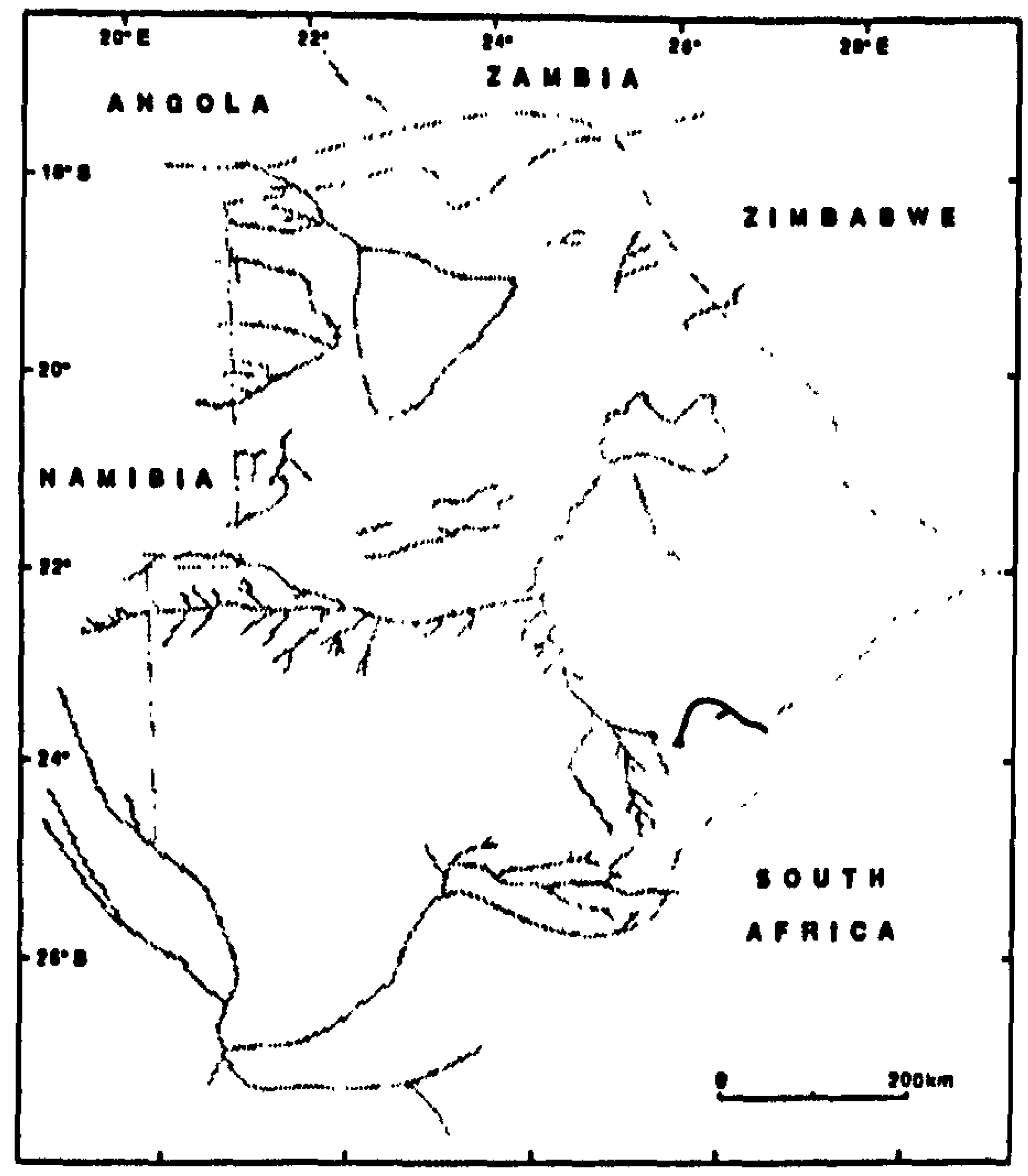


Figure 5.24: The location of the Serorome Valley.

The Bonwapitse Valley to which the Serorome is connected rises on a hardveld area and is still an ephemeral river. Other studies based in the Serorome include geological prospecting by Urangesellschaft (1978) and Gwosdz and Modisi (1983), and the study of silcrete profiles by Summerfield (1978, 1982).

(b) Field studies and aerial photography

The Serorome Valley was studied in 1990 over a 40 km headwater section of the valley to the south of Boatlaname village. Additional studies were made immediately adjacent to the Dibete veterinary cordon fence and in a segment 10 km west of the main Gaborone-Francistown road (figure 5.25).

The Serorome south of Boatlaname

The southernmost tributary of the Serorome (see figure 5.25) was encountered 70 km north of Molepolole on the Molepolole-Lephephe road. The form of the Serorome in this region (plate 5.44) consists of a broad, gently sloping valley, approximately 150 m across, with Kalahari Sand flanks. The overall relative relief is less than 5m, with a pan-lined base and occasional *Acacia* spp. in the valley floor. At 75 km north of Molepolole this tributary is joined by a much more prominent headwater valley, with an overall relief of 8m. The valley maintains this overall form, with a depth of 6-8 m, width of 100-200 m and infrequent pans in the valley floor, until approximately 103 km north of Molepolole (i.e. 6 km south of Boatlaname on figure 5.25).

Very few exposures of duricrust are seen along the valley in this headwater section, mainly due to the extensive Kalahari Sand cover. One exposure can be seen at 100 km north of Molepolole, where the main track crosses a valley meander. The valley flanks consist of a creamy-grey hard crystalline terrazzo-silcrete, similar to the silcrete exposed in the amphitheatre valley head of Letlhakeng Valley 1. This has been identified as a springline deposit by Shaw, Thomas and Nash (1993). The silcrete exposures are in excess of 2 m thick and show an opalline weathering rind. Uppermost sections of the profile are typically more silica-rich, with traces of calcite towards the profile base.

An interesting feature present on the eastern valley flank at 103 km north of Molepolole is an area of 2 m high grass-covered dunes of Kalahari Sand, indicative of dune-building activity since the formation of the valley. Also present in the valley floor near these dunes is a well revealing 1.5 m of sandy soil containing calcrete fragments.

The valley widens abruptly at around 5.5 km south of Boatlaname village (i.e. 103.5 km north of Molepolole), to reach a total width of 1.5 to 2 km in the vicinity of the village. This change in form is accompanied by the appearance of less sandy, darker vertisolic soils in the valley floor, possibly due to cultivation of lands near to Boatlaname. In the village itself, the main school is situated on top of a 4 m high promontory of tabulate and irregularly flaggy red-stained terrazzo silcrete, suggesting that the gentle valley flanks may contain buried duricrust outcrops.

The Serorome at the Dibete Fence

Following the line of the Dibete veterinary cordon fence the descent into the Serorome valley begins approximately 12 km southeast of Lephephe, with the valley base reached after a further 3 km. The valley width at this point is approximately 5.5 km. The very gently sloping flanks make estimation of the valley depth difficult, but the depth must be in around 40 m at this point, with a valley side slope of less than 1°. No channel is present in the valley floor, the base only distinguishable from the flanks by the presence of slightly darker vertisolic soils. A few duricrust exposures were seen on the northern flank at heights of 15-20 m above the valley base. These were creamy-grey terrazzo silcretes with a white outer opalline weathering crust. The escarpment area indicated on figure 5.25, approximately 30 km southeast of Lephephe, was the location for silcrete sample Lephephe 100 analysed by x-ray fluorescence (section 6.2.4).

The Serorome near the Gaborone-Francistown road

Where the main road cuts the valley, the base is at an altitude of 903 m asl, with a depth of incision of between 40 and 50 m (plate 5.45). The valley in this region is 1 to 1.5 km in width, with a flat valley floor. Duricrust exposures in the valley base are rare, with two exceptions. Firstly, a pit immediately adjacent to the main road contains 1.5 m of nodular to honeycomb calcrete, with cemented nodules of relatively crumbly micritic calcite. Some cavities are present within the profile particularly within the top 40 cm. The other valley floor exposure occurs at 9.8 km west of the main road where an area of hardpan calcrete outcrops in the base of the valley. Duricrusts on the valley flanks are more common, although few good exposures occur. The flanks generally have rubbly duricrusts in their uppermost sections, with a mixture of ferricrete and silcrete present. Large areas of duricrust rubble occur to the south of the valley, approximately 8.5 to 10 km due west of the road.

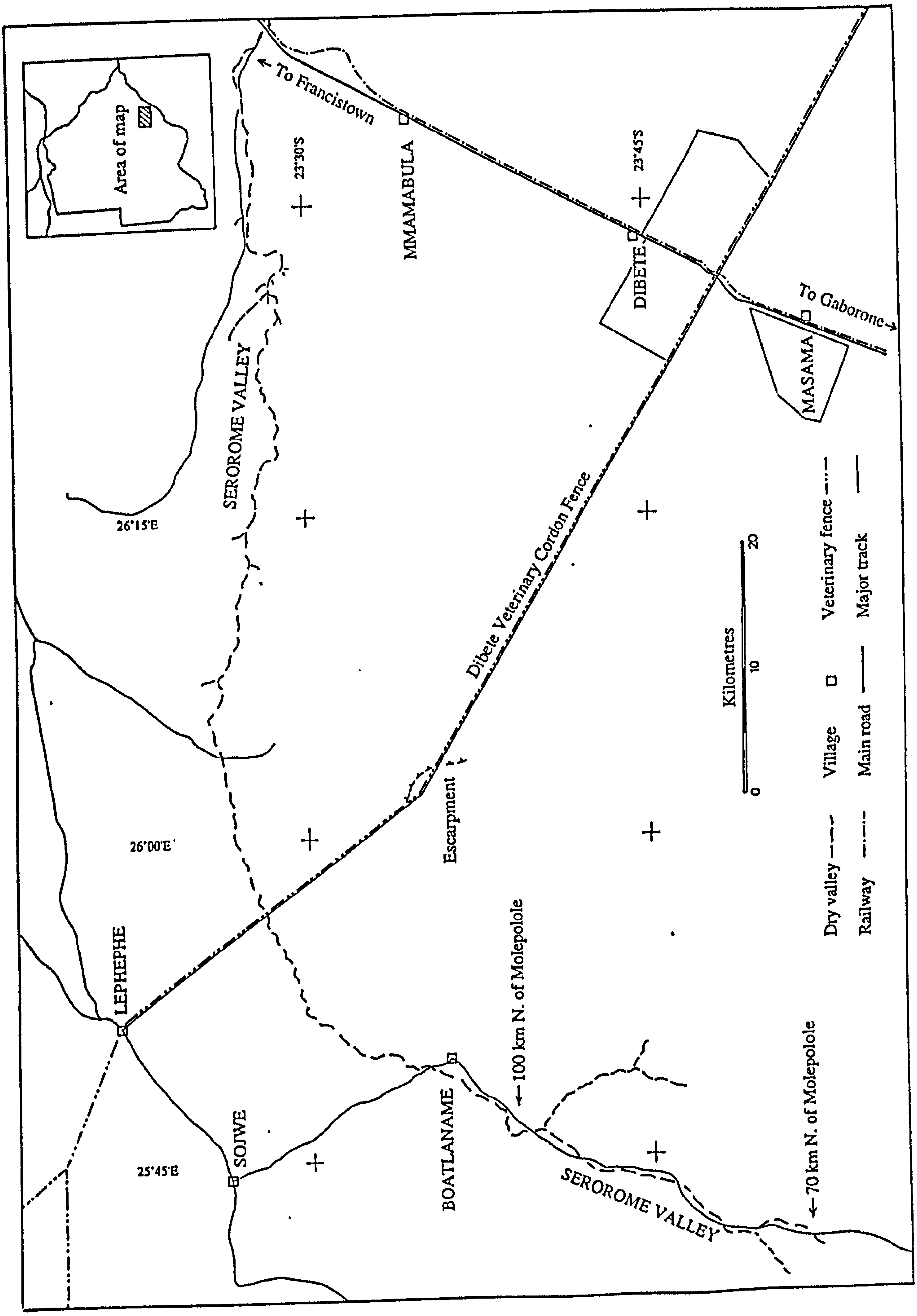


Figure 5.25: Locations along the Serorme Valley mentioned in section 5.3.4.

Plate 5.44: The Serorome Valley 70 km north of Molepolole.

Plate 5.45: The Serorome Valley where it is crossed by the Gaborone-Francistown road.

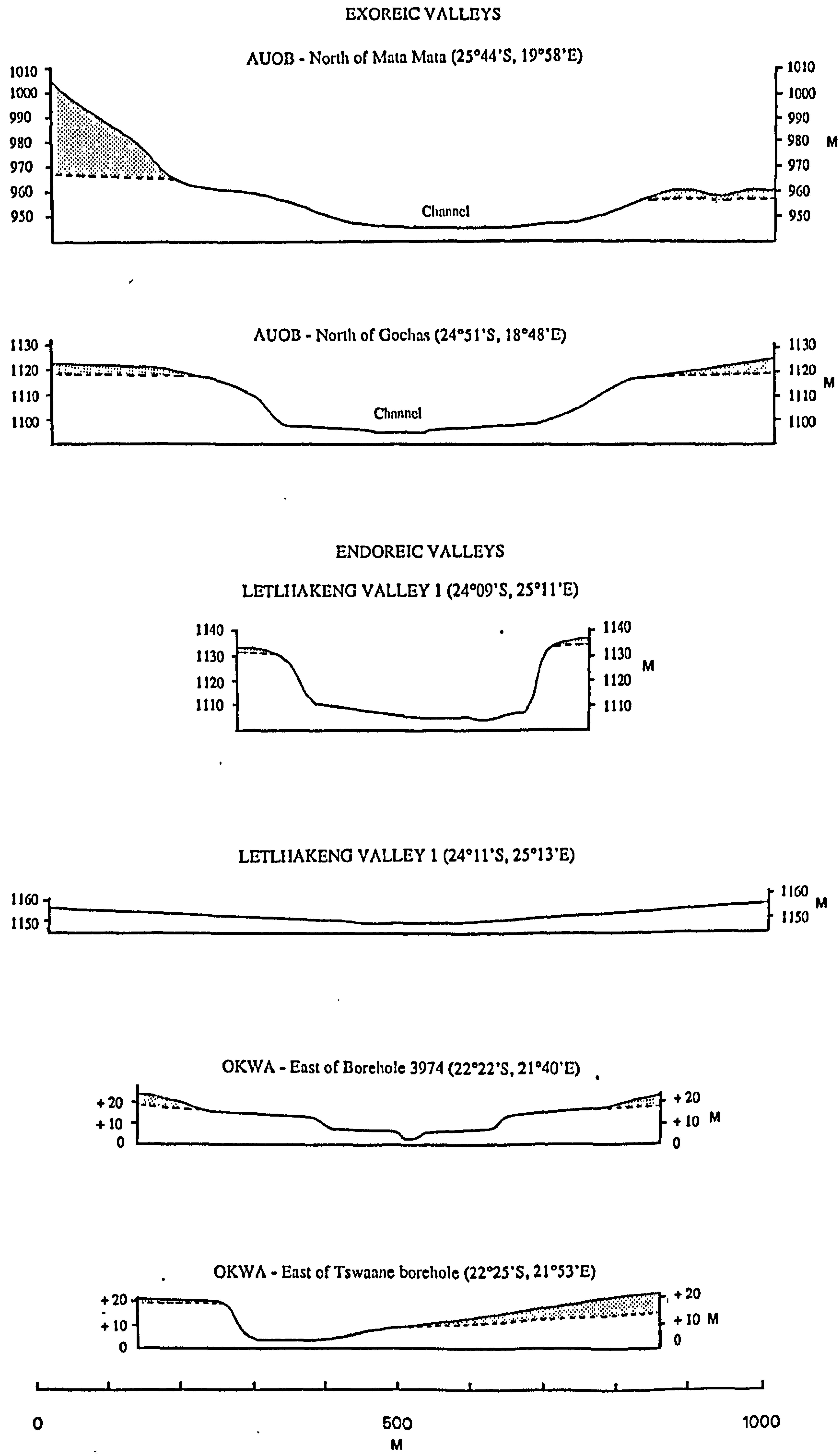


Figure 5.26: Cross-sections of selected exoreic and endoreic valleys.

5.4 Chapter summary

Field investigations and analyses of remotely-sensed imagery indicate the great variability in the form of Kalahari *mekgacha*, both within and between different systems. Such diversity of form might be expected, particularly when the great areal extent of many systems is considered; the Okwa-Mmone systems have a shared potential catchment in excess of 90,000 km^2 (Thomas and Shaw, 1991*a*). However, this variability is despite the marked homogeneity of the Kalahari Group sediments, which might be expected to produce homogenous valley forms.

Field studies confirm the general morphological characteristics identified by earlier accounts of *mekgacha* (e.g. Boocock and Van Straten, 1962), namely the three main stages of valley form present in most systems (summarised at the start of this chapter). The tripartite variation in form is particularly apparent in parts of the Mmone/Quoxo and Moselebe systems. It can also be clearly identified from studies of aerial photography for valleys such as the Xaudum to the west of the Okavango Delta. However, these general characteristics are somewhat oversimplified and do not apply equally to all networks. In particular, the Groot Laagte and Rooibrak valleys have very subdued overall forms, and do not exhibit the marked "gorge" section suggested by Boocock and Van Straten (1962) and Thomas and Shaw (1991*a*). In the case of the Rooibrak, this would appear to be largely attributable to an infill of sediment. The Okwa, Meratswe and Hanehai valleys also indicate the influence of bedrock inliers surrounded by Kalahari Group sediments along the course of a valley; within bedrock sections the courses of these valleys narrow considerably due to the width restraints placed upon them. Other valleys may show similar deviation from the idealised threefold variation in form, but were not visited in the field.

Another major feature identifiable from field studies is the difference in valley morphology between endoreic and exoreic *mekgacha* (figure 5.26). In particular, the Auob and Nossop, which rise in highland areas near to Windhoek in Namibia have much more deeply incised valleys, although the Kuruman and Molopo valleys show closer affinity in terms of form to the endoreic valleys. The role of comparatively recent fluvial activity in these systems appears to be the main influence contributing to channel form.

Field studies indicate the role of fluvial activity in shaping valley floor sediments and morphology. The most obvious sign of the action of either ephemeral or perennial rivers in *mekgacha* comes from the terrace levels identified in the Okwa, Xaudum, Kuruman and Moselebe systems. In the Okwa and Xaudum, these terraces presumably relate to changes in valley gradient due to fluctuations in base level caused by neotectonic movements associated with the Okavango Rift and Makgadikgadi Depression. The presence of valley floor sedimentary infill could, however, also be indicative of either reduced flows or an increase of sediment to valley systems. Incision to produce terraces, if developed by perennial rivers, could indicate either increased flows or decreased sediment inputs. Both the Okwa and Xaudum terrace materials are calcretised, most probably prior to a period of incision, thus maintaining a steep-sided terrace form. Terraces in the Moselebe are partly supported by silcretes, which suggests probable greater antiquity than the soft calcrete terraces in the Okwa, whilst the partially consolidated terraces along the Kuruman relate to flood events dating from the present day back at least to the Holocene. Channel patterns identified from aerial photography in the Ncamasere, Xaudum and Nossop valleys also testify to the role of fluvial activity in shaping valley floor sediments, although the channel patterns have developed upon a pre-existing floor within the confines of a larger valley. In the case of the Nossop, the last major flood to involve the entire length of the valley occurred in 1933-34, whilst no flows of such great extent have been recorded for the Ncamasere and Xaudum. Lag deposits and evidence of fluvial erosion on bedrock in the Okwa Valley also indicate periods of flow.

In contrast, field evidence for the action of groundwater in valleys is extremely limited apart from the valley head area of Letlhakeng Valley 1 where probable relict springlines occur. Even here, where the overall characteristics of the valley suggest development by groundwater sapping processes, the valley morphology suggests more recent fluvial activity which has eroded the back wall of the amphitheatre head. However, more certain evidence for the role of groundwater appears to be indicated by the broad scale structure of networks, which will be discussed in chapter 7.

Chapter 6

The relationship between duricrusts and Kalahari *mekgacha*

6.1 Introduction to duricrust analysis

Calcretes, silcretes and ferricretes, part of the duricrust suite, are rock types almost ubiquitously exposed within Kalahari *mekgacha*. Indeed, some of the main outcrops of duricrusts occur in the floors and flanks of dry valleys, which provide major topographic lows in the generally flat terrain of the Kalahari.

The relationship between *mekgacha* and duricrusts is of considerable importance in evaluating the hypotheses for valley development. As discussed in the preceding chapter, it is often assumed that Kalahari duricrusts formed prior to a period of valley incision, although some studies suggest that this may not be the case (Mabbutt, 1955; Shaw and De Vries, 1988). Studies of duricrust geochemistry and morphological variations from borehole records, outcrops and thin-sections can provide a record of the stages of, and environmental conditions during, the formation of a particular duricrust or duricrust suite. From this record of development it is possible to assess any relationship between a particular valley and its associated duricrusts. This chapter examines the possible interaction between duricrust development and valley systems in two ways. Firstly the definition, classification, mineralogy, possible models of origin and significance of duricrusts are considered in section 6.1. Secondly, the methodology and results of field and laboratory studies of duricrusts are discussed in section 6.2.

The study of duricrusts in the geomorphological and geological literature has traditionally developed by independent studies of individual duricrust types. Major studies and reviews include Reeves (1970), Goudie (1973*a*, 1983), Netterburg (1980), Watts (1980) and Wright and Tucker (1991) for calcretes, Mulcahy (1967) and Stephens (1971) for ferricretes and silcretes, and Langford-Smith (1978), Summerfield (1978, 1982, 1983*b*), Young (1985), Hesse (1989) and McBride (1989) for silcretes. However, as all three main types of duricrust as well as complex intermediate types (such as sil-calcrete and cal-silcrete) occur in association with Kalahari *mekgacha*, often even within the same profile, and all are interrelated in terms of possible origin, this chapter considers duricrusts as a whole.

6.1.1 Introduction and definitions

Much debate has been generated regarding the correct definition of a duricrust. Perhaps the best general definition is that given by Goudie (1973*a* p.5), who describes a duricrust as:

> "a product of terrestrial processes within the zone of weathering in which either iron... sesquioxides (in the case of ferricretes) or silica (in the case of silcrete) or calcium carbonate (in the case of calcrete)... have dominantly accumulated in and/or replaced an existing soil, rock or other weathered material, to give a substance which may ultimately develop into an indurated mass."

Essentially, calcretes and silcretes (terms first coined by Lamplugh; 1902, 1907) can be regarded as limestones and sandstones respectively. However, the main distinguishing characteristic of a duricrust is that formation occurred due to low-temperature physico-chemical processes at the near-surface, and not as a result of any metamorphic, igneous or diagenetic process due to deep burial (Summerfield, 1983*b*).

6.1.2 Distribution

Duricrusts are widespread over many arid and semi-arid parts of the world (cf. Goudie, 1973*a* pp.73-83), typically in warm areas with limited precipitation. Indeed, duricrusts have been reported from all continents except Antarctica although some polar examples are known (Summerfield, 1983*b*). They are known by a variety of local names; for example, calcrete is often termed *caliche* in the USA (Goudie, 1973*a* p.8), whilst silcrete occurring in Britain is known as *sarsen* or *puddingstone* (Summerfield, 1983*b* p.59).

Kalahari duricrusts are documented in a large number of references, stemming from the seminal observations of Passarge (1904). The distribution of duricrust types has been observed to follow the precipitation gradient in the Kalahari (Goudie, 1973*a*), with silcretes and calcretes dominating drier central and southern areas, and ferricretes mainly in the Northern Kalahari. However, as Thomas and Shaw (1991*a*) note, climatic conditions are not the only determinant of duricrust type, citing as evidence the abrupt boundaries in duricrust distribution at the Limpopo-Kalahari drainage divide and the presence of all three types of duricrust at some locations.

Duricrusts, though extremely widespread, do not always form conspicuous outcrops, principally because of the Kalahari's depositional setting. Studies of lithological borehole logs (e.g. Smit, 1977; Thomas, 1988*b*) reveal vast thicknesses of duricrusts beneath the surface as part of the Kalahari Group sediments. However, duricrusts are generally only exposed in topographic features such as valleys, pans and escarpments.

6.1.3 Classification of duricrusts

Duricrusts are highly variable in their composition and appearance, due to the wide range of host and cementing materials from which they are formed. In the Kalahari, host materials include bedrock, Kalahari Sand, various fluviolacustrine sediments and organic materials such as diatomaceous earths and shell beds (e.g. Netterberg, 1975, 1982; Thomas and Shaw, 1991*a*). The most common classifications, on the basis of meso- and microscale morphology are considered.

(a) Classification by morphology

Classification of duricrusts at the profile or hand-specimen scale is perhaps simplest for calcretes, with a number of schemes for classification available. The most useful of these is that shown in table 6.1, based upon Netterburg (1980) with subsequent modifications by Goudie (1983). Netterburg's scheme, with the exception of soil and powder categories, can also be applied in general to silcretes.

Table 6.1: Classification of calcrete by morphology (after Netterburg, 1980; Goudie, 1983).

Calcrete type	Characteristics and occurrence
Calcified soil	Weakly cemented horizons within soil profiles.
Powder calcrete	Fine powder with some carbonate replacement, associated with pans.
Nodular calcrete	Nodules or concretions in a matrix of carbonate; various occurrences.
Honeycomb calcrete	Nodules which have coalesced to form a honeycomb texture; various occurrences.
Hardpan calcrete	Indurated layer, often comprising cemented horizons of the above calcrete types, often found as a surface horizon or between nodular or powder calcretes; may include gravels.
Laminar calcrete	Laminated crust or layers < 25 cm in thickness, usually capping hardpan calcrete.
Boulder calcrete	Discrete to coalesced boulders of other calcrete types, often showing resolution; a secondary calcrete.

The morphological identification of silcretes is more complex since the most important differences between many silcrete types occurs at a microscopic level. Summerfield (1978, 1982, 1983*b*) chose to differentiate silcretes by the presence or absence of an associated weathering profile. Further subdivisions are dependent upon chemical and micromorphological criteria, as discussed below. The classification developed by Smale (1973) shown in table 6.2 is useful for purposes of field identification, although more detailed classification is only possible in thin-section and by chemical analysis.

McFarlane (1983) has developed a classification system for ferricretes in terms of their mineralogy and structure. However, as most Kalahari ferricretes have only a limited range of characteristics (Thomas and Shaw, 1991*a*), and were only encountered as isolated outcrops in two *mekgacha*, they will not be further considered.

Classification of duricrust profiles is often problematic, especially where a number of different types of material are juxtaposed or where diagenetic replacement has occurred within a profile. Examples of such complex profiles are described by a number of authors including Smale (1973), Goudie (1973*a*, 1983) and Arakel *et al.* (1989).

Table 6.2: Classification of silcrete by morphology (after Smale, 1973). See Table 6.3 for definition of fabric types.

Silcrete type	Silcrete characteristics
Terrazzo	Grain-supported fabric comprising approx 60% skeletal quartz grains with solutional cavities. Cement may be opal or cryptocrystalline. Conchoidal fracture common.
Conglomeratic	Pebbles of terrazo silcrete or other material. Conglomeratic fabric.
Albertina	Few skeletal grains, with a matrix of terrazzo type (usually crypto-crystalline). Floating or Matrix fabric.
Opalline	Few skeletal grains, comprising opaline, cryptocrystalline or chalcedonic silica. Massive, with a Matrix fabric.
Quartzitic	Grain-supported fabric, with an orthoquartzitic texture due to both cementation by and overgrowths of silica on skeletal grains.

(b) Microscale classification

Summerfield (1978, 1982, 1983*b*) identified four main silcrete fabric types (table 6.3), the general characteristics of which can also be applied to calcretes (with a calcite matrix instead of silica). Fabric types are based upon the quantity of skeletal grains, the amount of contact between grains and the type of matrix material (more correctly termed cement). The four main fabric types can be subdivided by means of the form of cement present and the presence/absence of glaebular structures within the duricrust (see section 6.1.5*b* below).

Calcretes can also be subdivided by microstructure with Alpha- and Beta-types representing two ends of a spectrum (Wright and Tucker, 1991). Alpha calcretes are dense micritic groundmasses containing no biogenic material, whilst Beta-types contain carbonate precipitated in association with micro-organisms.

(c) Other classification schemes

In addition, two other schemes exist for the classification of calcretes (Wright and Tucker, 1991). These are classification according to hydrological setting (under which pedogenic and groundwater calcretes can be considered as the two broad types; Carlisle, 1983) and division by mineralogy (Netterburg, 1980).

Table 6.3: Classification of silcrete by fabric type and other micromorphological characteristics (after Summerfield, 1978, 1983*b*).

Fabric Type	Fabric characteristics
GS Fabric	Grain-supported fabric. Framework of self-supporting skeletal grains. Sub-types: *a* Optically continuous overgrowths rimming detrital grains, *b* Chalcedonic overgrowths (usually optically length-fast), *c* Cryptocrystalline and opaline silica and microquartz mix filling interstitial voids.
F Fabric	Floating fabric. Skeletal grains (> 5% of the silcrete content) are not self-supported and float in the matrix. Grain dissolution and fretting may be present. Sub-types: *a* Massive (glaebules absent), *b* Glaebular (glaebules present).
M Fabric	Matrix fabric. Skeletal grains < 5% of the silcrete content, in a microquartz or cryptocrystalline cement. Sub-types: *a* Massive (glaebules absent), *b* Glaebular (glaebules present).
C Fabric	Conglomeratic fabric. Skeletal grains include detrital sediment > 4 mm which may include fractured bedrock, gravel or other duricrust fragments.

6.1.4 Geomorphological relations of duricrusts

Mapping of duricrust distributions in the Kalahari by Summerfield (1982 p.40) using information from lithological boreholes suggested that duricrusts are apparently best developed in the vicinity of pans and valleys. This may be due to the tendency to locate boreholes close to these features, although there is evidence to suggest that duricrust development and certain landforms may be genetically linked (e.g. Lawrance and Toole, 1984; Shaw and De Vries, 1988). In addition to pans and watercourses, Goudie (1973*a*) also notes the occurrence of calcretes underlying interdune *straats*.

Summerfield (1982) notes four main types of silcrete occurrence; totally or partially silicified calcrete profiles, silicified pan margin sediments, fluvial deposits and terraces, and silicified sands. Thomas and Shaw (1991*a*) add the silicified spring tufas of the Letlhakeng and Serorome valleys to this list. Calcretes have a similar range of occurrences, being associated with pan rims, valleys and in soil profiles (see section 6.1.6*d*).

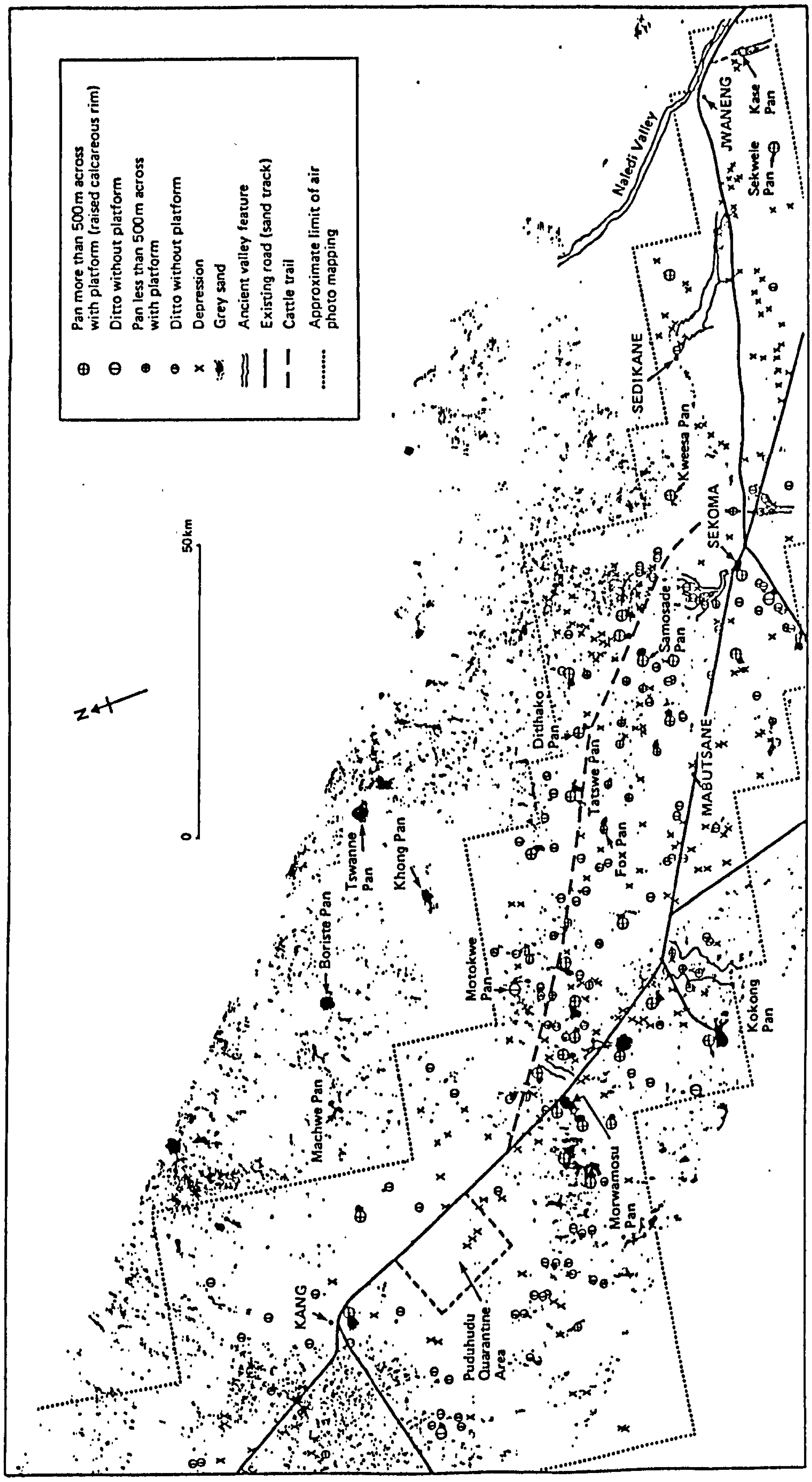


Figure 6.1: Calcrete-bearing landforms and features between Takatshwaane and Morwamosu Pan, Botswana (from Lawrance and Toole, 1984).

In the Kalahari, the most extensive surface suites of duricrusts are those associated with the Okavango Delta and Makgadikgadi Basin (section 3.4), described by a variety of authors (e.g. Cooke, 1977, 1980; Heine, 1978*a*; Shaw and Cooke, 1986). In Sua Pan, silcretes occur on and within the pan surface with calcretes commonly found around the pan edges, often cementing alluvial deposits. Summerfield (1982) records increasing levels of SiO_2 in channels towards the distal end of the Okavango Delta, which may partly explain the increased prominence of silcretes in the Thamalakane and Boteti Rivers (Thomas and Shaw, 1991*a*).

From remote-sensing and field studies in the Jwaneng-Kang-Lone Tree areas, Lawrance and Toole (1984) note a relationship between calcrete type and landform (figure 6.1). Highly indurated calcrete deposits were found in pans with a well developed raised pan rim or "platform", with the probability of finding aggregate-quality calcrete decreasing for pans without a platform, depressions and grey sand areas.

In other non-depositional environments away from the Kalahari, duricrusts are often much more conspicuous, acting as cap-rocks and thus controlling slope evolution (Goudie, 1973*b*; Hagedorn, 1988). For example, silcrete in the South African Cape coastal zone exists mainly as cappings, representing a dissected residual land surface (Summerfield, 1983*d*). In Australia, ferricretes commonly occur as caprocks due to relief inversion (Ollier, 1991*b*). Such cap-rocks can act as a source material for the development of other duricrusts (section 6.1.6*d*).

6.1.5 Mineralogy, chemistry and micromorphology

(a) Chemistry and mineralogy

The precise chemical and mineralogical definition and classification of particular duricrust types is not simple, due mainly to the wide range of materials under consideration. There is little agreement regarding the arbitrary distiction between, for example, a calcrete and sil-calcrete. Table 6.1 (after Netterberg, 1980; Goudie, 1983) shows that calcrete alone varies from essentially calcium carbonate-cemented soil to an extremely indurated bouldery form, with associated variations in $CaCO_3$ and silica content.

(i) Calcretes

Studies of Kalahari calcrete grab samples in the Serorome valley of eastern Botswana by Gwosdz and Modisi (1983 p.82) show $CaCO_3$ and SiO_2 contents varying from 24.6 to 87.1% and 7.2 to 65.3% respectively. Corresponding figures (Gwosdz, 1981, 1982; Gwosdz and Modisi, 1983 p.149-150) for calcretes from boreholes in Shaw and De Vries' (1988) Valley 2 south of Letlhakeng, Botswana, show $CaCO_3$ values ranging between 28.5 and 94.1% and SiO_2 content from 3.9 to 67.5%. As Gwosdz and Modisi (1983 p.86) note, the dominant feature of calcrete mineralogy appears to be the strong antipathy between calcium carbonate and silica, a factor which allows for the range of intermediate duricrust types, such as sil-calcretes and cal-silcretes. Summerfield (1982 p.48) quotes a SiO_2 content of between 85 and 95% as the arbitrary boundary line for a cal-silcrete.

Goudie (1972*a*; 1973*a* p.18-19) has calculated a global mean calcium carbonate content of 79% from his study of calcretes in a number of countries, with values ranging from 27 to 99%. Other constituents of calcrete include iron and aluminium sesquioxides (commonly less than 1% concentration) and magnesium carbonate and magnesium oxide (from 1.5-20% and 0.5-13% respectively). The magnesium content of a particular sample varies with the dominant carbonate type present; low MgO and $MgCO_3$ when calcite is the main carbonate; higher values if dolomite is present. Higher levels of dolomite have been recognised in calcretes developed in Mg-rich host materials (Watts, 1980) and from areas with elevated groundwater salinities (Mann and Horwitz, 1979; Watts, 1980). Manganese oxide and various clay minerals may also be present (Watts, 1980; Goudie, 1983). In addition, concentrations of the uranium-bearing mineral carnotite (with up to 102 ppm uranium) have been noted from calcretes in many Kalahari *mekgacha* (Union Carbide, 1980*d*) as well as in valley calcretes from Western Australia (Carlisle *et al.*, 1978).

(ii) Silcretes

In terms of chemistry and mineralogy, silcretes are comparatively simple, highly siliceous materials, with Summerfield (1978, 1982) recording silica levels in the range 92-98% for Kalahari silcrete samples. Iron, magnesium, titanium and aluminium oxides are also present in small quantities, as is calcium carbonate. Mineralogy reflects both the host material properties and subsequent diagenesis. Quartz is the common form of silica present within silcrete, mainly in the form of skeletal grains where present, with microquartz, chalcedony, and opalline silica in its various forms commonly forming the matrix cement (Summerfield, 1982; Chadwick *et al.*, 1987*b*). Studies in the Paris Basin by Thiry and Millot (1987) identify that silica precipitation starts with amorphous opalline forms, with subsequently more crystallographically organised forms (in the order chalcedony and lucetite, microcrystalline quartz and finally megaquartz) being precipitated. Summerfield (1983*c*) attributes this sequence to a temporal decrease in the rate of movement of silica-rich waters through the host material. In pedogenic silcrete profiles, the latter stages of this sequence are more commonly found within the upper sections of profiles.

(b) Micromorphology

When viewed in thin-section, duricrusts display a variety of micromorphological features, many of which have significance in the identification of their history of development. These are, again, in part inherited from the original host material but are also a result of near-surface diagenesis (Smale, 1973; Summerfield, 1978). Summerfield (1983*b*) subdivides these features into characteristics of the duricrust fabric and specific diagenetic features, to which should be added characteristics derived from the host material. These will be discussed in turn.

(i) Fabric characteristics

The principle variations in duricrust fabric are shown in table 6.3 (after Summerfield, 1978, 1983*b*). This scheme, although designed specifically for silcretes, is also broadly applicable to calcretes, with calcium carbonate replacing silica as the matrix material.

GS- (grain supported) fabrics are typical of cemented riverine and aeolian sands, with the sand grains forming a framework within which cement is precipitated by a process of passive void-filling (for uncemented sands) or replacement of a pre-existing matrix (Summerfield, 1983*b*). The cement may take a micro- or macro-crystalline form, dependent upon the presence or absence of coatings upon skeletal grains (Heald and Larese, 1974; McBride, 1989), the chemical constitution of porewaters and the time allowed for precipitation to occur (Summerfield, 1983*b*; Thiry and Millot, 1987). Coatings on quartz grains are generally thought to inhibit optically continuous macro-quartz and megasparite overgrowths (Heald and Larese, 1974), whilst porewater impurities cause disordered crystal growth (Thiry and Millot, 1987).

F- (floating) fabrics consist of skeletal grains "floating" within a cement or matrix, whilst M- (matrix) fabrics have less than 5% skeletal grains. Such fabrics can be formed by either displacement and/or replacement of the skeletal grains by the cementing agent (Goudie, 1983). In both cases, grains often show signs of surface re-solution, particularly where replacement has occurred. Field evidence for displacement as a significant mechanism in the formation of silcretes is lacking (Summerfield, 1983*b*), although the development of pseudo-anticlines and other displacive features in calcrete profiles is well documented (e.g. Blank and Tynes, 1965; Watts, 1977). M-fabrics are most likely to be a result of replacement of existing F-fabric cements (Summerfield, 1983*b*) or of lithologies such as shale with a small average quartz grain size.

C- (conglomeratic) fabric duricrusts typically consist of clasts of either bedrock, gravel or brecciated duricrust greater than sand size in an F-fabric matrix (Summerfield, 1982).

(ii) Features inherited from host material

Where a duricrust has developed from or replaced pre-existing bedrock, some features of the original host material may be retained. For example, Goudie (1983 p.103) includes photography of a micro-brecciated calcretised schist in thin-section, which clearly retains its schistose fabric.

(iii) Diagenetic features

A number of features are found in duricrusts which relate to diagenetic processes operating in a near-surface environment. These include laminar features, concentric structures (or glaebules), colloform features, complex vugh- (or void-) fills and replacement features.

Laminar features.

A common feature of the exposed surface of duricrusts is an outer rind of hard, often laminated material, which contains comparatively few skeletal grains (Wright, 1989). Such laminar features consist of alternating layers of light and darker brown authigenic microcrystalline calcite, opal or chalcedony (Walls *et al.*, 1975; Goudie, 1983; Thiry and Millot, 1987). The banded appearance is attributed to differential staining by iron and manganese oxides, and may also occur as linings along solutional pipe structures in duricrusts (Goudie, 1983). Biogenic calcretes often exhibit laminar structures.

Concentric or glaebular structures

Glaebular features up to several centimetres across are common in calcretes and also in F- or M-fabric silcretes associated with weathering profiles (Goudie, 1973*a*; Summerfield, 1983*b*,*d*). Where concentric features are found in non-weathering profile silcretes, it is likely that they are inherited from nodular calcretes which have been replaced by silica. The exact method of formation of concentric structures is equivocal, but must be due to centripetal enrichment (Summerfield, 1983*d*). In calcretes, such structures may or may not have central nuclei around which calcite has been precipitated (Siesser, 1973). Goudie (1983 p.105) describes three types of concentric structure; spherical pellets of authigenic micrite lacking a nucleus; ovoids comprising a nucleus with one or more laminae enclosing it; and pisolites which are essentially ovoids greater than 2 mm in diameter. Glaebules in silcretes may also have an undifferentiated internal structure, and, in the case of weathering profile silcretes, commonly have higher concentrations of iron and TiO_2 (in the form of anatase) than the surrounding matrix (Summerfield, 1983*c*).

Colloform features

These are cusp-like laminations found in weathering-profile silcretes, mainly those with F- and M-fabrics, commonly containing concentrations of anatase (Summerfield, 1983*b*). Colloform features usually consist of vertically-stacked, concave-upward cusps, although horizontally extended forms also occur (Summerfield, 1983*c*). Such features are apparently a result of rhythmic geopetal late-stage infilling of voids, or occasionally pedotubules (Summerfield, 1983*b*,*d*), although this origin is equivocal.

Vugh-fills and replacement features

Many duricrusts show late-stage, often complex, void-filling by silica and/or calcium carbonate. Such vugh-fills are of great significance in establishing the history of development of a particular duricrust, indicating the chemical composition of porewaters during the late stages of cementation (Summerfield, 1983*c*). Studies of silcious vugh-fills in silcretes are the best documented (e.g. Summerfield, 1978, 1982, 1983*b*,*c*; Milnes and Thiry, 1986; Thiry and Millot, 1987), although calcite infills in silcretes (Summerfield, 1982, 1983*c*) and silicification of calcretes (Arakel *et al.*, 1989; Arakel, 1991) have been noted.

The most common inward sequence of siliceous infill from the walls of a vugh in silcrete is opal-chalcedony-microquartz-megaquartz (Summerfield, 1982). There is also a tendency for crystal size to increase towards the centre of the vugh, especially where megaquartz is present at the vugh centre. This sequence reflects both the changing composition of porewaters as various forms of silica come out of solution, but also the declining rate of porewater flow as voids are progressively infilled and porosity is reduced (Summerfield, 1983*c*). Detailed mineralogical studies by Thiry and Millot (1987) make further distictions between the order of crystallisation of different forms of opalline silica, with opal-A crystallising before opal-CT. Chalcedony in vugh-fills can occur in optically length-slow and length-fast forms (the crystal growing parallel or perpendicular to the mineral's epsilon optical vibration direction respectively), with the former being common in association with carbonates and in alkaline environments (Watts, 1980). Where a vugh-fill contains alternating length-slow and length-fast chalcedony, Summerfield (1983*c*) suggests that this may reflect localised changes in pH at the time of precipitation.

In calcretes, diagenetic sequences of siliceous infill and replacement of calcite have been shown to vary in relation to the position of the regional water table (Arakel *et al.*, 1989). In the vadose zone, infill is most common in porous and brecciated calcretes, with chalcedonic and opalline silica filling veins and lining pores. Replacement and displacement of calcite may also occur in this zone to give a brecciated appearance. The most extensive silicification within the zone where groundwater levels fluctuate, where some calcrete sections were observed to contain 30% silica by weight (Jacobson *et al.*, 1988). In this zone, silica precipitation in voids alternates between bands of fibrous chalcedony and thin layers of opalline silica, producing a layered cement with spherulitic chalcedony commonly filling voids. The layered void-fill sequence may be due to either fluctuations in the water table or inputs of fresh water, both of which would temporarily change pH levels (Arakel *et al.*, 1989). Replacement processes are most common in the phreatic zone, although opalline and length-fast fibrous chalcedonic void fills do occur. Replacement of microspar calcite cement by diagenetic chalcedony produces chalcedonic spherulites. In places, these spherulites coalesce to produce silcrete with a floating fabric containing areas of micrite in a silica matrix. Such replacement may progress until total coalescence occurs, with subsequent silicification in voids producing a fully indurated silcrete (Arakel *et al.*, 1989). Late-stage diagenetic processes may ultimately lead to the microcrystalline replacement of void-linings and void fills, partially or totally destroying earlier diagenetic fabrics (Arakel, 1991).

6.1.6 Duricrust formation

Prerequisites for the formation of any duricrust (in this case silcrete or calcrete) are a source of silica or calcium carbonate, the availability of moisture to act as a transporting agent from the source to the site of precipitation and a mechanism to cause precipitation. However, the exact amount of moisture necessary and the climatic significance of this, along with the actual source and mode of precipitation for a particular duricrust is still a matter of debate. When considering the environmental conditions necessary for duricrust formation it is essential to note that *two* environments need to be taken into account; one where silica or carbonate is produced and another where it is precipitated (Ollier, 1991*a*). Sources of precipitate, mechanisms of precipitation and the environmental conditions required for duricrust development are now considered.

(a) Precipitate sources

A number of potential sources of material exist for duricrust development, which have varying significance for different duricrust types. These include chemical breakdown of silicates within bedrock, direct dissolution of quartz grains, lakes and pans, vegetation and leaf litter, micro-organisms, atmospheric dust, rainfall, surface runoff, and groundwater, all of which can act as either local or non-local sources (Goudie, 1973*a*, 1978, 1983; Summerfield, 1978, 1982; Wright and Tucker, 1991).

The breakdown of silicates in bedrock is of greatest importance as a source of silica for the development of silcretes associated with weathering profiles (Summerfield, 1981). In the absence of a weathering profile, direct dissolution of quartz grains or grain abrasion may be more important. The formation of dew (Summerfield, 1982) and occurrence of rainfall may be of significance for grain dissolution.

Silica is also released during the replacement of quartz by calcite (Summerfield, 1982), as a result of the inverse solubility ranges of quartz and calcite in alkaline environments (silica is soluble above pH 9). Lake and pan environments can also produce locally alkaline conditions conducive to the dissolution of quartz (Summerfield, 1978). Vegetation litter (with the exception of $CaCO_3$ fixing species) may have a converse effect upon porewater pH, especially in weathering environments, giving rise to highly acidic conditions conducive to silicate weathering (Summerfield, 1983).

Certain types of vegetation, as well as organisms such as diatoms (Passarge, 1904; Du Toit, 1954) and micro-organisms (Thomas and Shaw, 1991*a*; Shaw *et al.*, 1991), have a localised role in duricrust development, acting as sources of silica or carbonate. The role of organisms such as carbonate-fixing algaes (cf. Lancaster, 1977; Walker, 1973*b*), trees, shrubs and mollusca as sources of $CaCO_3$ is better documented (see Goudie, 1983 and Thomas and Shaw, 1991*a* for summaries). Less localised sources of material for precipitation come from atmospheric dust inputs (Goudie, 1978), ground- (Kirchner and Tredoux, 1975) and surface-waters (Summerfield, 1982 p.57).

(b) Mechanisms of precipitation

A number of possible mechanisms exist for the precipitation of silica and calcium carbonate. Considering silica precipitation firstly, Siever (1962) and Summerfield (1983*b* p.76) note that evaporation, cooling, cation reaction, adsorption by solids, the neutralization of strongly alkaline solutions and the action of certain organisms can all lead to the removal of silica from solution. Of these possible mechanisms, evaporation, changes in pH, the effects of biotic life processes, cooling and reaction with cations are considered the most important in duricrust genesis.

The relatively slow rate of change of silica solubility with temperature suggests that cooling may not be a major factor at earth-surface temperatures. Indeed, the rate of reaction of silica to supersaturation is considerably slower than rates of change in temperature (Summerfield, 1983*b*). Silcrete formation by evaporation would also be restricted to surface examples only, and cannot be used to explain silcrete lenses within calcrete nor thick silcrete profiles (Summerfield, 1982).

The main control of silica precipitation, and therefore silcrete genesis, is any change in pH. The species of silica precipitated and in solution is, however, highly dependent upon the area of adsorptive surfaces and the ionic strength of solutions (Chadwick *et al.*, 1987*a*). A pH of greater than 9 is required for silica dissolution, and such alkaline conditions are mainly encountered in pan and lacustrine environments in the Kalahari. Summerfield (1982) notes that the dominant mode of silcrete development appears to be the replacement of a host material, rather than direct precipitation. In a pan environment, the silicification of pan clays is the most likely mode of silcrete genesis, although Summerfield (1981, 1982 p.60) recognises possible problems with such clay replacement, particularly concerning the effects of NaCl and Al upon silica solubility. It has also been suggested that cation reactions during the mixing of silica-rich groundwater with downward percolating NaCl-rich waters may lead to silica precipitation (Frankel and Kent, 1938; Smale, 1973; Watson, 1989). However, Arakel *et al.* (1989) suggest that silica petrogenesis may be, above all, dependent on the texture and amount of impurities within the host material.

Compared with silica precipitation in near-surface environments, mechanisms of calcite precipitation are well understood. Calcrete formation results from the solution and precipitation of carbonate by means of the following simplified reaction (Goudie, 1983 p.112):

$$CO_2 + H_2O + CaCO_3 \Leftrightarrow Ca^{2+} + 2HCO_3^-$$

Following the reaction to the right, dissolution of $CaCO_3$ occurs, which results from either low pH, decreased temperature or increased CO_2 partial pressure. Precipitation of $CaCO_3$ occurs (reaction proceeds to the left) due to evaporation, decrease in CO_2 partial pressure, freezing, the addition of Ca^{2+} by the common ion effect or by biological activities (Goudie, 1983, Chadwick *et al.*, 1987*a*). Evaporation of near-surface porewater can cause precipitation by two methods; evaporation of vadose waters can lead to rapid carbonate saturation and hence precipitation, and CO_2 may be lost as a result of decreased pore pressure (Goudie, 1983; Wright and Tucker, 1991).

The interaction between solution and precipitation of carbonates and silica is highly dependent upon local levels of pH; as noted, the presence of high levels of carbonate in solution can lead to silica dissolution (Garrels and Christ, 1965), whilst temporary lowering of pH can cause $CaCO_3$ precipitation. This reaction is responsible for the wide range of possible materials within the calcrete-silcrete duricrust suite. If, however, groundwaters are saturated with respect to SiO_2, precipitation of silica cements may occur without a major change in the local chemical environmental conditions.

(c) Environmental parameters for duricrust formation

A number of environmental factors are apparently necessary for duricrust development, although none are unequivocal. Duricrusts are invariably associated with minimal local relief (e.g. Stephens, 1971; Goudie, 1983; Summerfield, 1978), with low rates of erosion and a relatively stable surface also desirable to allow the slow process of precipitation to proceed (Summerfield, 1982). The degree of drainage necessary is a matter of debate; a well drained profile may have excessive leaching and hence hinder duricrust development, whilst very poor drainage may prevent porewater movement.

The importance of climate in duricrust development has been noted by almost every author on the subject, with a general agreement that most duricrusts form under semi-arid to arid climates (section 6.1.7). Many authors even specify the degree of aridity required for a particular type of duricrust to form. For example, Bond (1963) proposed 250 to 300 mm pa for Kalahari silcrete formation and Goudie (1983) suggested 400 to 600 mm pa for calcretes. Strong seasonal temperature and precipitation variations are usually considered important, although, as noted above, diurnal variations leading to dew formation may be as important in certain environments. Goudie (1973*a*) notes that Kalahari duricrust types follow the precipitation gradient in general. Thus, ferricretes are more common in the wetter Northern Kalahari, with silcretes and calcretes occurring in the Kalahari core. However, as will be seen from the following section, other local environmental factors, at a variety of scales, can be equally important in determining duricrust development.

Finally, it should be noted that whilst virtually all studies of Kalahari duricrusts consider them to be comparatively old, the environmental parameters discussed above are derived from modern conditions. Unless the present-day formation of duricrusts can be proven (only Shaw *et al.* [1991] have documented evidence for contemporary development), all attempts at defining parameters necessary for formation are at best suspect. This makes the possible palaeoclimatic significance of duricrusts, a common theme in the literature, difficult to assess (section 6.1.7).

(d) Models of duricrust formation

A number of different models have been proposed for duricrust genesis (see Goudie, 1983; Summerfield, 1983*b*; Thomas and Shaw, 1991*a* for summaries). These can be essentially grouped into those involving vertical transfer of material in solution (also termed "relative accumulation" or "pedogenic" models), those where material in solution is transported by lateral movement of water (termed "absolute accumulation" or "non-pedogenic" models) and those involving biogenic accumulation.

Duricrusts in the Kalahari are, however, almost certainly polygenetic (Thomas and Shaw, 1991*a* p.79), and the significance of different models varies spatially. As a result, it is often difficult to distinguish a single main mode of formation at any one site. Factors contributing to duricrust development act at a variety of scales (Summerfield, 1983*b*). The actual precipitation of silica or calcium carbonate depends upon microscale geochemical conditions, but these in turn depend upon mesoscale porewater movements. Such movements are themselves reliant upon larger scale hydrologic and often geomorphological controls, with climatically, tectonically and anthropogenically related water table levels also responsible.

The role of micro-organisms in duricrust development is reviewed by Wright (1989) and will not be considered here. Goudie (1983 p.114) mentions another possible duricrust origin, that being the mechanism leading to the formation of reconstituted and detrital calcretes (transportation and subsequent cementation of weathered calcrete fragments). These possible origins for duricrust development are comparatively site specific and can account for only certain occurrences. Therefore the remainder of this section will discuss vertical and lateral transfer models more fully, these being of more universal significance.

(i) Vertical transfer models

Mechanisms of duricrust development involving relative accumulation due to predominantly vertical movement of silica and $CaCO_3$ enriched soil- or ground-water can be subdivided into *per ascensum* and *per descensum* models.

Per ascensum models of duricrust formation involve the mechanisms of capillary rise and subsequent evaporation of moisture from saturated solutions, leading to the precipitation of material at or near the ground surface (Goudie, 1973*a*). The main criticism levelled at this model is that once material has been precipitated at the surface and has formed a crust, this prevents the further evaporation of soil waters (Summerfield, 1983*b*). It is unlikely to be applicable to the Kalahari, where the extensive sand cover in many areas is liable to inhibit capillary rise (Summerfield, 1982).

Per descensum models use the downward percolation of saturated waters as the agent for duricrust formation (Stuart and Dixon, 1973). Water table fluctuations are regarded by Smale (1973) as a requirement for silcrete genesis, with the meeting of percolating low pH or salt and/or oxide charged solutions and upward rising groundwater leading to precipitation at the interface zone. Summerfield (1983*b*) concludes that a combination of the downward percolation of silica combined with water table fluctuations is the most likely model of genesis for South African weathering profile silcretes.

(ii) Lateral transfer models

In addition to the more "traditional" vertical transfer models for duricrust development, the lateral transfer of material in solute form by geomorphic agencies may also be a significant formative mechanism. Lateral transfer models envisage absolute accumulation of material, and can be subdivided into lacustrine/pan models and fluvial/sheet flood or groundwater models. Calcareous tufa formation will also be considered at the end of this section.

Lacustrine/pan models

These are most applicable to pan duricrusts although the precise mechanism of duricrust formation involved is uncertain (Summerfield, 1982). Under present-day conditions pans often provide the only standing water within Kalahari *mekgacha*. As such, duricrust formation in a pan environment may be of great significance when considering duricrusts associated with valleys. The conditions experienced in a pan environment i.e. the seasonal availability of water which can be subsequently evaporated and the often highly variable pH, are generally conducive to mineral mobilisation and duricrust development (cf. Jacobson *et al.*, 1988). Coates *et al.* (1979) record silcrete infilling lacustrine mud-cracks in the Makgadikgadi Basin of Botswana. The occurrence of silica-fixing diatoms (Passarge, 1904; Du Toit, 1954) and shell material can act as a potential duricrust source.

Fluvial/Sheet flood models

These were first suggested as mechanisms for silcrete development in Australia (see Stephens, 1971), where the presence of siliceous hardpans was noted in areas of periodic flooding. The association of calcrete development and fluvial processes has also been documented (e.g. Carlisle *et al.*, 1978; Mann and Horwitz, 1979; Netterburg, 1975, 1982; Arakel, 1986; Arakel *et al.*, 1989).

Duricrust accumulation by fluvial processes can involve deposition within channels or valleys, deposition from sheet floods, and/or lateral seepage of groundwater and throughflow water (Goudie, 1983). Valleys, like pans, are potentially important sites for duricrust development since water tables are generally closer to the surface in the vicinity of topographic depressions. Discussing calcretes formed within valleys, Mann and Horwitz (1979) distinguish "vadose" (pedogenic) from "phreatic" (non-pedogenic) calcretes, terming the latter groundwater calcretes (after the nomenclature of Netterberg, 1969*b*). Vadose calcretes are usually viewed as forming by processes such as evaporation at the capillary fringe (Goudie, 1983). Mann and Horwitz propose the following four stage model (figure 6.2), for groundwater calcrete development (1979 pp.301-2):

(A) A precursor to the development of a groundwater calcrete is a broad, alluvially filled drainage line with a shallow groundwater system.

(B) Carbonate ions are transported and precipitated at or below the water table where the solubility of calcite is exceeded.

(C) Precipitation proceeds in the phreatic zone, forming consolidated pods and domes of calcrete, which may form surface mounds by displacement.

(D) As calcrete forms beneath the water table, older calcrete is displaced and pushed upwards, where hardening and silicification can take place. Secondary processes such as carnotite emplacement may also occur at this stage, as may surface reworking.

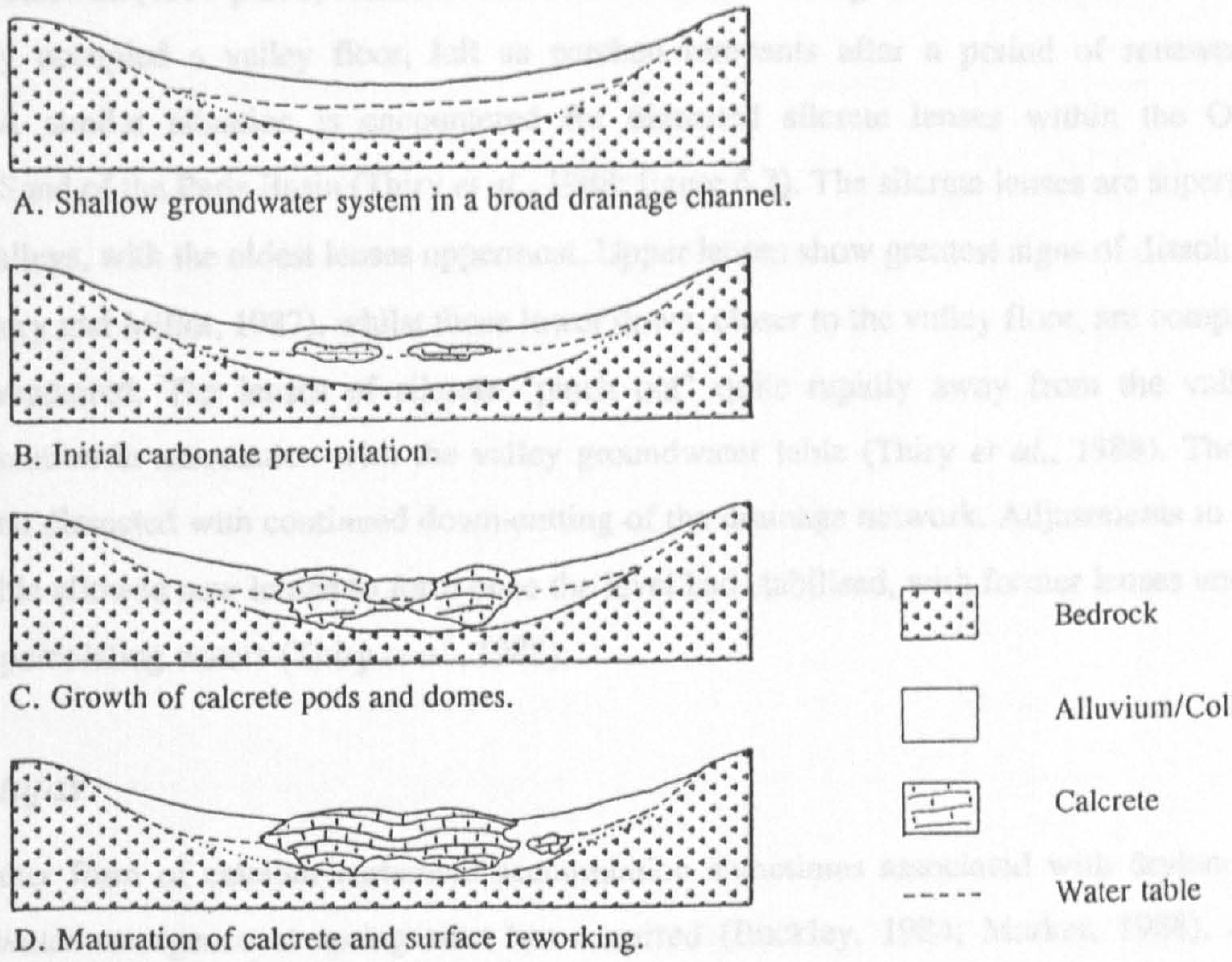


Figure 6.2: The stages in the development of groundwater calcrete (after Mann and Horwitz, 1979).

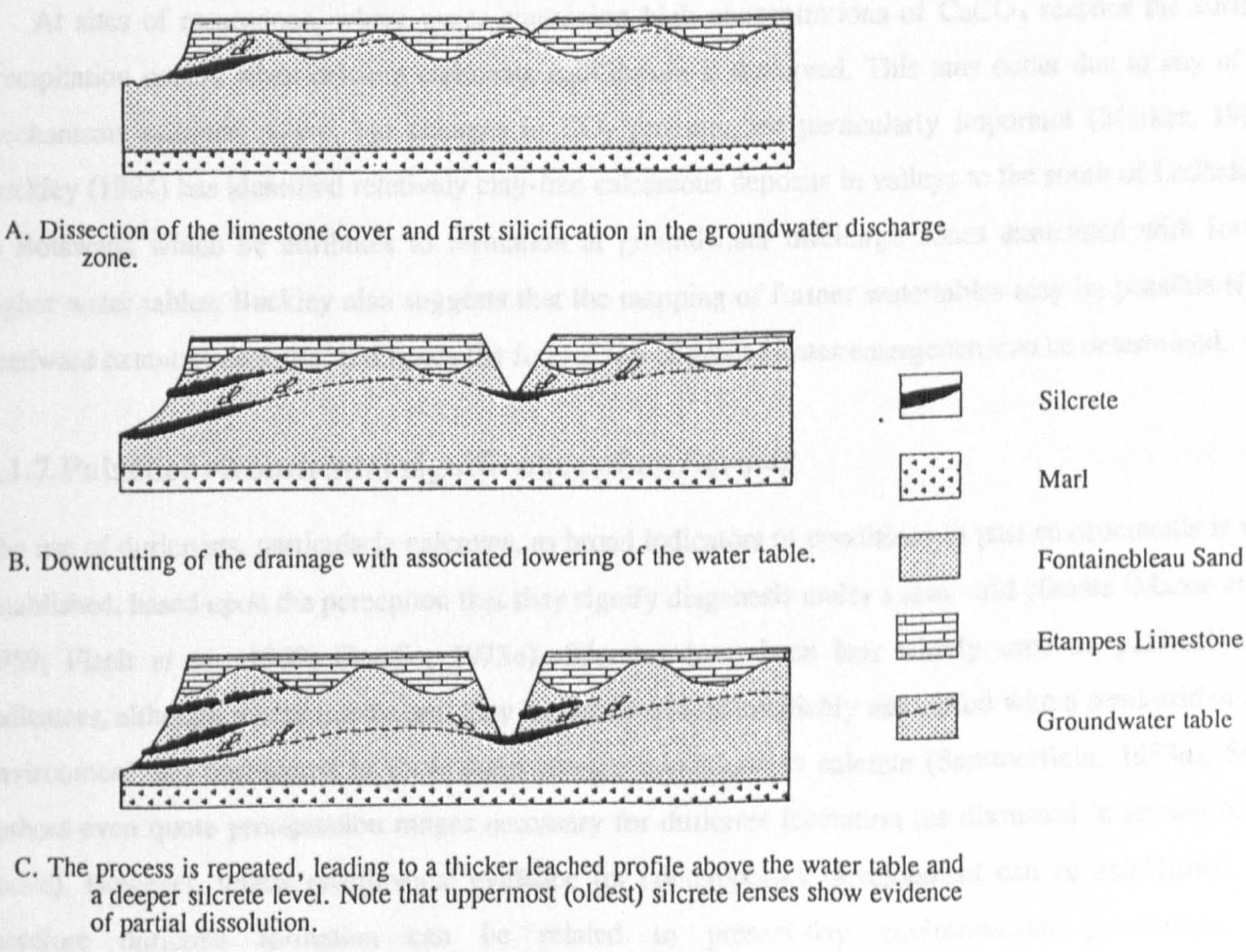


Figure 6.3: Successive stages of quartzite development in the Fontainebleau Sand of the Paris Basin. In the diagram, the uppermost quartzite is the oldest and shows most signs of dissolution and weathering, whilst the youngest, freshest lens is found lowest in the valley (after Thiry *et al* (1988).

Mann and Horwitz (1979 p.296) record the occurrence of dissected groundwater calcretes in Australia which formerly occupied a valley floor, left as perched remnants after a period of renewed valley downcutting. A similar situation is encountered for dissected silcrete lenses within the Oligocene Fontainebleau Sand of the Paris Basin (Thiry *et al.*, 1988; figure 6.3). The silcrete lenses are superposed up the flanks of valleys, with the oldest lenses uppermost. Upper lenses show greatest signs of dissolution and weathering (Thiry and Millot, 1987), whilst those lower down, closer to the valley floor, are comparatively fresh and unweathered. The lenses of silcrete "pinch out" quite rapidly away from the valley side, suggesting formation in association with the valley groundwater table (Thiry *et al.*, 1988). The silcrete lenses have been dissected with continued down-cutting of the drainage network. Adjustments to the level of the water table allowed new lenses to form once the level had stabilised, with former lenses undergoing dissolution by percolating waters (Thiry *et al.*, 1988).

Calcareous tufas

Tufas are another form of calcium carbonate accumulation sometimes associated with dryland valleys where groundwater emergence at spring sites has occurred (Buckley, 1984; Marker, 1988). Southern African tufas include deposits formed at springs and resurgences, waterfalls and in turbulent streams, with Marker (1988) providing a thorough review of recent literature. Whilst not strictly classifiable as tufas (Goudie, 1986*a*), valley calcretes are often cited as tufaceous deposits.

At sites of resurgence, where water containing high concentrations of $CaCO_3$ reaches the surface, precipitation occurs when calcium carbonate equilibrium is disrupted. This may occur due to any of the mechanisms outlined above, but changes in CO_2 pressure are particularly important (Marker, 1988). Buckley (1984) has identified relatively clay-free calcareous deposits in valleys to the south of Letlhakeng in Botswana which he attributes to formation at groundwater discharge zones associated with former higher water tables. Buckley also suggests that the mapping of former watertables may be possible if the headward extent of deposits, and hence the former site of groundwater emergence, can be determined.

6.1.7 Palaeoenvironmental significance of duricrusts

The use of duricrusts, particularly calcretes, as broad indicators of conditions in past environments is well established, based upon the perception that they signify diagenesis under a semi-arid climate (Mason *et al.*, 1959; Flach *et al.*, 1969; Goudie, 1973*a*). Silcretes have been less widely used as palaeoclimatic indicators, although where referenced, they are again almost invariably associated with a semi-arid or arid environment and considered to form under similar conditions to calcrete (Summerfield, 1983*a*). Some authors even quote precipitation ranges necessary for duricrust formation (as discussed in section 6.1.5*c* above). However, unless unequivocal evidence for contemporary development can be established, and therefore duricrust formation can be related to present-day environmental parameters, the palaeoenvironmental significance of any occurrence is questionable.

Table 6.4: Diagnostic petrographic features in southern African weathering and non-weathering profile silcretes (Summerfield, 1983*a*).

Petrographic feature	Weathering profile silcrete	Non-weathering profile silcrete
Optically continuous overgrowths	Absent	Uncommon
Chalcedonic overgrowths	Absent	Uncommon
Length-slow chalcedony vugh-fills	Absent	Common
Glaebules (authigenic)	Abundant	Absent
Colloform features	Abundant	Absent

Whilst a semi-arid origin is reasonable for most calcretes and many instances of silcrete, studies in southern Africa (e.g. Summerfield, 1978, 1982, 1983*c*) suggest that certain silcretes have developed under more humid conditions. As noted above, Summerfield (1978) distinguishes two groups of silcretes; those associated with a weathering profile and those where no such profile is present. These two groups also have different spatial distributions, with weathering profile silcretes found mainly in the Cape coastal zone and non-weathering profile silcretes generally restricted to the Kalahari. Summerfield (1983*a*) also notes major geochemical and petrographic differences between the two silcrete types, particularly in terms of the trace minerals they contain. He specifically subdivides silcretes on their TiO_2 content, with weathering profile types being titanium-enriched in comparison to titanium-poor non-weathering profile types.

The main petrographic differences between the two silcrete types are shown in table 6.4. Summerfield (1983*a*) also notes that weathering profile silcretes tend to exhibit F- or M-fabrics, a feature which he attributes to the likelihood of the silcrete having developed from a highly weathered host material. Non-weathering profile silcretes have more variable petrographic characteristics, with a wider variety of fabric types.

Interpreting the reasons for these differences in silcrete type, Summerfield (1983*a*) suggests that the general absence of deep-weathering profiles in the Kalahari (with exceptions such as the Serowe escarpment) indicates semi-arid conditions throughout much of the Cenozoic. However, for periods prior to the Quaternary there is a lack of independent palaeoclimatic evidence to confirm such a suggestion. Petrographically, non-weathering profile silcretes tend to lack silicified weathering profile clays, which provide further evidence for formation under a semi-arid or arid climate. Conversely, Summerfield (1983*a*) considers weathering profile silcretes in Cape coastal zones to be formed in association with intense weathering. Studies of absolute elemental gain-loss in these silcretes (Summerfield, 1984) indicate enrichment in TiO_2 with respect to bedrock. As titanium is only mobile under low pH conditions, an acid environment, probably with abundant vegetation, is indicated at the time of mobilisation (Summerfield, 1983*a*). A similar climatic interpretation has been given by Thiry (1989) for the TiO_2-rich silcretes mentioned in section 6.1.6*d* (ii) above.

Twidale and Hutton (1986) consider that Summerfield's (1983*a*) climatic conclusions based purely upon TiO_2 alone are insufficient. They argue that concentrations of other elements such as yttrium and zirconium should also have been presented, since without this evidence titanium may simply have relatively accumulated rather than being mobilised and emplaced at the time of silicification. Summerfield (1986) counters this argument by noting that absolute elemental gain-loss calculations give no indication of the considerable volume of bedrock reduction necessary for such high levels of relative titanium enrichment.

As previously mentioned, the environmental conditions conducive to duricrust formation in both the source area and site of precipitation need to be taken into account. Dury (1968) and Stephens (1971) both consider the source material for certain Australian silcretes to be derived from more humid areas, with precipitation occurring in areas where it is either more arid or drainage is impeded. Taylor and Ruxton (1987) use palynological evidence supported by geomorphic information to propose a wet and warm climate for silcretes in southeast Australia. There are other suggestions of silcrete formation in humid environments, such as in well-drained upland soils (Van De Graaff, 1983) and in the swampy floodplains of rivers (Wopfner, 1978). However, particularly in the latter example, local environmental factors may be more significant than general climatic controls.

In summary, previous studies suggest that, from consideration of a silcrete's fabric, geochemistry and petrography, together with careful assessment of local environmental factors, it may be possible to identify conditions at its time of formation, although problems of interpretation remain. On the basis of Summerfield (1983*a*), a silcrete developed under conditions conducive to deep-weathering should have a high level of TiO_2, will most probably have an F- or M-fabric and contain both glaebular and colloform features. On the other hand, silcretes developed under semi-arid conditions will not contain such features but will include chalcedonic overgrowths and/or vugh-fills, a wider variety of fabric types and will exhibit low levels of TiO_2. Phases of silicification in calcretes may also be interpreted in similar ways if features such as colloform structures and the particular species of quartz present are considered. However, as has been noted, if environmental site factors are most important it may be extremely difficult to interpret the palaeoclimatic significance of a particular duricrust type. Whilst calcretes are generally regarded as being indicative of semi-arid climates, studies of the association between plant roots and vadose zone calcrete development in Australia (Semeniuk and Seale, 1985) suggest differently. If roots are essential to development in certain locations, then under a semi-arid climate when less plants are present, this would hinder development unless capillary rise mechanisms increased in importance. However, it should be noted that the study considers only specific calcrete types developed in Holocene coastal sands and may not be universally applicable.

6.1.8 Dating duricrust formation

If duricrusts are used as palaeoenvironmental evidence, then it is clearly desirable to be able to date their time of formation. Relative dating of southern African calcretes has been undertaken by Netterberg (1969*a*). However, successful determination of the absolute age of a duricrust is very difficult. The three main absolute dating methods are uranium-series, radiocarbon and electron spin resonance techniques, with U-series and ^{14}C only applicable to dating calcrete formation (Radtke *et al.*, 1988).

ESR is presently the only available technique for absolute dating of silcretes (Radtke and Brückner, 1991), but is still in its preliminary stage of development. Other possible methods for dating silcrete formation including fission-track and palaeomagnetic iron dating (Summerfield, 1983*b*), and thermoluminescence techniques may be applicable in the future (Thomas and Shaw, 1991*a*). No available method can take into account the problem that duricrust development is almost certainly not a discrete event. Most techniques rely on the use of crushed samples, and can therefore only represent an average date, particularly where multiple stages of development have taken place. Wright (1978) suggests at least three phases of Kalahari silcrete development since the Cretaceous, which would render techniques such as ESR redundant. In the case of calcrete, contamination of a sample by percolating waters may occur, in addition to the multiple generations of carbonate which may be incorporated within a duricrust (Goudie, 1983).

Radiocarbon dating is the most commonly applied dating technique, particularly where archaeological remains or shells suitable for dating are present. However, as Rust *et al.* (1984) warn, the use of calcrete as a palaeoenvironmental indicator is dangerous, not least due to problems of contamination and hence spurious radiocarbon dating. Also, as the upper age limit for the radiocarbon technique is around 40,000 years BP (or 100,000 years BP if accelerated), it only has limited applicability. Relative dating is however possible, where a duricrust is overlain by a datable horizon such as a basalt flow or contains shell material, although in the latter case there are still potential problems with contamination.

6.2 Methodology and results of duricrust analysis

Having considered the possible modes of origin, mineralogy and significance of duricrusts in the preceding sections, the remainder of this chapter addresses the relationship between Kalahari duricrusts and *mekgacha*, concentrating particularly upon water movements and duricrust formation.

Under the major hypotheses of Kalahari valley development discussed in chapter 4 it is possible that the presence of a valley has affected local hydrological conditions and allowed for absolute accumulation of $CaCO_3$ and SiO_2 due to water table fluctuations and by movement of lateral throughflow. However, it is also possible for the duricrusts to have been formed prior to valley incision. Therefore to evaluate the hypotheses it is necessary to establish whether the duricrusts within a valley existed prior to the formation of that valley.

From the formational models presented in section 6.1.6 above, there are four ways in which a duricrust exposed within a *mokgacha* at the present day may have developed;

(i) Duricrust formation occurred due to pedogenic processes operating prior to valley incision, with duricrusts subsequently exposed by valley downcutting. An identifiable duricrust stratigraphy should be present.

(ii) The duricrust developed as a direct result of the presence of a valley by accumulation of minerals in the watertable zone beneath the valley flanks, in the way proposed by Thiry *et al.* (1988), with subsequent incision separating the duricrust from the valley floor.

(iii) The duricrust developed by groundwater-related processes ("phreatic" types of Mann and Horwitz, 1979) with precipitation at or beneath a shallow valley watertable and subsequent downcutting leading to perched remnants. Calcareous tufas may be considered under this possible mode of formation.

(iv) Duricrusts formed under pan conditions (the pans possibly located in a proto-valley), with precipitation caused by pH changes due to lateral movements of water into alkaline pan conditions. Whilst not directly relevant to valley development, the presence of standing water in pans located within valleys may have generated localised environmental conditions suitable for mineral precipitation.

If a pedogenic origin can be shown, then there is evidence to support a "fluvial" origin for *mekgacha* (i.e. the valley has incised through pre-existing duricrusts). If not, then it can be inferred that the crusts have probably developed as a direct result of processes related to the existence of the valley and the effects this would have on local hydraulic gradients. Clearly this simplifies the argument, as it is possible that a valley incised its channel by fluvial erosion, thus influencing local hydrological conditions and causing diagenetic alteration to pre-existing duricrusts (and also leading to the generation of further duricrusts, possibly by pedogenic processes operating within valley slope materials). The duricrusts may also have been modified by, or developed in association with, another conduit for water movement such as a geological lineament or fracture. It is, however, very difficult to assess the impact of a valley upon local hydrology during the early stages of its development. Clearly the development of any linear depression would have influenced flowpaths of groundwater and throughflow by altering hydraulic gradients.

However the present status of watertables at levels well below valley floors (Thomas and Shaw, 1991*a*) raises problems for the analysis of watertable-valley interactions.

The degree to which a "groundwater" origin may be inferred is highly dependent upon the nature of the particular duricrust type within a valley. Additionally, the fact that a groundwater-related duricrust may be identified within a valley does not automatically imply the role of groundwater processes in valley development. It would only be possible to state that the valley existed prior to the formation of the duricrust suite and no definite conclusions regarding the mode of valley origin can be made. Despite these limitations, duricrust studies play an important role in evaluating hypotheses for valley development when viewed in conjunction with the results of both the preceding and following chapters.

Four approaches to the study of duricrusts were used. Firstly, the field distribution of duricrusts types was mapped by considering profile variations in selected valleys with extensive duricrust exposures. From this technique, any stratigraphy within valley duricrusts at the selected locations could be noted. Clearly, studying only those duricrusts which were well-exposed may have necessarily biased sampling towards more indurated duricrust varieties or those occurring within valleys which are more deeply incised. However, it was felt that the other methods of duricrust study compensated for this problem. Under the second main approach, information on duricrust thickness, composition and type was considered, principally from lithological borehole logs drilled in the vicinity of valleys. From these studies, the variation in characteristics of duricrusts associated with valleys was assessed. The third method utilised thin-section analyses of duricrust samples to assess microscopic and mineralogical indicators of the environmental conditions during their development (as discussed in section 6.1.7). Finally, x-ray fluorescence analysis of samples was utilised to evaluate bulk chemistry, with particular attention to TiO_2 concentrations (section 6.1.7) and their potential significance in recognising the environment of duricrust formation. Discriminant analysis of bulk chemistry data was carried out to identify geographical variations in duricrust composition, and to compare the bulk chemistry of silcrete samples from this study with those of Summerfield (1978, 1982, 1983*d*). This chapter now considers the methods and results of each of these approaches in turn.

6.2.1 Studies of duricrusts in profile

(a) Methodology

The first method of duricrust analysis employed was the study of duricrust profiles in areas where extensive exposures of silcrete and calcrete occurred. Whilst outcrops of duricrust were present intermittently in all valley systems, exposures were often spatially and vertically limited. As a result, the main areas for this study were the valleys to the south of Letlhakeng, in the Kweneng District of Botswana, with a reconnaissance level study in the Auob Valley in eastern Namibia. The rationale for duricrust profile study was to assess evidence for correlation between different duricrusts in various profiles. If duricrust sequences correlate over large distances and a stratigraphy can be identified, notably perpendicular to the valley axis, then it is most likely that the duricrust suite existed prior to the presence of the valley. If, however, diagenetic alteration beneath the water table or the lateral movement of groundwater through valley flanks has occurred during valley development, it is possible that no such correlation will be identifiable. A lack of stratigraphy may be interpreted in either of two ways; it may be that duricrusts have formed as a consequence of the presence of a valley, or that movements of water towards the valley over long time periods have altered pre-existing duricrust suites to the extent that any previous stratigraphy is rendered unidentifiable. Clearly, a lack of evidence for stratigraphy does not necessarily preclude the existence of duricrusts prior to valley incision, but means that careful identification of the mode of origin of duricrusts is required.

In both the Letlhakeng valleys and the Auob, study took the form of basic geological description of duricrust sequences. In Letlhakeng, duricrusts were studied in a number of transects across valleys with all major variations in lithology, morphology and appearance noted, together with estimated thicknesses for individual "beds". In the Auob, study was at a reconnaissance-level, with exposures studied mainly on the west side of the valley due to difficulties of access.

(b) Results

(i) Letlhakeng valleys

Duricrusts are exposed almost continuously in Valleys 1, 2 and 3 to the south of Letlhakeng village (24°06' South 25°02' East). As noted in the preceding chapter, the best exposures occur in the form of low cliffs, whilst the majority of outcrops in the Letlhakeng areas are poorly exposed in valley flanks. Taking outcrops on both flanks of the three valleys into consideration, the total length of exposure in the area is in excess of 80 km. In order to assess variations in the type of duricrust exposed, a series of valley flank transects were studied. The transects were spaced at either half or one kilometre intervals along each valley dependent upon exposure quality, starting at the southernmost duricrust exposure within each valley. This resulted in a total of sixty transects being surveyed, as shown on the insets of figures 6.4 and 6.5, with the precise locations of transects given in table 6.5. The results for Valley 1 are shown in figure 6.4 whilst those for the main valley of Valley 2 and the main and tributary valleys of Valley 3 are given in figure 6.5. It should be noted that no attempt has been made to correlate between profiles on the diagrams due to the highly variable nature of the duricrusts under consideration; such correlation would imply possible lithostratigraphic relationships which do not appear to be present. Kalahari Sand was found overlying duricrusts in most profiles, and is not indicated on the diagrams.

Letlhakeng Valley 1

Figure 6.4 demonstrates variations in duricrust type within Valley 1 in long-profile, with an overall drop in the valley bed of some 51 m over the 8 km stretch of valley studied (an average gradient of 0.36°). Transects were surveyed from the amphitheatre valley head area at grid reference 24°09'30" South 25°12'00" East to 24°05'00" South 25°03'30" East (table 6.5). It should be noted that the base of the schematic profiles shown in the figure do not necessarily correspond with the valley base, so that the long-profile is not markedly stepped as the figure suggests. Typically, duricrusts are exposed in low cliffs at the top of debris slopes, although only debris slopes of > 5 metres vertically are included on the diagrams.

Table 6.5: Grid references of transects surveyed in the Letlhakeng area. All transects were surveyed at 90° to the valley axis, with grid references referring to a position on the valley axis. Grid references for transects marked * are for positions on the valley flank not the valley axis.

No.	Grid reference		No.	Grid reference	
	South	East		South	East
VALLEY 1			VALLEY 2		
H	24°09'35"	25°12'00"	2	24°09'38"	25°05'00"
1	24°09'33"	25°11'44"	4 *	24°10'01"	25°04'51"
2	24°09'27"	25°11'28"	5 *	24°09'27"	25°04'21"
3	24°09'16"	25°11'14"	6 *	24°09'01"	25°04'54"
4	24°09'15"	25°10'54"	7	24°08'01"	25°04'08"
5	24°09'09"	25°10'33"	8	24°08'30"	25°03'58"
6	24°09'00"	25°10'22"	9	24°08'11"	25°04'03"
7	24°08'53"	25°10'07"	10	24°07'34"	25°03'52"
8	24°08'50"	25°09'45"	11	24°07'09"	25°03'30"
9	24°08'38"	25°09'31"			
10	24°08'27"	25°09'21"	VALLEY 3		
11	24°08'14"	25°09'19"			
12	24°08'01"	25°09'10"	1	24°08'42"	25°00'31"
13	24°07'58"	25°08'54"	2 *	24°08'38"	25°00'51"
14	24°07'45"	25°08'30"	3 *	24°08'34"	25°01'21"
15	24°07'29"	25°08'00"	4 *	24°08'42"	25°01'44"
16	24°07'08"	25°07'33"	5 *	24°08'58"	25°01'49"
17	24°06'40"	25°06'55"	6 *	24°08'50"	25°01'30"
18	24°06'27"	25°06'30"	7 *	24°08'47"	25°01'10"
19	24°06'17"	25°05'50"	8 *	24°08'53"	25°01'05"
20 *	24°05'24"	25°05'10"	9 *	24°09'03"	25°00'52"
22 *	24°06'09"	25°06'15"	10	24°09'09"	25°00'22"
23 *	24°06'08"	25°05'31"	11	24°09'25"	25°00'17"
24 *	24°06'11"	25°05'22"	12	24°09'16"	25°00'12"
25 *	24°06'30"	25°04'43"	15	24°09'09"	24°59'00"
26 *	24°06'30"	25°05'00"	16	24°09'04"	24°59'35"
27	24°05'35"	25°04'31"	17	24°08'51"	25°00'12"
28	24°05'00"	25°03'37"	18	24°08'25"	25°00'19"
			19	24°08'04"	25°00'20"
			20	24°07'32"	25°00'28"
			21	24°09'37"	25°00'28"
			22	24°09'19"	25°00'24"
			23	24°07'02"	25°00'37"
			24	24°06'39"	25°00'51"

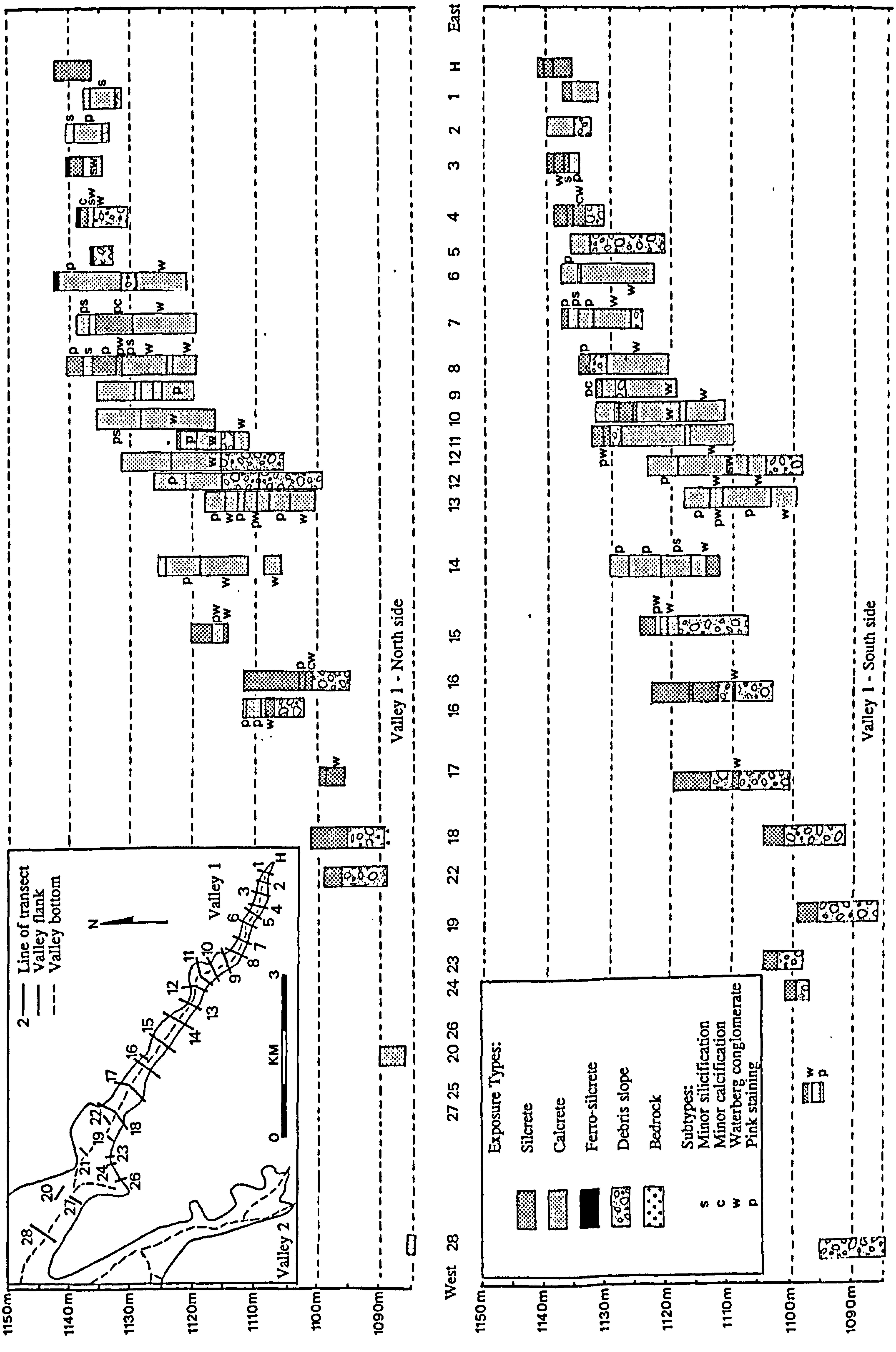


Figure 6.4: Variations in duricrust type and thickness; Letlhakeng Valley 1.

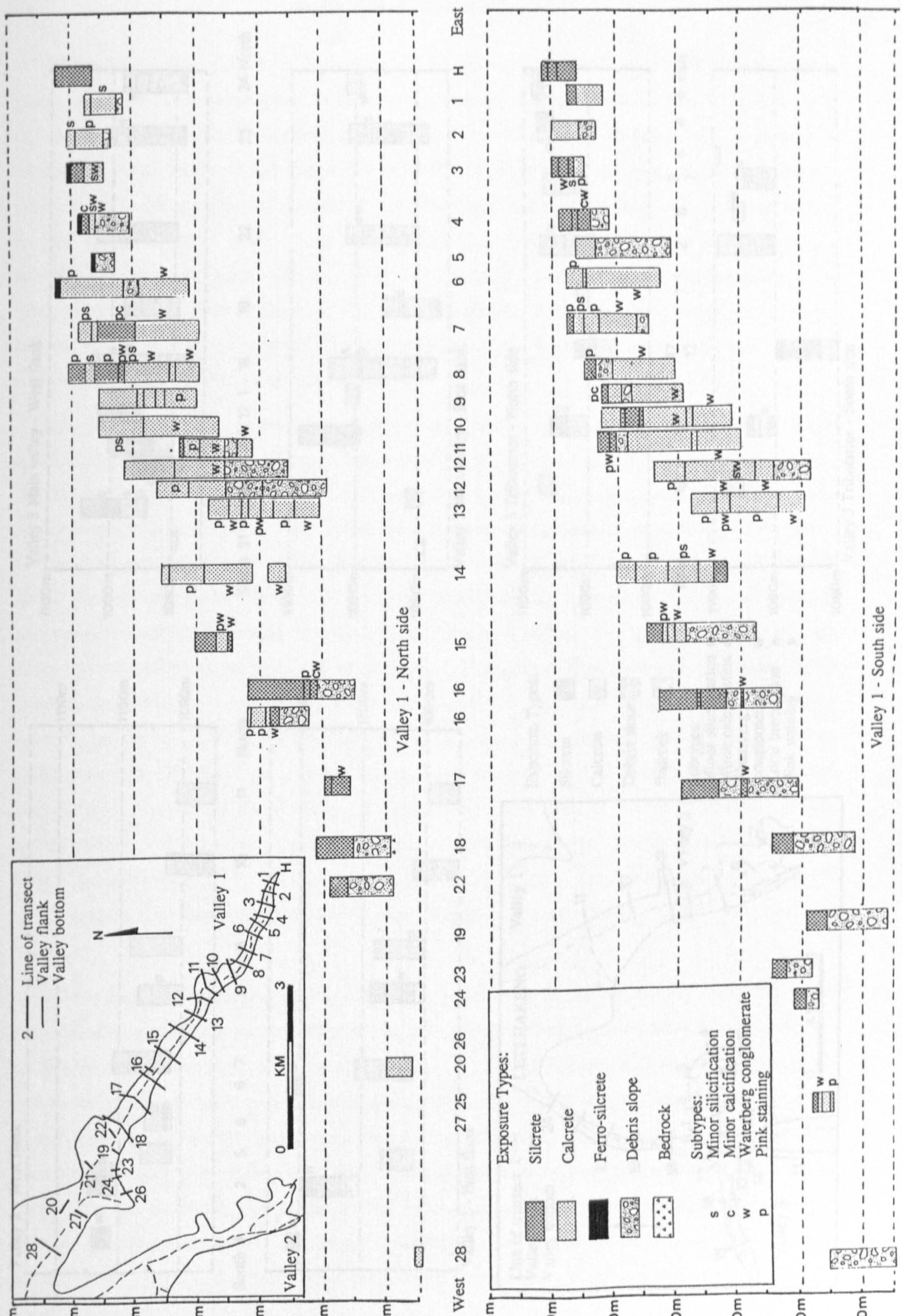


Figure 6.4: Variations in duricrust type and thickness; Letlhakeng Valley 1.

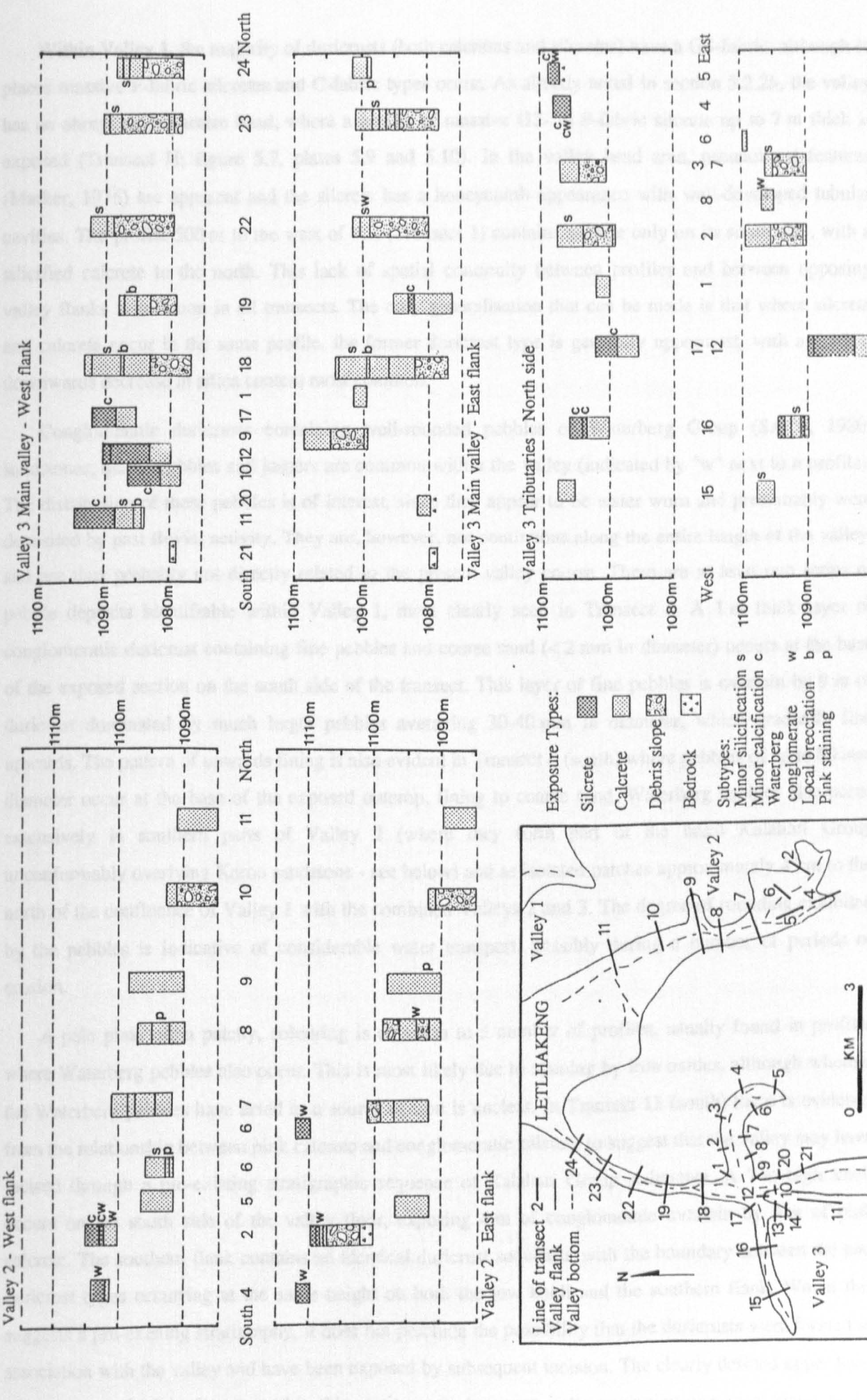


Figure 6.5: Variations in duricrust type and thickness; Letlhakeng Valleys 2 and 3.

Within Valley 1, the majority of duricrusts (both calcretes and silcretes) have a GS-fabric, although in places massive F-fabric silcretes and C-fabric types occur. As already noted in section 5.2.2*b*, the valley has an abrupt amphitheatre head, where a profile of massive GS- to F-fabric silcrete up to 7 m thick is exposed (Transect H; figure 5.7, plates 5.9 and 5.10). In the valley head area, pseudokarst features (Marker, 1976) are apparent and the silcrete has a honeycomb appearance with well-developed tubular cavities. The profile 500 m to the west of this (Transect 1) contains silcrete only on its south side, with a silicified calcrete to the north. This lack of spatial continuity between profiles and between opposing valley flanks is common in all transects. The only generalisation that can be made is that where silcrete and calcrete occur in the same profile, the former duricrust type is generally uppermost, with a gradual downwards decrease in silica content most common.

Conglomeratic duricrusts containing well-rounded pebbles of Waterberg Group (SACS, 1980) sandstones, quartz pebbles and jaspers are common within the valley (indicated by "w" next to a profile). The distribution of these pebbles is of interest, since they appear to be water worn and presumably were deposited by past fluvial activity. They are, however, not continuous along the entire length of the valley, and are thus probably not directly related to the present valley course. There are at least two series of pebble deposits identifiable within Valley 1, most clearly seen in Transect 8. A 1 m thick layer of conglomeratic duricrust containing fine pebbles and coarse sand (< 2 mm in diameter) occurs at the base of the exposed section on the south side of the transect. This layer of fine pebbles is overlain by 9 m of duricrust dominated by much larger pebbles averaging 30-40 mm in diameter, which gradually fine upwards. The pattern of upwards fining is also evident in Transect 6 (south) where pebbles of up to 80 mm diameter occur at the base of the exposed outcrop, fining to coarse sand. Waterberg pebbles also occur extensively in southern parts of Valley 2 (where they form part of the basal Kalahari Group unconformably overlying Karoo sandstone - see below) and as isolated patches approximately 4 km to the north of the confluence of Valley 1 with the combined Valleys 2 and 3. The degree of rounding exhibited by the pebbles is indicative of considerable water transport, possibly during a number of periods of erosion.

A pale pink, often patchy, colouring is common to a number of profiles, usually found in profiles where Waterberg pebbles also occur. This is most likely due to staining by iron oxides, although whether the Waterberg pebbles have acted as a source of iron is unclear. In Transect 13 (south) there is evidence from the relationship between pink calcrete and conglomeratic calcrete to suggest that the valley may have incised through a pre-existing stratigraphic sequence of Kalahari Group sediments. A 7 m high knoll occurs on the south side of the valley floor, exposing 4 m of conglomerate overlain by 3 m of pink calcrete. The southern flank contains an identical duricrust sequence, with the boundary between the two duricrust types occurring at the same height on both the low knoll and the southern flank. Whilst this suggests a pre-existing stratigraphy, it does not preclude the possibility that the duricrusts were formed in association with the valley and have been exposed by subsequent incision. The clearly defined upper level of conglomeratic deposits exposed in this section may, however, indicate erosion or a local hiatus in the deposition of the Kalahari Group. Finally, the occurrence of an isolated small area of ferro-silcretes beneath Kalahari Sand on the uppermost parts of the north sides of Transects 3, 4, 5 and 6 should be noted.

Letlhakeng Valley 2

Valley 2 (profiles surveyed from grid reference 24°10'30" South 25°04'30" East to 24°07'15" South 25°03'30" East; table 6.5) shows a similar lack of correlation between duricrusts (figure 6.5), although GS- to F-fabric silcretes and cal-silcretes mainly occur near the head of the valley with calcretes dominating to the north. In contrast to Valley 1, exposures of duricrusts tend to occur as sloping valley flanks rather than low cliffs. As a result, exposures are often covered by slope deposits and shallow soil, obscuring any sedimentary variations which may be present. Waterberg pebbles occur extensively both within conglomeratic duricrusts and as lag deposits covering the flanks of southern parts of the valley. These conglomerates are part of the basal Kalahari Group sediments, and can be seen immediately overlying Karoo sandstone in the east side of Transect 2 (plate 5.14). From studies of borehole logs drilled within the Molopo Valley, Smit (1977) relates similar basal Kalahari conglomerates to former northerly-flowing rivers, although Gould and Rathbone (1985) and Levin *et al.* (1985) indicate additional southerly directed buried channels beneath the Molopo Valley. As deposition of the Kalahari Group commenced in the Jurassic/Cretaceous, the conglomerates are clearly of some antiquity (Thomas and Shaw, 1991*a*).

Letlhakeng Valley 3

Valley 3 (profiles surveyed from grid reference 24°09'30" South 25°00'30" East to 24°06'40" South 25°00'12" East; table 6.5) is of greater interest since it contains extensive outcrops of duricrust not only within its main trunk valley but also in two tributaries. Duricrusts are well-exposed, both as cliffs up to 6 m in height at the top of debris slopes and on valley sides. In the trunk valley, the main outcrops of silcrete occur only on the west flank from Transect 17 southwards. Other silcrete outcrops occur in the tributary valleys, with localised areas of Waterberg pebbles present in the eastern tributary. As in Valley 2, these are part of the basal Kalahari Group, with Karoo Kweneng sandstone bedrock outcropping close to Transect 5, south of Transects 11 and 21 and in the centre of the main valley where the two tributaries converge (plate 5.16). Of these bedrock outcrops, only the sandstone appearing in the main valley centre and that exposed to the south of Transect 11 show evidence for calcification. Sandstone exposed in the centre of the main valley gives a weak reaction to dilute hydrochloric acid, indicating the presence of some calcium carbonate. The exposure of Kweneng sandstone to the south of Transect 11 is immediately overlain by sand and calcrete rubble, the lowest outcrop of consolidated calcrete occurring 40 cm above the sandstone. This calcrete contains very fine bedding structures and would appear to be altered sandstone, suggesting weathering associated with calcification.

The only semi-continuous duricrust horizon seen in the three Letlhakeng valleys occurs within the main trunk of Valley 3. On the east side of Transect 19 an outcrop of cal-silcrete occurs, traceable for over 50 m north and south of the profile transect line. The transect was taken across a section of valley containing a large pan, which is now dammed and used for livestock watering during the summer. The extent of the cal-silcrete outcrop broadly coincides with the maximum limits of the pan. It is possible that the siliceous horizon formed as a result of lateral movement of groundwaters associated with an earlier pan at a time when the valley floor was at a higher level than present. A lowering of the valley bed may have occurred, leaving the cal-silcrete some 8 m above the present pan floor. A problem with this suggestion is that no comparative cal-silcrete horizon occurs on the western valley flank, which is perhaps due to a slight lateral migration of the valley during its incision.

General implications of duricrusts from Letlhakeng

The profile studies outlined above indicate that there is considerable variation in duricrust lithology both within and between valleys. Even taking the altitudinal differences between the three valleys into account (see figures 6.4 and 6.5) there does not appear to be a common stratigraphic sequence in the Letlhakeng area. One common duricrust type found within all valleys is a compact hardpan calcrete, usually containing no signs of silicification apart from a silica-rich laminar surface rind. This hardpan is commonly exposed in valley floor areas, and is particularly prevalent close to Letlhakeng village. The overall form of exposures, particularly their domed morphology, suggests that these duricrusts formed as valley calcretes in the manner suggested by Mann and Horwitz (1979).

The fact that there is no apparent correlation between the duricrust profiles studied in any of the three Letlhakeng valleys suggests that they were either formed after the incision of the valleys or have been considerably altered during valley development. The duricrusts studied were highly variable in terms of fabric, appearance and chemistry, even within individual profiles, making generalisations very difficult. This high degree of variability was also noted by Shaw and De Vries (1988). There is little evidence for small-scale tectonic activity having displaced duricrust horizons on either side of valleys, which could help explain such a lack of correlation. Indeed, the only place where a local faultline occurs at the surface (from Landsat imagery interpreted for the report by Mallick *et al.*, 1981) is in the western tributary valley of Valley 3, and no post-duricrust displacement is evident.

Another possibility is that duricrusts did exist prior to valley development, but during (or even after) incision the duricrusts on opposing valley flanks developed independently as a result of lateral throughputs of groundwater. It is likely that opposing valley sides would have individual local groundwater environmental conditions, with possible differences in pH and in levels of dissolved silica and calcium carbonate depending upon the provenance of such solutes. This could explain the lack of anything but a general similarity between most opposing valley flanks, but would account for the occurrence of conglomeratic duricrusts on either side of valleys. It could also explain the presence of anomalies such as the area of ferro-silcrete on the north side of Valley 1 (Transects 4, 5 and 6). The relationship between pink and conglomeratic calcrete types in Transect 13 of Valley 1 outlined above indicates that there was some pre-existing stratigraphy within duricrust host materials but does not suggest that the duricrusts themselves existed prior to valley development.

Finally, the relationship between the duricrusts exposed in the Letlhakeng valleys and the overlying Kalahari Sand should be considered. Thomas and Shaw (1991*a* p.62) note that a possible approach to correlating the Kalahari Group over large areas is by the study of unconformities in the sedimentary sequence. The conglomeratic duricrusts at the base of the Kalahari Group rest uncomformably upon the underlying Karoo bedrock, this break in deposition being best exposed in Valley 2. The break between conglomerates and overlying calcrete in Valley 1 Transect 13 may indicate another hiatus or erosion surface. However, only one consistent zone of lithological unconformity was evident within the Kalahari Group in the Letlhakeng area, this being between the upper Kalahari Sand and underlying duricrusts. As will be discussed below, Kalahari duricrusts usually contain a high proportion of clastic material, commonly quartz grains. In the vicinity of the Letlhakeng valleys, the Kalahari Sand varies in colour between red and ochre due to iron oxide coatings, whilst grains within duricrusts viewed in thin-section (see section 6.2.3) rarely show such coatings. The fact that the coated sand does not appear to be incorporated within duricrusts and that windblown sands often cover duricrust exposures implies a definite break between the formation of the duricrust suite and the deposition of the Kalahari Sand.

(ii) Auob valley

The form of the Auob valley has already been discussed in section 5.3.1, with particular attention paid to its depth of incision into the surrounding Kalahari "Limestone Plateau" (Mabbutt, 1957). Exposures of duricrusts of varying quality occur along most of the valley length, with major exposures occurring to the south of Stampriet in Namibia and continuing until the confluence with the Nossop valley in the Kalahari Gemsbok National Park (section 5.3.1*b*). Whilst exposures are more extensive than those in the vicinity of Letlhakeng village, they are invariably much higher and of less easy access. The problems of adequately sampling duricrusts from 9 m vertical cliffs atop 15 m scree slopes, difficulties of access to farmland and being unable to collect samples within the Gemsbok Park made an extensive survey similar to that undertaken at Letlhakeng impracticable.

Major duricrust outcrops first occur along the Auob Valley to the south of Stampriet and are associated with the appearance of a "badlands" topography (plate 5.31). Unlike the Letlhakeng valleys, there is evidence for a consistent duricrust stratigraphy along the valley which can be identified between exposures at Witkrans Farm (24°26'S 18°30'E) and the boundary of the Gemsbok Park. The exposure at Witkrans indicates the typical duricrust sequence encountered along the Auob. A 7 m high outcrop occurs at the top of an 8 m high debris slope on the eastern side of the valley immediately east of the farm. Uppermost parts of the exposure could not be reached, but the top 5.0 m of the duricrust cliff appeared to consist of a highly indurated iron-stained calcrete with a crumb-like texture. Beneath this indurated material, 2.0 m of a considerably softer white calcrete horizon occurs, with differential weathering producing an overhang at the boundary of the harder and softer lithologies. Both upper and lower sections of the cliff exhibit a laminated appearance due to traces of iron-staining within individual horizons. Despite this horizontal to sub-horizontal staining, no bedding was present within the exposure.

Exposures seen further south allow more detailed examination of the morphology of the duricrust sequence, particularly at Kalkheuval Farm (24°45'45"S 18°44'25"E) where samples were collected for thin-sectioning (section 6.2.3) and x-ray fluorescence analysis to determine bulk chemistry (section 6.2.4). An overall view of the valley from the sampling location is indicated on plate 5.32. The sample profile contains an identical sequence to that seen at Witkrans, with 5.5 m of indurated iron-stained calcareous duricrust overlying approximately 3.2 m of a much softer white calcrete. This cliff of duricrust occurs at the top of an 18-20 m high debris slope at an angle of between 24 and 27°. Unlike the exposure at Witkrans, an overhang does not occur at Kalkheuval to indicate the differential resistances of the two duricrust types. Instead, there is a break of slope at the contact of the two lithologies with the more indurated material forming a vertical cliff above a 45° slope of softer calcrete partially covered by the debris slope.

Within this simplified sequence of harder and softer duricrust there are, however, other lithological variations which are further discussed in section 6.2.3. The uppermost 2 m of the exposure are partly silicified, this induration contributing to the resistance of the duricrust. The top of the cliff has an almost horizontal surface which, unlike exposures seen at Letlhakeng, does not exhibit pseudokarst features. The upper plateau surface does, however, contain occasional hollows visible on aerial photography which may indicate removal of material in solution. Pebbles of jasper and quartzite (up to 5 mm diameter) are incorporated into the upper sections, with a 30 cm thick conglomerate band occurring at 2.1 m below the top of the profile. Variations in the iron content are indicated by differences in colour between upper and lower sections. No obvious lineations were evident within the profile, neither were any bedding structures.

The road cuttings at Grensplaas (25°29'30"S 19°29'00"E) and Okampuma (25°02'30"S 18°54'00"E) farms and exposures at Grauwater (25°09'30"S 18°58'00"E) allow further analysis of the upper sections of the duricrust profile. The road cutting at Okampuma reveals over 5 m of exposure with three apparent lithologies present. These consist of a lower 1.5 m of nodular calcareous duricrust with a crumb-like appearance, overlain by 2.5 m of a similar material with a significant coarse gravel content and an uppermost 1.0 m of brittle silicified calcrete. The upper exposures are of most interest as the silicification has occurred by a replacement of calcium carbonate within the duricrust to produce patches of darker siliceous cement within the iron-stained calcrete matrix. There is additional evidence of horizons within the duricrusts, possibly indicative of a pedogenic origin. At Grensplaas, 1.5 m of nodular duricrust are exposed, equivalent to that seen in the lower sections of the exposure at Okampuma, with 2.0 m of the same material exposed at the Auob-Elephants confluence. The exposure at Grauwater allows inspection of the junction between the lower and upper duricrust types described above from further north, with extensive alcove development present at the interface of the two duricrust types. The appearance of a localised indurated layer forming a terrace level at 7 m above the valley floor in the eastern valley flank at Simon Koper Farm (24°50'S 18°47'E) has been commented upon in section 5.3.1*b*. This layer consists of approximately 2 m of silicified calcrete of a similar form to that seen at the top of cliff exposures, but only occurs over 4 km of the valley flank and does not appear to outcrop in the western flank.

The extent of the Kalahari plateau in the vicinity of the Auob and Nossop valleys has also been commented upon in section 5.3.1*b* (above). The occurrence of two distinct duricrust types is not restricted only to exposures occurring in the valley flanks. Numerous gullies and dendritic tributaries cut through the plateau in the area of "badlands" topography and reveal an identical sequence of duricrust exposures up to 500 m away from the valley axis (e.g. to the north of Twee Rivier Farm; 25°27'30"S 19°26'30"E). The sequence is also traceable northwards along the Elephants River from its confluence with the Auob. This consistent stratigraphy would explain the flatness and extent of the Kalahari plateau, with the upper silicified duricrusts providing a resistant cap-rock. As noted in the preceding chapter, the occurrence of this cap-rock leads to differential weathering which has contributed to the development of the "badlands" topography.

Finally, the form of the duricrust exposures within the Kalahari Gemsbok National Park needs to be considered, although descriptions are necessarily brief since detailed analysis and sample collection was not permitted. Exposures are not as thick as those seen up-valley and show little variation in lithology, the most common duricrust type being a nodular silicified calcrete as seen in the upper sections of cliffs described above. The major difference between exposures outside and within the Gemsbok Park is in profile form, the morphology of exposures in the park being more complex. Profiles are generally horizontally bedded, but in places exhibit undulating horizonation, with the horizons appearing to reflect the topography of the valley side. As detailed inspection was not possible, it is difficult to assess whether or not the duricrust formed pedogenically, with the undulations representing the surface topography at the time of formation. Major undulations occur where minor tributaries appear to have once entered the valley, the duricrusts apparently mimicking the cross-sectional form of these tributaries.

From the above discussion it would appear that there is a different relationship between valleys and duricrusts in the case of the Auob and Letlhakeng valleys. No obvious stratigraphy occurs in the duricrusts exposed at Letlhakeng whereas the Auob to the north of the Gemsbok Park appears to be incised through a definite sequence of crusts. The duricrust/valley relationship within the Gemsbok Park, however, suggests that duricrusts formed after the development of minor tributary valleys.

6.2.2 Duricrusts in borehole records

As previously mentioned, it is possible to ascertain compositional and thickness variations in duricrusts from lithological borehole records where sufficiently detailed information is recorded. In the case of boreholes in the vicinity of Kalahari *mekgacha* this degree of logging accuracy is comparatively limited; few boreholes drilled in conjunction with mineral or groundwater exploration or more recent boreholes can be considered sufficiently reliable. The spatial distribution of boreholes is also highly variable, with only four areas containing a sufficiently large number of borehole logs available for detailed analyses to be undertaken. These are in the Rooibrak and Xaudum valleys from drilling by Union Carbide (1979*d* and 1980*d* respectively), in Letlhakeng Valley 2 due to calcrete prospecting by Gwosdz (1981, 1982) and Gwosdz and Modisi (1983), and in the vicinity of the Gaotlhobogwe valley (Letlhakeng Valley 1) where a large number of boreholes exist as a result of various projects. Borehole logs for Letlhakeng Valley 1 were studied at the Botswana Department of Geological Survey in Lobatse and at Wellfield Consulting Services in Gaborone.

(a) Duricrusts beneath the Rooibrak valley

Variations in the thickness of calcrete beneath the Rooibrak valley (from Union Carbide, 1979*d*) are shown in figure 6.6. From two transects drilled perpendicular to the main valley, it can clearly be seen that the calcrete is in a lenticular form, thinning rapidly away from beneath the centre of the valley where it has a maximum thickness of approximately 8m. Also of interest is that the valley appears to be situated directly above a much deeper trough in the underlying Ghanzi Group quartzites and shales.

The calcrete body has a similar form to the groundwater calcretes described by Mann and Horwitz (1979) and shows "pinching out" similar to the duricrust lenses studied by Thiry *et al.* (1988), both discussed in section 6.1.6 above. The calcrete is also clearly genetically related to the valley. From its cross-sectional form it appears to have developed after the deposition of the Kalahari Sand, incorporating sand within a carbonate cement. However, given the negligible relief of both the valley and the underlying bedrock trough, it is difficult to envisage development of the calcrete by groundwater sapping processes. Indeed, the form of the calcrete lens appears to suggest formation by water moving laterally away from or into the valley, although whether this was as a result of permanently flowing or periodically standing water is difficult to assess. Drilling along the valley (not shown on the figure) does not appear to indicate significant calcrete thickness variations. However, the bedrock trough could be indicative of deep groundwater flow along preferential flow paths beneath the valley. Such flow could be responsible for weathering of the quartzite bedrock immediately below the valley course, although the borehole records in Union Carbide (1979*d*) provide no information on the weathered status of the bedrock.

(b) Duricrusts beneath the Xaudum valley

Figure 6.7 shows the lithological variations beneath the Xaudum valley in northwestern Botswana, with boreholes drilled as part of mineralogical prospecting by Union Carbide (1980*d*). The figure shows a progressive thickening of the duricrust and sand content beneath the valley from west to east, but with disruption of the sedimentary sequence resulting from movement along a graben-like structure between boreholes C1 and C16 (numbering by Union Carbide, 1980*d*). The orientation of this graben is approximately N330°, similar in direction and possibly age to the faults controlling the Okavango "Panhandle" (Mallick *et al.*, 1981).

Whilst the thickening of the Kalahari Group towards the east (away from the Kalahari rim) is not surprising, there are certain unusual features within the sedimentary sequence. Of particular interest are the shells (borehole C3) and lignite deposits (boreholes C1 and C4), situated within a series of downfaulted "valley calcretes". Unfortunately, no information is included in Union Carbide (1980*d*) regarding the cross-sectional form of calcrete bodies within the valley, nor the species of shell present.

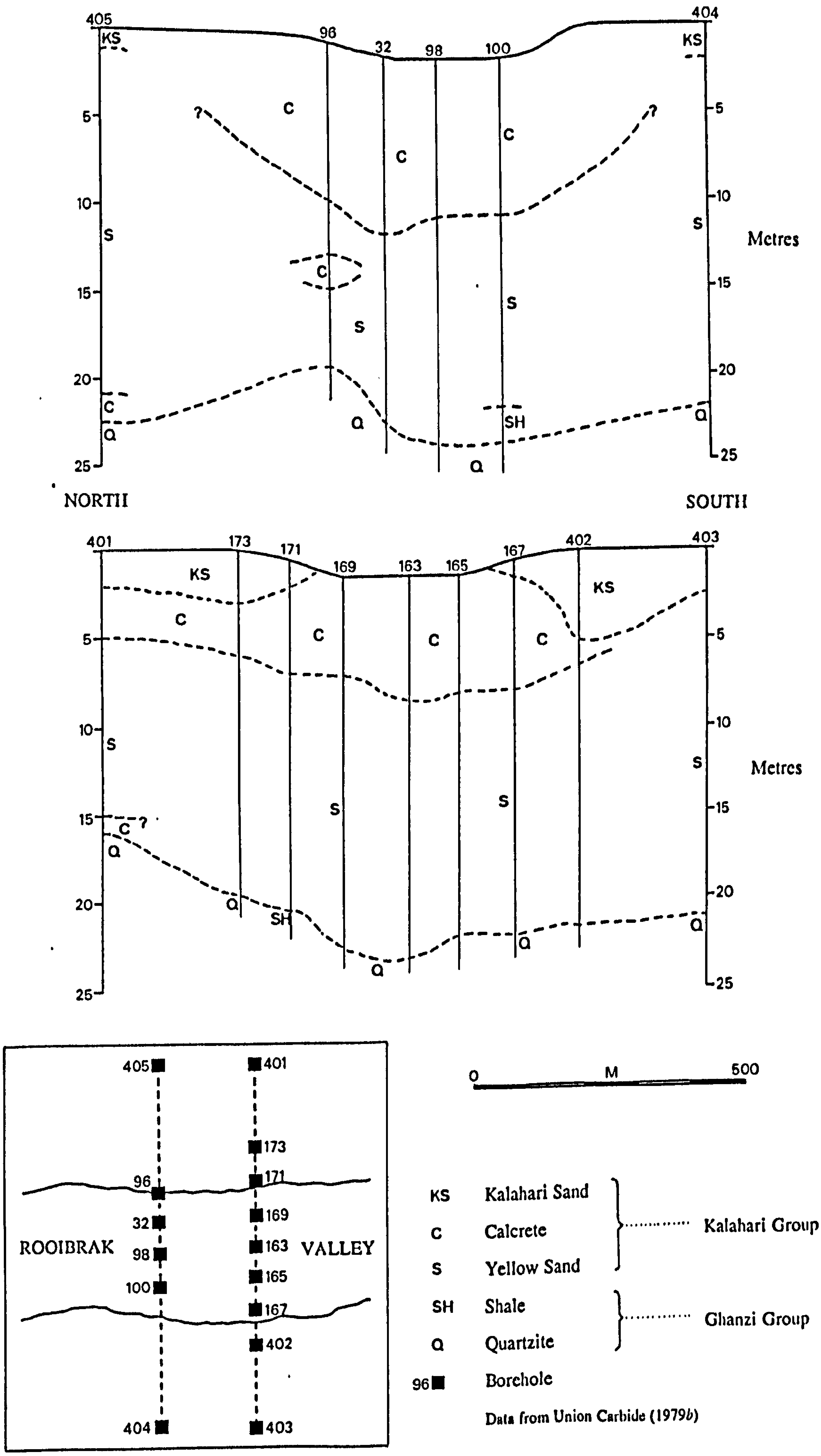


Figure 6.6: Thicknesses of calcrete beneath the Rooibrak valley, 63 km due east of Ghanzi, Botswana (after Union Carbide, 1979*d*).

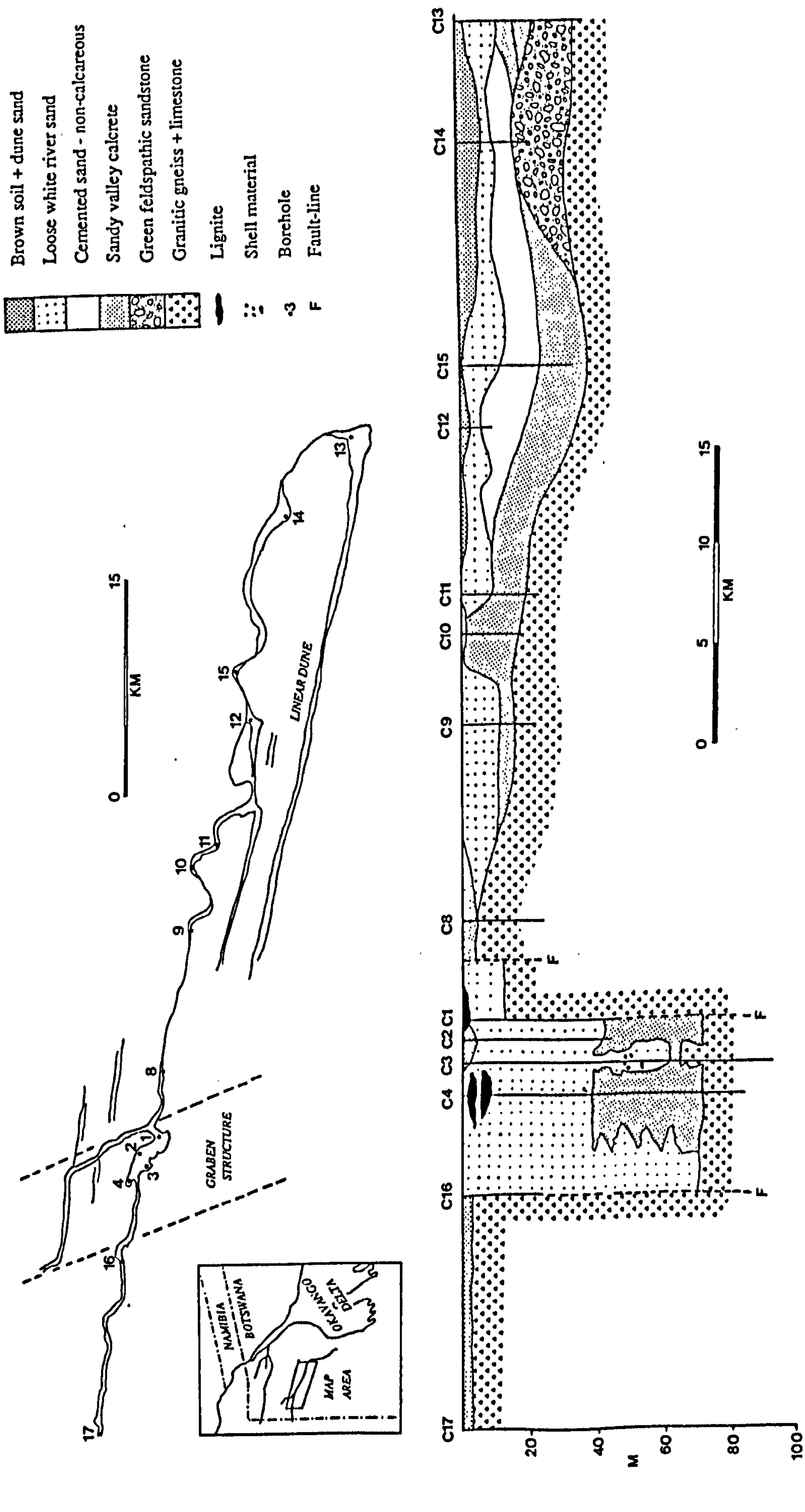


Figure 6.7: Lithological variations beneath the western Xaudum valley, northwest Botswana (after Union Carbide, 1980*d*).

However, the presence of shell and lignite deposits, assuming they are directly related to the present valley, could indicate former periods of standing water within the valley. By nature of their location 50m below the present ground surface, the shell deposits are clearly of some antiquity. More importantly they indicate that valley development was initiated prior to neotectonic movements along the faultlines (Wright, 1978; Thomas and Shaw, 1991*a*). The timing and frequency of movement along the parallel faultlines is unclear, but is likely to have occurred continuously over a long period at a sufficiently slow rate to allow gradual sedimentary infill. The comparatively shallow lignite lenses are more recent deposits, possibly related to formation in a pan environment in the graben, as also noted from studies in the Dobe valley (Helgren and Brooks, 1983).

(c) Borehole studies around Letlhakeng Valley 1

The variations in thickness of the Kalahari Group sediments beneath Letlhakeng Valley 1 are shown in figure 6.8, compiled from lithological borehole logs held at the Botswana Department of Geological Survey in Lobatse and at Wellfield Consulting Services in Gaborone. The boreholes included on the figure (with the exception of boreholes 6514 and 6518) are all drilled within, or are only a few metres above the valley floor, with locations confirmed by field reconnaissance. The figure includes both along- and across-valley variations of duricrust type where boreholes contained sufficient lithological detail. There is a degree of overlap between figures 6.4 and 6.8, with Transect 5 on figure 6.4 corresponding to Borehole 6500 on figure 6.8, and the valley head (Transect H) lying approximately 2 km east of this borehole. Borehole numbers correspond to the national numbering scheme in operation in Botswana (Sekwale, 1984).

The main observation which can be made regarding variations in thickness of the Kalahari Group directly beneath Letlhakeng Valley 1 is that pre-Kalahari bedrock is considerably closer to the surface to the west of borehole 6609. This marked change in Karoo surface topography does not coincide with the location where the valley abruptly deepens ("Head" on the figure), and as such cannot simply be attributed to removal of sediment due to incision of the valley.

Borehole records also show evidence for the action of deep-weathering processes in the vicinity of Valley 1. One example can be seen from borehole 6514, where an 8 m high cave (J. Farr, pers comm) was encountered within Karoo shales at a depth of 90 m. Further examples of the effects of groundwater flow along preferential flowpaths occur in borehole 4695 (described in Von Hoyer *et al.*, 1985) with solution and decomposition noted to depths of 125 m in Karoo sandstone and calcite infillings in underlying shales and dolerite.

Sedimentation within the Karoo depositional sub-basin in this area was strongly controlled by a northeast-southwest basement escarpment which is traceable from the Jwaneng wellfield to Botlhapatlou, passing beneath Letlhakeng (Buckley, 1984; Smith, 1984). As a result, the depth to the Karoo basement (away from valley courses) to the immediate north of Letlhakeng is over 300 m whilst only approximately 100 m to the south of the village (Buckley, 1984). The Karoo escarpment separates areas of deltaic siltstones, shales and coal measures within the depositional sub-basin from sandstones to the southeast (Smith, 1984; Shaw and De Vries, 1988). Lithological variations across the escarpment, particularly in regard to permeability changes induced by increased clay content within shales, may provide conditions for groundwater discharge (Shaw and De Vries, 1988). Buckley (1984) notes that the Karoo coal measures in this area only allow circulation of groundwater in fractures and bedding planes. Additionally the variations in depth of pre-Kalahari topography beneath valley floors indicated on figure 6.8 may also cause movement of groundwater towards the surface. However, in order to confirm this statement detailed geophysical evidence of spatial variations in Karoo depth would be necessary.

In a further use of geological boreholes, the spoil adjacent to recently drilled boreholes was examined to assess duricrust variations beneath the valley floor to the southeast of the amphitheatre valley head. A more complete picture of the along-valley variations in duricrust type can be obtained if this information is viewed in conjunction with the profile studies in section 6.2.1*b(i)*.

From these observations an overall increase in $CaCO_3$ content was noted away from the valley head. Similarities were apparent between duricrusts removed from borehole 6513 and those exposed in the valley head some 1.5 km to the northwest. In both cases a crystalline GS- to F-fabric silcrete is found. This is in contrast to the pink silica-cemented silcrete breccia found in spoil at borehole 6479. Borehole spoil further "upstream" of the valley head is mainly poorly indurated calcrete, suggesting that silcrete may be only associated with areas close to the valley head. This association may be partly due to a forced rise of groundwater as a result of permeability reductions in the Karoo bedrock close to Letlhakeng (Shaw and De Vries, 1988), or because of the rise of the pre-Kalahari surface seen in figure 6.8. However, whilst the study of borehole spoil may be useful it is impossible to determine the depth from which spoil has been drilled, and only qualitative judgements can be made.

(d) Duricrusts in Letlhakeng Valley 2

The calcrete profiles in Valley 2 to the south of Letlhakeng are shown in figure 6.9, compiled from base maps and borehole logs given in Gwosdz (1981, 1982) and Gwosdz and Modisi (1983). The lithological logs are only sufficiently detailed to distinguish the type of calcrete present (see table 6.1). However, this study provides useful information on the variation of calcrete type away from the flank of the valley, with the boreholes sited in a grid pattern at a spacing of approximately 100 m.

In all three pairs of logs shown on figure 6.9, it can be seen that the profile nearer to the valley course contains more hardpan calcrete. It is also apparent from boreholes 2 & 5 and 1 & 6 (numbering by Gwosdz and Modisi, 1983) that the amount of partially calcified sand is greater further away from the valley. This would suggest that the calcrete types are related to the presence of the valley, although not as clearly as in the Rooibrak valley (above). However, there is evidence for thickening of duricrusts in proximity to and beneath the valley.

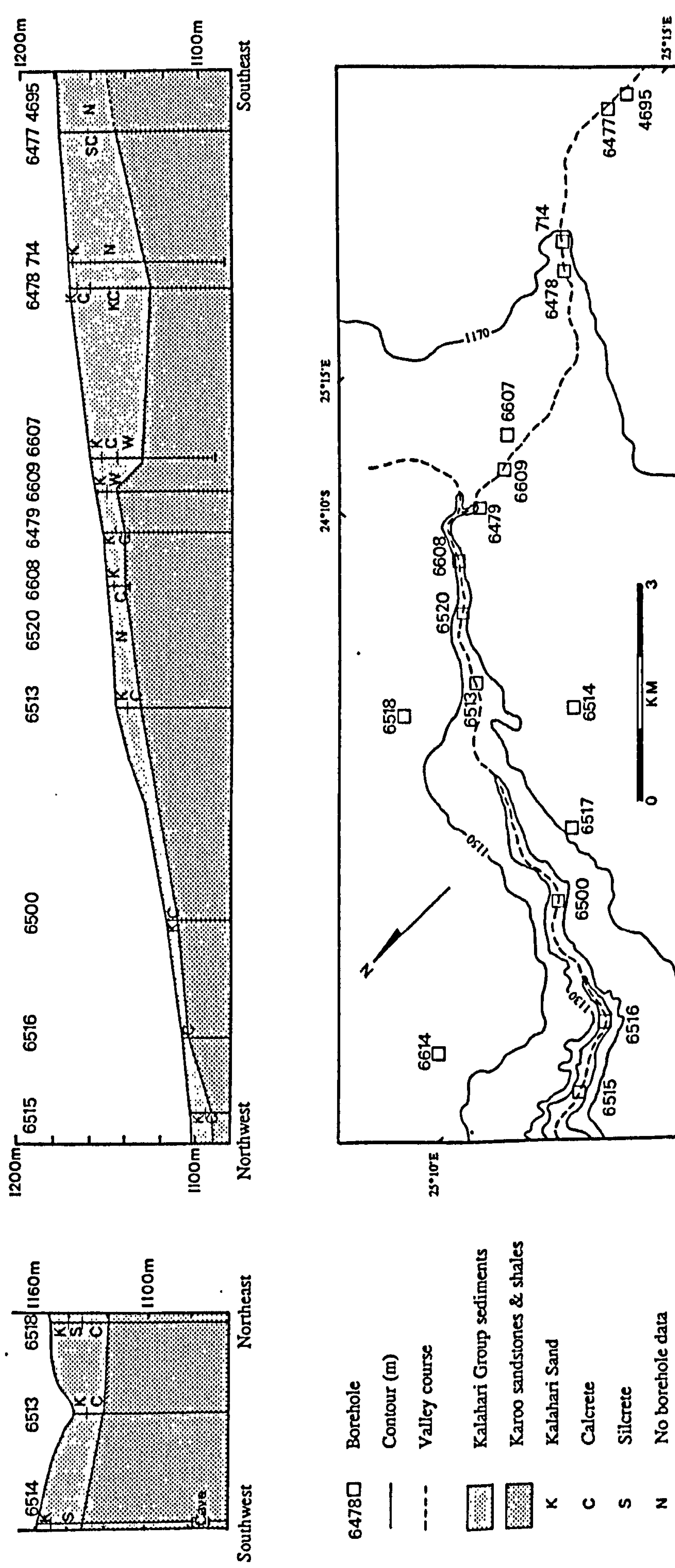


Figure 6.8: Variations in the thickness of the Kalahari Group sediments beneath the Gaotlhobogwe Valley (Letlhakeng Valley 1).

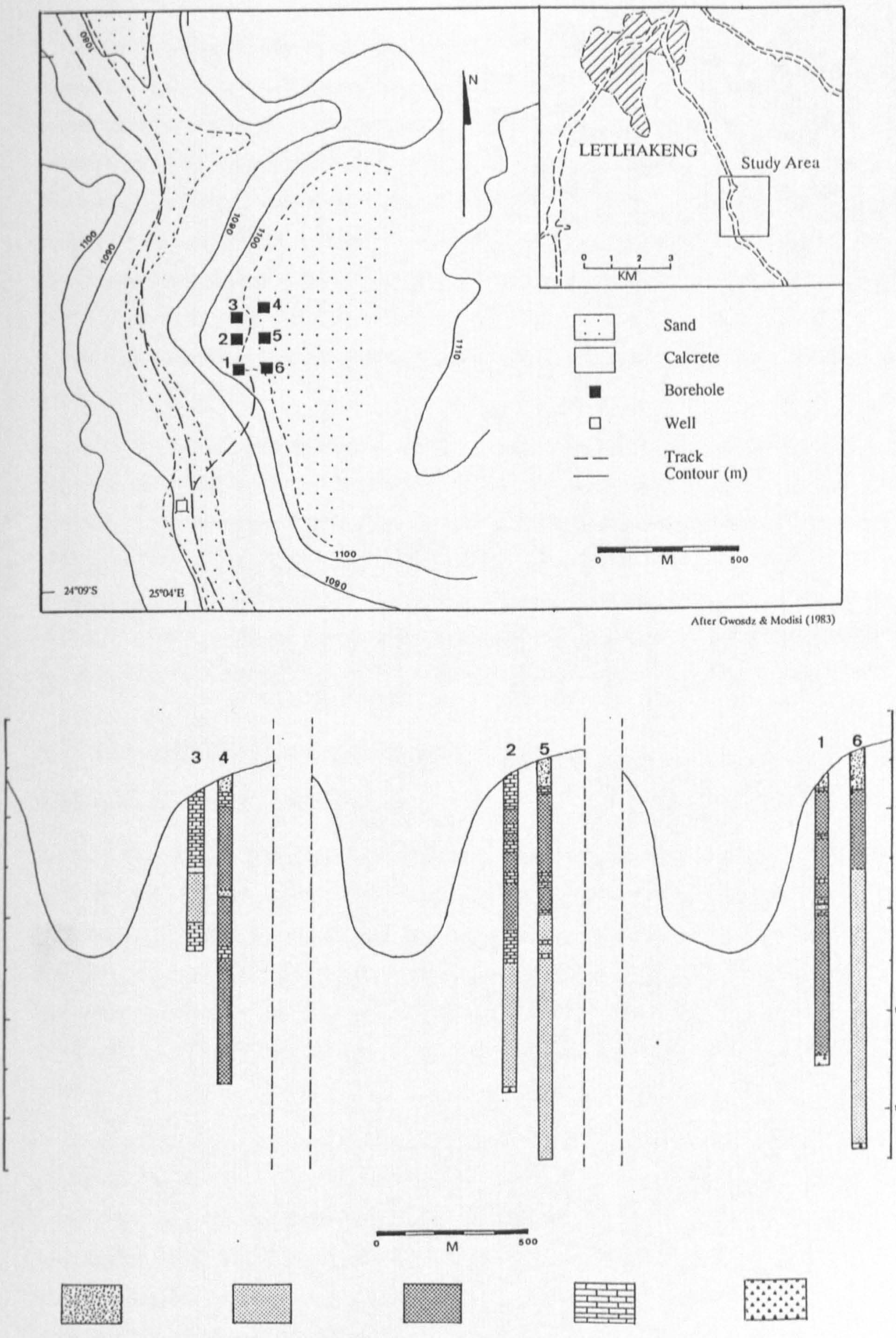


Figure 6.9: Calcretes in Letlhakeng Valley 2 (after Gwosdz and Modisi, 1983).

There appears to be little stratigraphical correlation between any of the cores, in particular the hardpan layers, although this may be a result of logging accuracy. If hardpan layers formed in close proximity to a former water table it would be expected that some form of correlation would be evident. Also of interest are silica enriched layers identifiable from chemical analyses of cores by Gwosdz and Modisi (1983). These siliceous layers are found at a depth of 12.65 to 12.90 m in borehole 1 and at 9.05 to 9.45 m in borehole 3, coinciding with the base of a hardpan layer in both cores. There is, however, no apparent relationship between these layers as they occur at approximately 1090 m and 1094 m asl in boreholes 1 and 3 respectively. This lack of correlation is consistent with studies of silicification in Australian groundwater calcretes, where diagenetic silica was present as lenses and laterally discontinuous layers (Arakel *et al.*, 1989).

From these results, it is difficult to assess whether the valley has been incised through a layer of duricrust or whether duricrust has formed as a direct result of the presence of the valley. However, the fact that the calcrete appears to be thickest near to the thalweg and apparently becomes less consolidated further away from the valley axis suggests that the duricrust formed in the presence of the valley. Also, the absence of correlated horizons within the profile tends to refute the idea that the valley cut through a pre-existing sequence of duricrusts. As discussed above, it is still difficult to determine whether this apparent lack of correlation is due to alteration of earlier duricrusts by laterally or vertically moving groundwater.

6.2.3 The analysis of duricrusts in thin-section

(a) Rationale and methodology

The preceding sections have described variations in duricrust morphology at a macro- and meso-scale, with particular reference to variations in duricrust types associated with *mekgacha*. The majority of previous studies of duricrust morphology, however, have concentrated upon variability within profiles and at a microscopic level. Petrographic studies at a micro-scale (e.g. Summerfield, 1978; Arakel *et al.*, 1989) have been used to assess the origin and mode of development of duricrusts from mineralogy and sequences of silicification, with particular attention paid to the environmental conditions under which formation took place.

In all studies the overall form of a duricrust, especially its fabric, the weathered status of any skeletal material and the constituents of the cement are taken into account. In the case of silcretes, the presence of specific diagenetic features have been considered diagnostic of the environment of formation, with Summerfield (1982) particularly noting the importance of the presence/absence of colloform features, types of mineral overgrowths and mineralogical sequences within vugh fills. The identification of the mode of origin of calcretes, particularly assessing whether ancient specimens developed as groundwater calcretes or by pedogenic processes, is more problematic (Wright and Tucker, 1991). However, as noted above, the size of matrix crystals, sequences of silicification, the presence of organic material or biogenic structures and laminar features are likely to be important indicators of origin.

In order to evaluate the relationship between Kalahari duricrusts and *mekgacha*, thin-sections of samples (approximate thickness 60 µm) were prepared using standard techniques, and were analysed with a polarising monocular microscope. Friable calcrete samples from profiles LET V3 B, OKWA 3 and OKWA 4 were impregnated with resin prior to thin-sectioning, as they were otherwise too soft for grinding. In all cases, samples were of fresh material which had not been subject to surface weathering. Analysis of thin-sections was undertaken in a random order, with a detailed petrographic description made of each sample. Terminology used in the description of thin-sections is shown in table 6.6.

An approximate indication of mineral composition was made for each sample by a "grain-counting" technique. 200 points were randomly selected on each section using a point-counter and the material at each point identified and classified into one of the following categories; skeletal grains (quartz grain, shell fragment, pebble, other), matrix material (macro-quartz cement, micro-quartz cement, carbonate cement, silica void-fill, carbonate void-fill) or void space. These point counts were subsequently converted into percentages and results graphically presented as total silica and total carbonate (figures 6.10 to 6.15).

Table 6.6: Terminology employed in duricrust thin-section description (after Summerfield, 1983*c*).

Mineral form	Crystal morphology
Megaquartz	Coarse equant (crystal width > 20 µm)
Microquartz	Fine equant (crystal width < 20 µm), typically with needle-point extinction
Length-fast chalcedony	Fibrous (epsilon vibration normal to long axis of fibres)
Length-slow chalcedony	Fibrous (epsilon vibration parallel to long axis of fibres)
Disordered chalcedony	Fibrous, but with undulating extinction
Spherulitic chalcedony	Ovoid silica patches with cross-shaped extinction
Cryptocrystalline silica	Individual crystals not resolvable in thin-section
Opalline silica	Microscopically amorphous to fairly coarsely crystalline
Microcrystalline calcite (Micrite)	Individual crystals not resolvable in thin-section
Sparry calcite	Coarse equant (crystal width > 20 µm)

Table 6.7: Details of duricrust and bedrock sample sites at which specimens were collected for analysis, together with numbers of samples from each location analysed in thin-section (TS) and by x-ray fluorescence (XRF).

Sample Number		Analysis TS	Analysis XRF	Location of sample site	Grid Reference of sample site
LET V1	A1-A25	15	8	Letlhakeng Valley 1 Head	24°09'35"S 25°12'00"E
LET V1	B1-B22	13	7	Letlhakeng Valley 1 Head	24°09'35"S 25°12'00"E
LET V1	C1-C21	14	-	Letlhakeng Valley 1 Transect 2 North flank	24°09'27"S 25°11'28"E
LET V1	13C	-	1	Letlhakeng Valley 1 Transect 3 North flank	24°09'16"S 25°11'14"E
LET V1	106	-	1	Letlhakeng Valley 1 Borehole 6513	24°10'00"S 25°12'00"E
LET V1	24A	-	1	Letlhakeng Valley 1 Transect 22	24°06'09"S 25°06'15"E
LET V2	B1-B25	18	3	Letlhakeng Valley 2 Transect 8 East flank	24°08'30"S 25°03'58"E
LET V3	A1-A15	8	-	Letlhakeng Valley 3 Transect 12	24°09'16"S 25°00'12"E
LET V3	B1-B22	12	-	Letlhakeng Valley 3 Transect 18 West flank	24°08'25"S 25°00'19"E
LET V3	C1-C30	14	1	Letlhakeng Valley 3 Transect 19 East flank	24°08'04"S 25°00'20"E
Okwa	2A-2G	4	-	Okwa Valley (Gobololo)	22°24'00"S 20°54'15"E
Okwa	3A-3G	4	-	Okwa Valley (12 km W of Ghanzi-Jwaneng road)	22°24'00"S 21°43'50"E
Okwa	4A-4G	4	4	Okwa Valley (Tswaane)	22°24'20"S 22°51'10"E
Okwa	68	-	1	Okwa Valley (Tswaane)	22°24'20"S 22°51'10"E
Auob	115A-115J	10	2	Auob Valley (Kalk-heuval Farm)	24°45'45"S 18°44'25"E
Moselebe	108	-	1	Moselebe Valley	25°17'30"S 24°37'00"E
Lephephe	100	-	1	Escarpment SE of Lephephe	23°34'00"S 26°02'00"E
Bosutswe	200	-	1	Escarpment at Bosutswe	21°55'00"S 26°30'00"E

(b) Sample site locations

In total, 116 thin-sections were analysed from eleven sample profiles distributed between the Okwa, Auob and Letlhakeng valleys. As noted in section 6.2.1, these valleys contain extensive exposures of duricrusts. The exact locations and numbers of sections examined for each profile are indicated in table 6.7; the locations of transects in the Letlhakeng valleys are also shown in figures 6.4 and 6.5.

Six of the eleven sample profiles were cliffs or vertical exposures of duricrust; LET V1 A, LET V1 B, LET V3 A, LET V3 B, OKWA 4 and AUOB 115. The remaining profiles were sampled down valley-side slopes. Profile LET V1 C consisted of a 1.5 m vertical exposure above 5.25 m of duricrust with a surface slope of 21°. LET V2 B comprised 8.7 m of material with a surface slope of 9°, and was sampled directly downslope of borehole 1 on figure 6.9. LET V3 C was a complex profile, with 13.5 m of duricrust exposed in slopes of between 9° and 13°. Profile OKWA 2 was sampled through 0.6 m of terrace calcrete at a slope of 6°, whilst OKWA 3 sampled 0.8 m of calcrete exposed in a pan flank at slopes of between 4° and 9°.

(c) Results

The general duricrust characteristics, fabric and specific diagenetic features will be now be described for each of the eleven profiles, together with details of intra-profile variations. Full descriptions are given in Appendices 1 to 5. The following section considers the environmental implications of these results and their significance to *mekgacha* development.

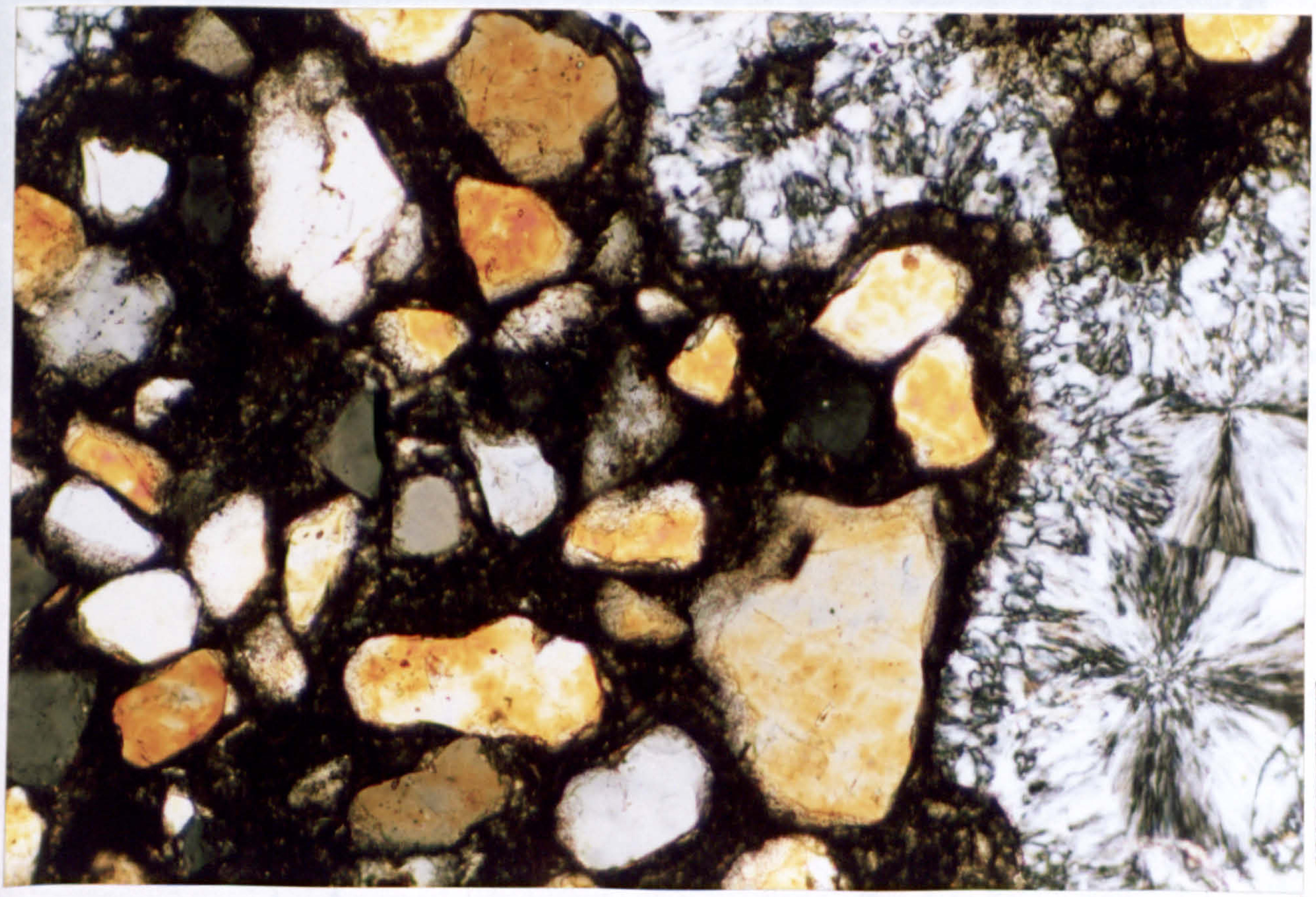
Letlhakeng Valley 1 Profiles A and B

Both profiles consist of vertical exposures of GS- to F-fabric silcretes with skeletal grain contents of between 32.5 and 70%. Grains are dominate by quartz, with additional minor quantities of plagioclase and microcline feldspar, augite, tourmaline, apatite, opaque minerals and quartzite. Some skeletal grains show signs of minor fretting or dissolution, but this is relatively uncommon.

The silcrete is densely cemented throughout both profiles with minimal pore space. The matrix is dominated by a mixture of cryptocrystalline quartz and disordered chalcedony with isolated areas of microcrystalline quartz identifiable by its needlepoint extinction. In Profile B, the matrix developed in the order: opalline or cryptocrystalline silica as a grain coating (often Fe-stained) followed by cryptocrystalline silica or microquartz and then disordered or length-fast chalcedony.

In addition to the complex matrix sequence, void linings and fills are equally complex, although generally show a sequence of opalline silica to chalcedony to micro- or megaquartz. There is also a pattern within chalcedony precipitation, with disordered chalcedony commonly followed by length-fast chalcedony and finally spherulites of chalcedony which sometimes coalesce at the centre of voids. The full sequence of silicification was not encountered in any one void or fissure, and distinct breaks in precipitation were evident in most void linings. A maximum number of eight individual stages of deposition could be identified in any one void (sample 18 Profile A; plate 6.1), consisting of opalline silica to length-fast chalcedony to opalline silica to length-fast chalcedony to opalline silica to length-fast chalcedony to length-slow chalcedony to microquartz and megaquartz.

The major difference between the void-fills of Profiles A and B was the presence of late-stage microcrystalline calcite at the centre of voids in Profile B below a height of 1.75 m (i.e. sample 16 downwards). There was no evidence of replacement between the calcitic infill and underlying opalline silica, the change between the two precipitates being abrupt. The only evidence for replacement in either profile is in sample 1 of Profile B (height 6.25 m) where the disordered chalcedony void-fill and matrix material appear to merge in places.



0.5 mm

Plate 6.1: Thin-section LET V1 A18: GS- to F-fabric silcrete, with quartz grains set in a cryptocrystalline quartz and disordered chalcedony matrix. Void fill consists of (inwards) opalline silica to length-fast chalcedony to opalline silica to length-fast chalcedony to opalline silica to length-fast chalcedony to length-slow chalcedony (with extinction crosses) to microquartz and megaquartz. (Cross polars. x 10 magnification).

Letlhakeng Valley 1 Profile C

Profile C is a complex silcrete and calcrete profile (figure 6.10), with extensive evidence for diagenetic alteration and silicification. F- or M-fabrics are most common (although one section exhibited a GS-fabric), with a skeletal quartz content of between 3 and 41%, but more commonly below 15%. Other mineral constituents include minor feldspars and opaques.

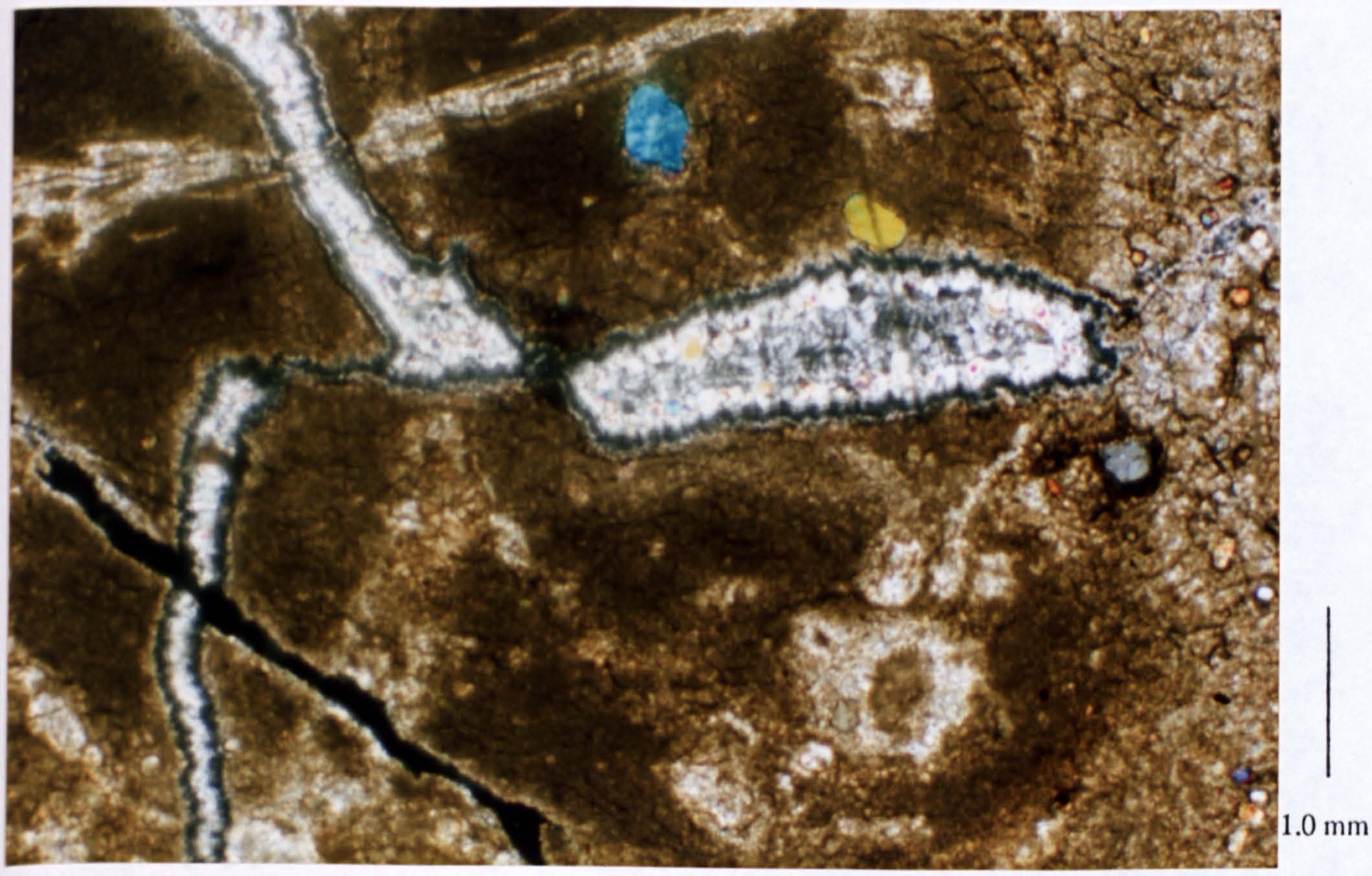


Plate 6.2: Thin-section LET V1 C3: M-fabric calcrete, with a matrix of microcrystalline to sparry calcite. Main void fill consists of layered opalline silica and macroquartz, and cuts across a saparry calcite void fill. (Cross polars. x 5 magnification).

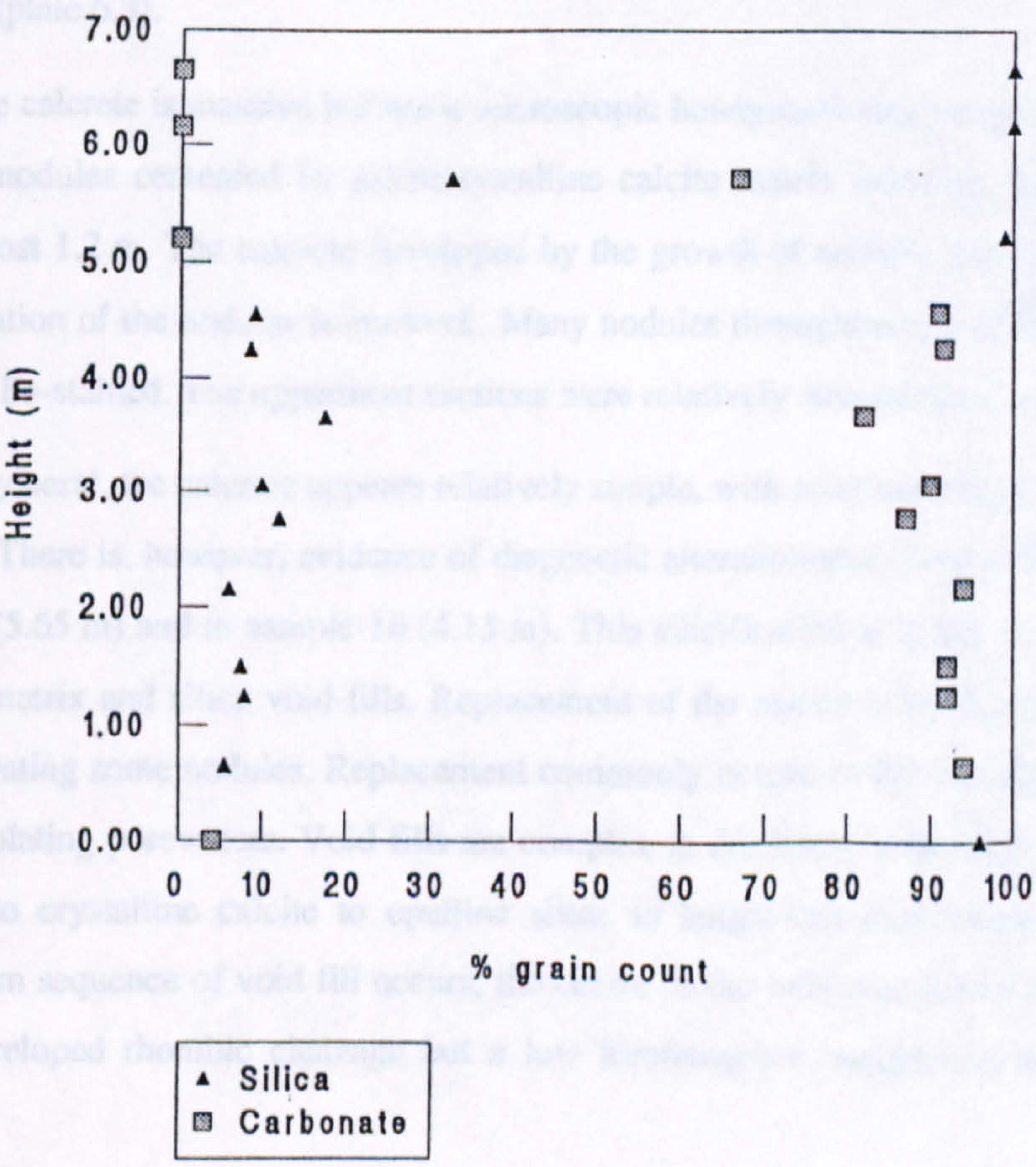


Figure 6.10: Variations in silica and carbonate content for Letlhakeng Valley 1 Profile C.

The matrix composition of samples 1, 2, 4 and 21 (at heights of 6.65, 6.15, 5.20 and 0.65 m respectively) is dominantly irregularly distributed cryptocrystalline silica, disordered chalcedony and microquartz. However, the remainder of the profile has a matrix of microcrystalline or sparry calcite with a sugary texture. Sparry calcite occurs in patches up to 6 mm diameter. There is evidence throughout the profile for replacement of the matrix calcite by disordered chalcedony, with occasional patches of chalcedony away from voids.

Void fills are equally complex, and containing both silica and carbonate. The simplest lining consists of microcrystalline calcite, but sequences of microcrystalline calcite to disordered chalcedony and also complex opal-chalcedony-megaquartz fills occur. The most complex sequence consists of calcite overlain by five distinct layers of opalline silica, with megaquartz and finally further microcrystalline calcite at the centre (plate 6.2). A 0.25 mm layer of sparry calcite crystals oriented perpendicular to the void wall also occurs in one void, sandwiched between layers of chalcedony.

Letlhakeng Valley 2 Profile B

This sloping profile contains 8.7 m vertical sequence of M- to F-fabric calcrete, with a skeletal quartz content varying between 1 and 21% as indicated on figure 6.11. Other minerals present in minor quantities include plagioclase feldspar, opaques and isolated grains of tourmaline. Skeletal grains show no evidence of grain fretting or dissolution. Individual valves of bivalve molluscs up to 0.3 mm long also occur (samples 3, 4, 13, and 23), with the shell material diagenetically altered to crystalline or microcrystalline calcite (plate 6.3).

The calcrete is massive but has a microscopic honeycomb appearance consisting of coalesced micritic calcite nodules cemented by microcrystalline calcite matrix material, which was present in all but the uppermost 1.2 m. The calcrete developed by the growth of nodules (up to 0.3 mm in diameter) with later cementation of the nodular framework. Many nodules throughout the profile show a calcite coating which is often Fe-stained. The uppermost sections were relatively structureless, consisting of micritic calcite.

In general, the calcrete appears relatively simple, with most void-linings consisting of microcrystalline calcite. There is, however, evidence of diagenetic alteration and silicification between samples 9 (5.90 m) and 11 (5.65 m) and in sample 16 (4.15 m). This silicification is in the form of partial replacement of the calcite matrix and silica void-fills. Replacement of the matrix is by disordered chalcedony with opalline silica coating some nodules. Replacement commonly occurs in the vicinity of voids, suggesting alteration by percolating porewaters. Void fills are complex in this zone, with sequences including microcrystalline calcite to crystalline calcite to opalline silica to length-fast chalcedony to sparry calcite. Where this maximum sequence of void fill occurs, the centre of the void consists of a single crystal of calcite with a well-developed rhombic cleavage but a low birefringence, suggesting possible further replacement by silica.

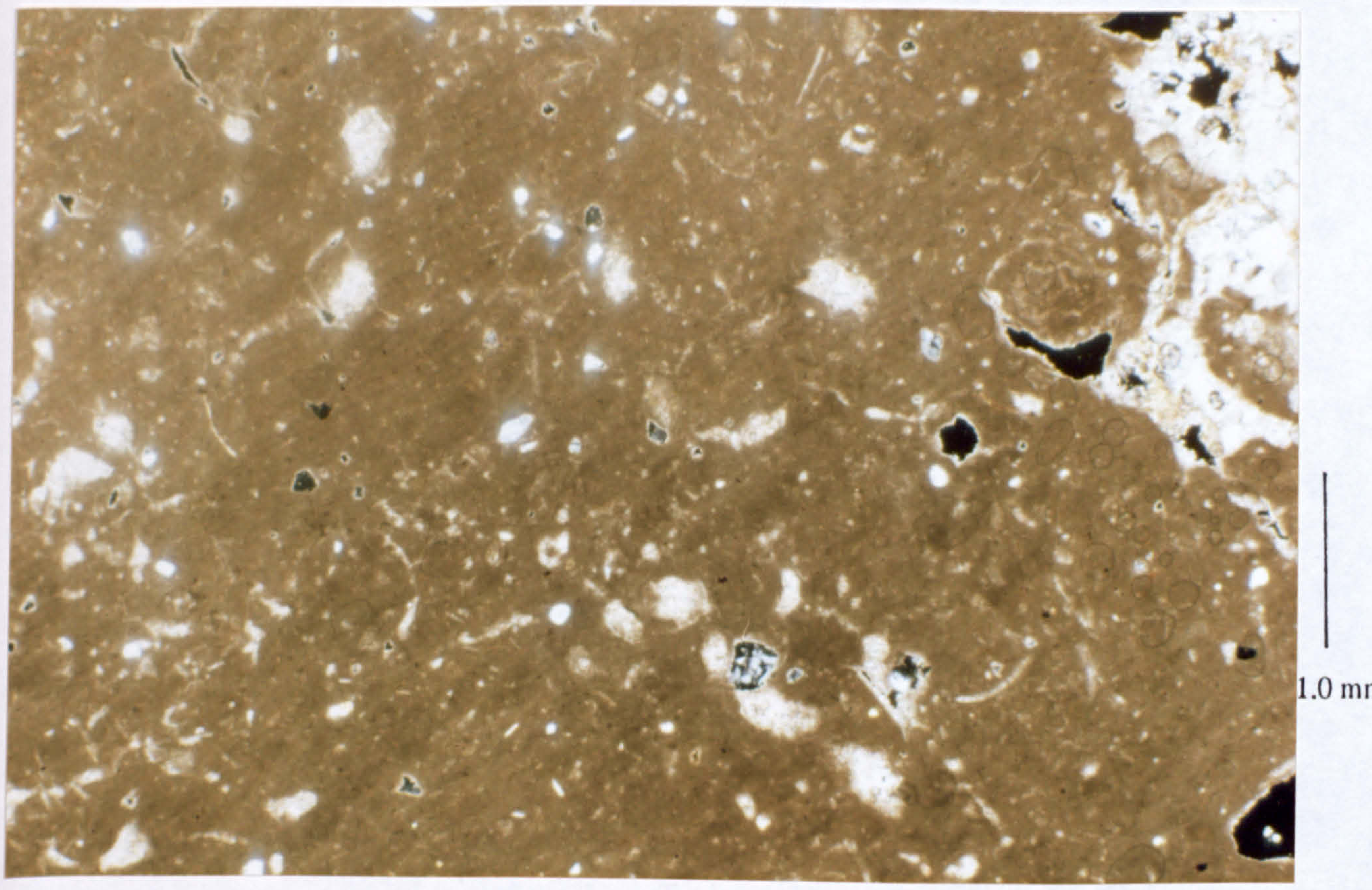


Plate 6.3: Thin-section LET V2 B23: M-fabric calcrete, with quartz grains and skeletal shell material in a microcrystalline calcite matrix. (Cross polars. x 5 magnification).

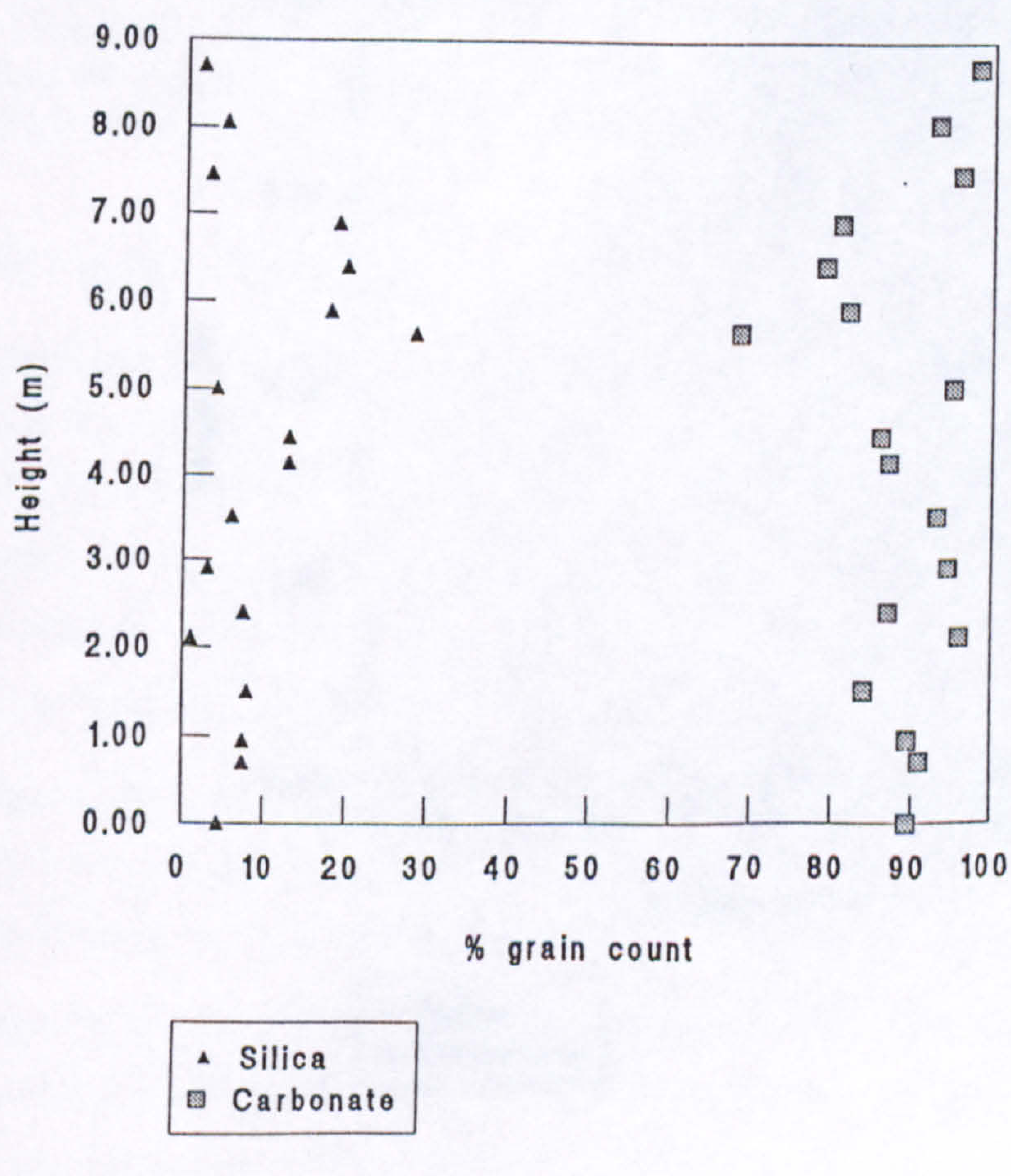


Figure 6.11: Variations in silica and carbonate content for Letlhakeng Valley 2 Profile B.

The most complex thin-section was from sample 27 (0.95 m), which consists of glaebules with brown-stained edges and nodules set in matrix material. The glaebules encompass microcrystalline calcite which in turn surrounds quartz. In places, the "walls" of the glaebules are breached by fissures which radiate from the centre of the glaebule and appear to represent expansion/shrinkage cracks. The cracks have been filled by microcrystalline calcite which also surrounds the nodules.

Letlhakeng Valley 3 Profile A

Valley 3 profile A comprises silcrete overlying calcrete (figure 6.12), with evidence that the silcrete has developed by diagenetic replacement. The duricrusts have skeletal quartz (plus opaques) contents of between 2.5 and 32% and exhibit predominantly F- but also M-fabrics. The majority of the matrix material is microcrystalline calcite, with the exception of sample 1 (3.5 m) which is entirely cryptocrystalline silica and micro-quartz. All sections have a nodular appearance producing a framework which has subsequently been infilled by cementing material.

Replacement of the calcite cement by cryptocrystalline silica is evident throughout the profile, with the matrix of sample 9 (1.5 m) comprising 36% silica and 56% carbonate. Throughout much of the matrix the nodules are cemented by a coating of microcrystalline calcite. However, where replacement has occurred opalline silica coats nodules.

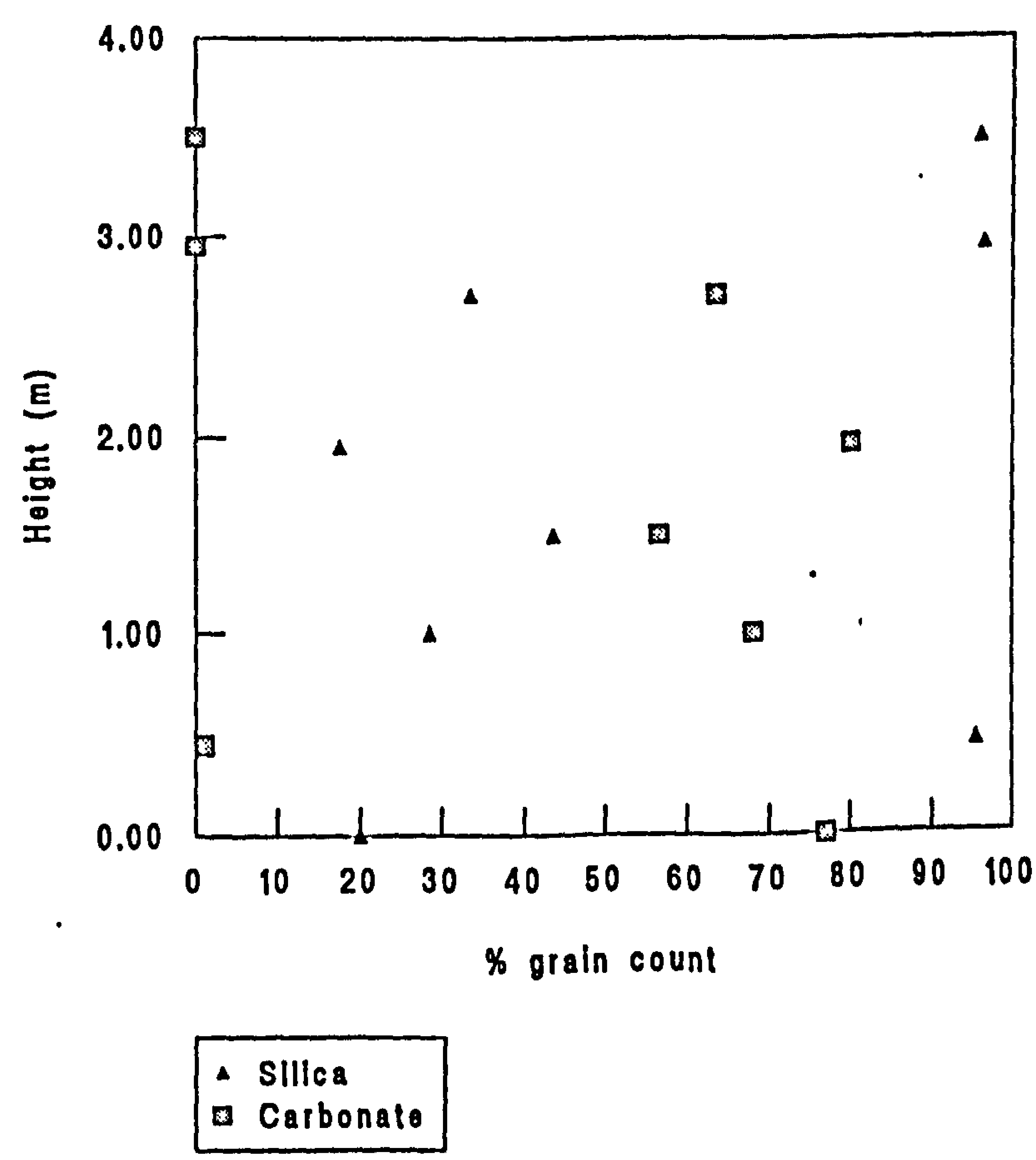


Figure 6.12: Variations in silica and carbonate content for Letlhakeng Valley 3 Profile A.

Void-fills are dominated by various forms of silica, with the most common sequence being opalline silica to disordered or length-fast chalcedony. Additionally, one void contained sparry calcite at its centre (plate 6.4). The most complex void fill consisted of opalline silica to spherulitic length-fast chalcedony to opalline silica, with length-slow chalcedony in the centre.

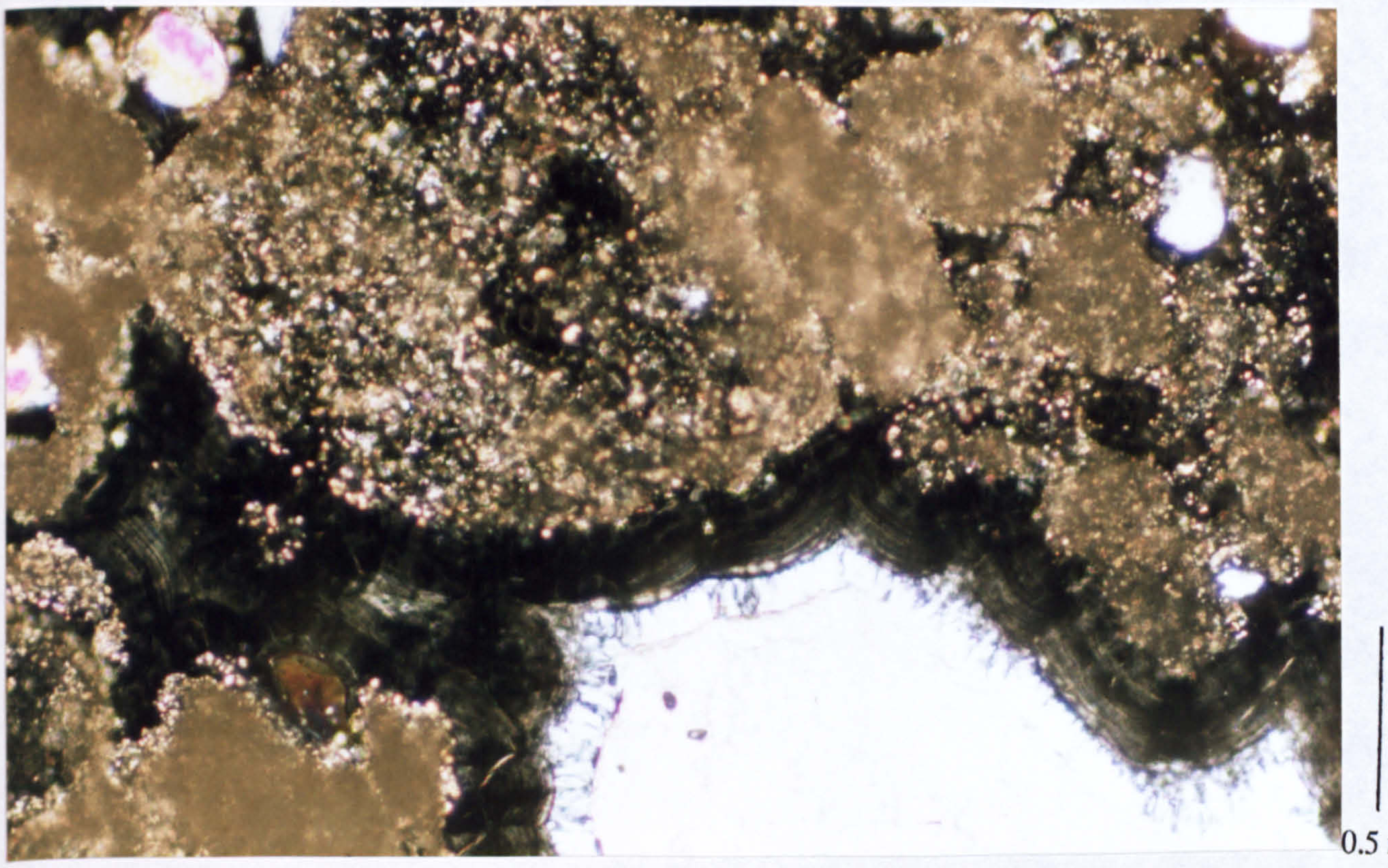


Plate 6.4: Thin-section LET V3 A11: M-fabric calcrete with chalcedony replacement of calcite matrix material. Void fill consits of (inwards) opalline silica to length-fast chalcedony to (white) sparry calcite. (Cross polars. x 10 magnification).

Letlhakeng Valley 3 Profile B

Profile B consists of an authigenic nodular calcrete with some replacement and silicification apparent in the upper part of the profile, but a declining downward silica content (figure 6.13). The calcretes contain between 0.5 and 13% quartz grains in an F- to M-fabric, with additional opaques and feldspars in minor quantities. Shell material is also present in samples 17 to 22 (i.e. below a height of 1.55 m), with all shells replaced by microcrystalline calcite. The shells consist of single or paired mollusc valves up to 1.1 mm in length, with one bivalve in sample 20 having its valves still hinged together.

The calcrete matrix is dominated by nodules of microcrystalline calcite cemented by interstitial calcite. Only in the uppermost samples (5.00 m and above) is there evidence of localised replacement of the matrix material by micro-quartz and cryptocrystalline silica.

Most voids are lined by microcrystalline calcite, although some contain siliceous precipitates. There is a difference between linings of linear and sub-circular voids in the upper parts of the profile. Both linear and sub-circular void linings include opalline silica and length-fast chalcedony but also opalline silica and calcite. However, opalline silica, length-slow chalcedony and calcite additionally occur in pores. Opalline grain coatings are uncommon but occur near areas of matrix replacement.

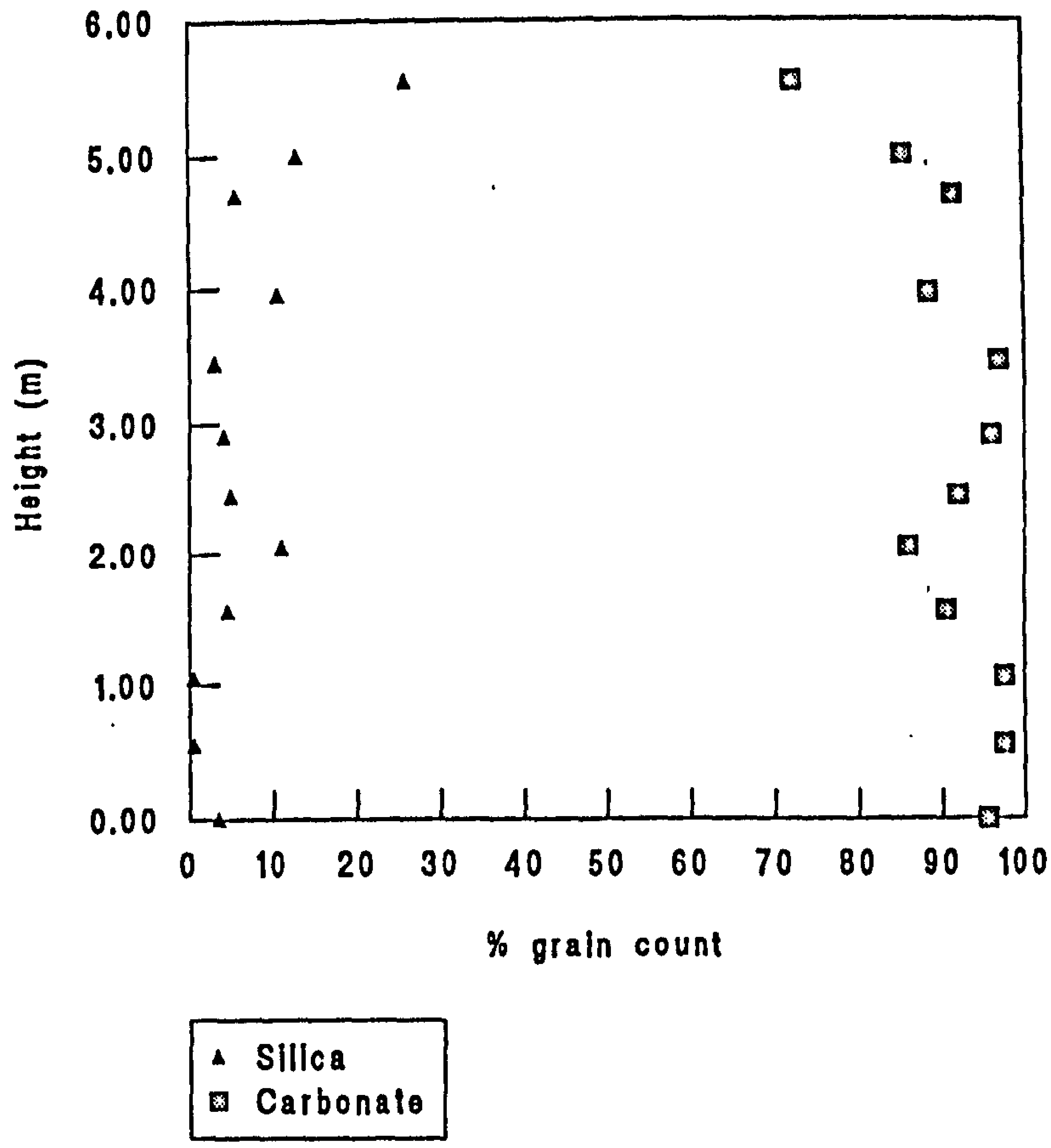


Figure 6.13: Variations in silica and carbonate content for Letlhakeng Valley 3 Profile B.

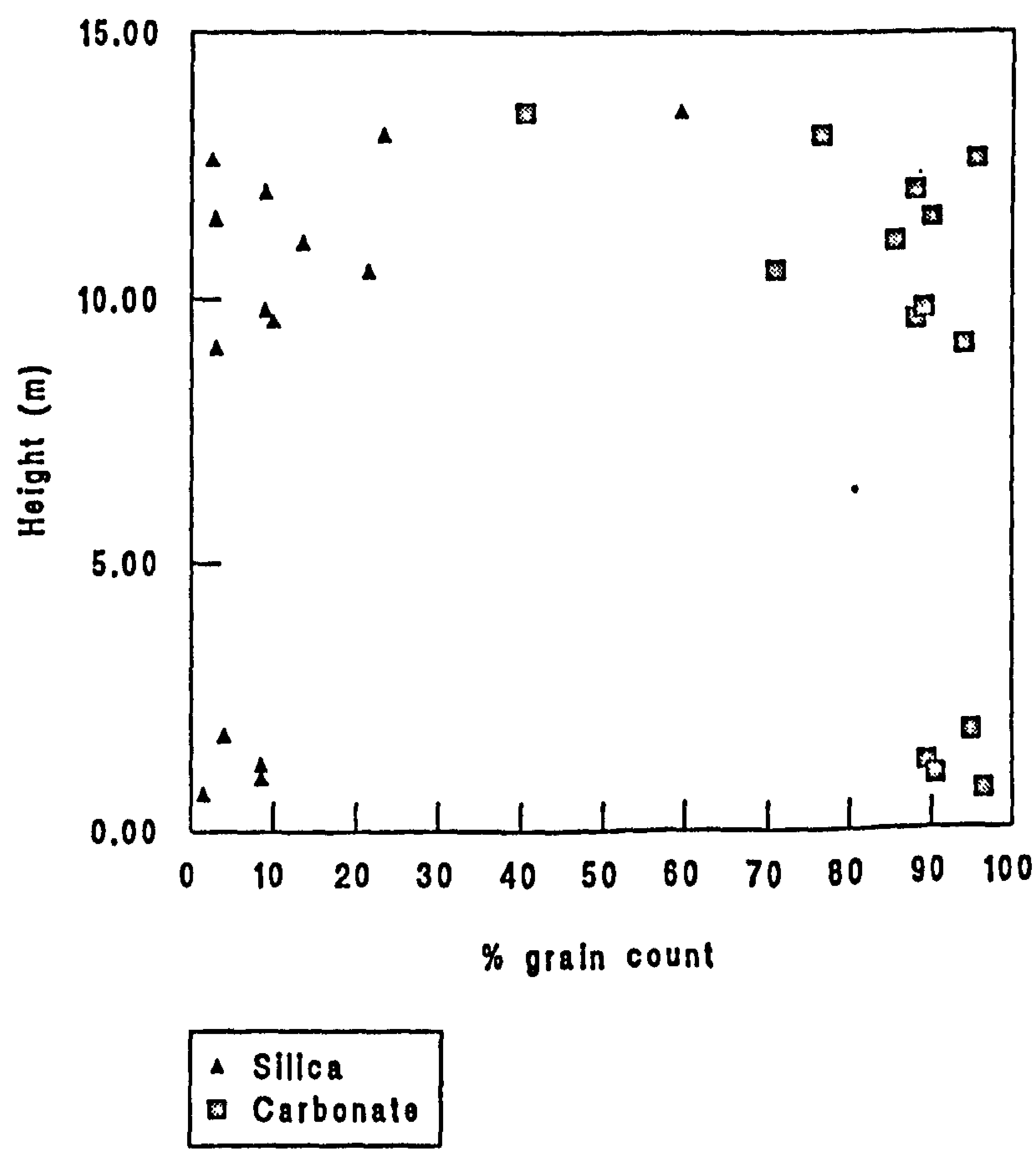


Figure 6.14: Variations in silica and carbonate content for Letlhakeng Valley 3 Profile C.

Letlhakeng Valley 3 Profile C

Samples in Profile C (figure 6.14) are predominantly simple authigenic nodular or pisolitic calcretes with a skeletal quartz content of between 1.5 and 12.5%, together with minor quantities of opaque minerals. The calcretes have a dominantly F- to M-fabric. The general structure of all samples is a cemented nodular latticework with interstitial microcrystalline and sparry calcite. Pisolites reach diameters of up to 3.9 mm in lower samples. In the uppermost sections (samples 1 and 4) there is evidence of replacement of the matrix material by disordered chalcedony, particularly in the immediate vicinity of voids, with one entire nodule replaced by chalcedony.

Void fills are commonly a simple lining by microcrystalline calcite, but there is evidence of circulation of siliceous porewaters, particularly in upper parts of the profile. Void fills (above 13.05 m) include linings of sparry calcite overlain by cryptocrystalline and opalline silica, with subsequent length-fast chalcedony and calcite. Length-slow chalcedony fill also occurs, sandwiched between two layers of calcite.

Auob 115

The Auob profile is highly complex, consisting dominantly of calcite but with considerable evidence of replacement and silicification. The duricrusts contain between 2.5 and 23.5% skeletal quartz grains (commonly less than 10%), with minor quantities of opaque minerals and augite, and have an F- to M-fabric (figure 6.15).

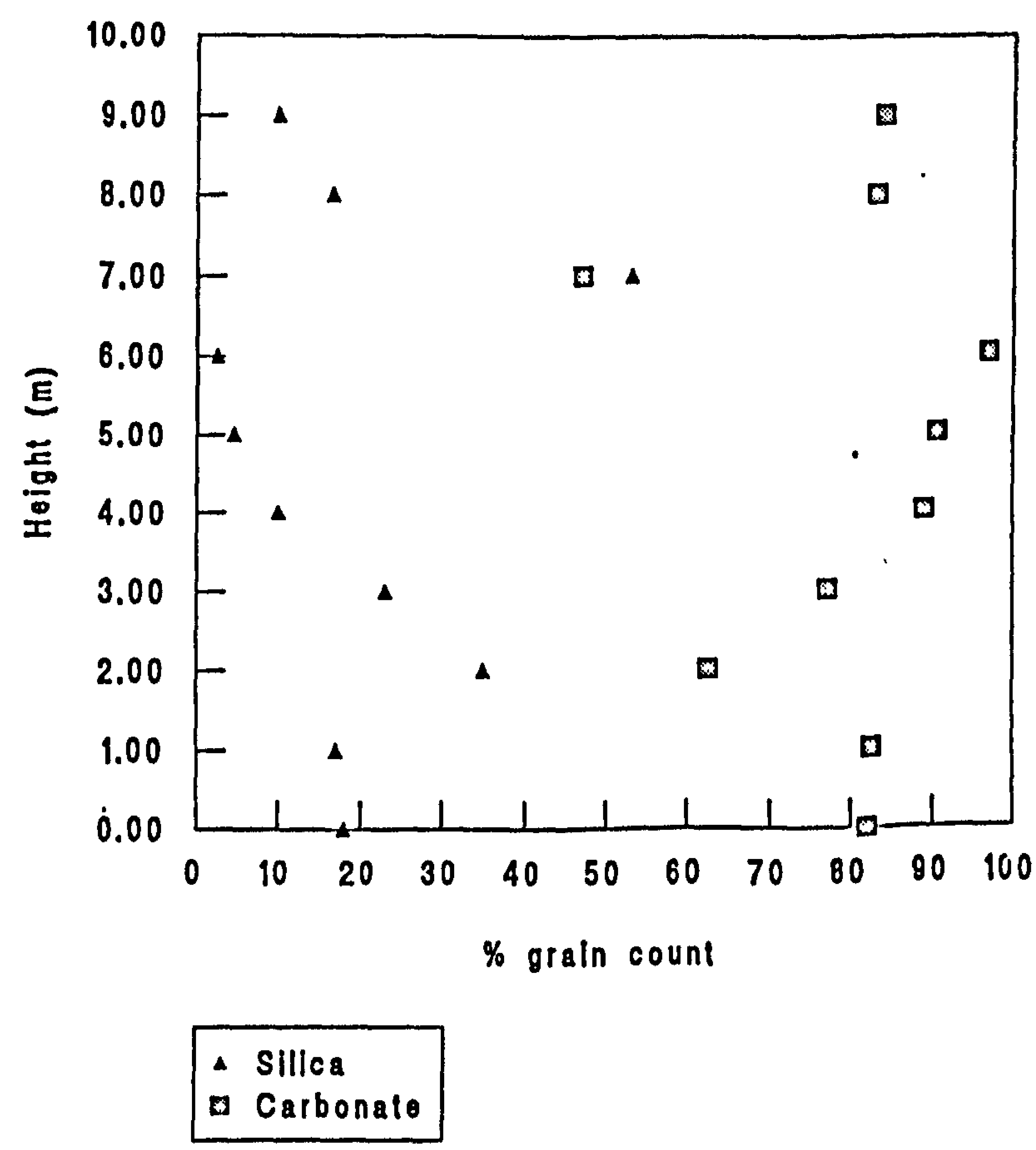


Figure 6.15: Variations in silica and carbonate content for Auob Valley Profile 115B.

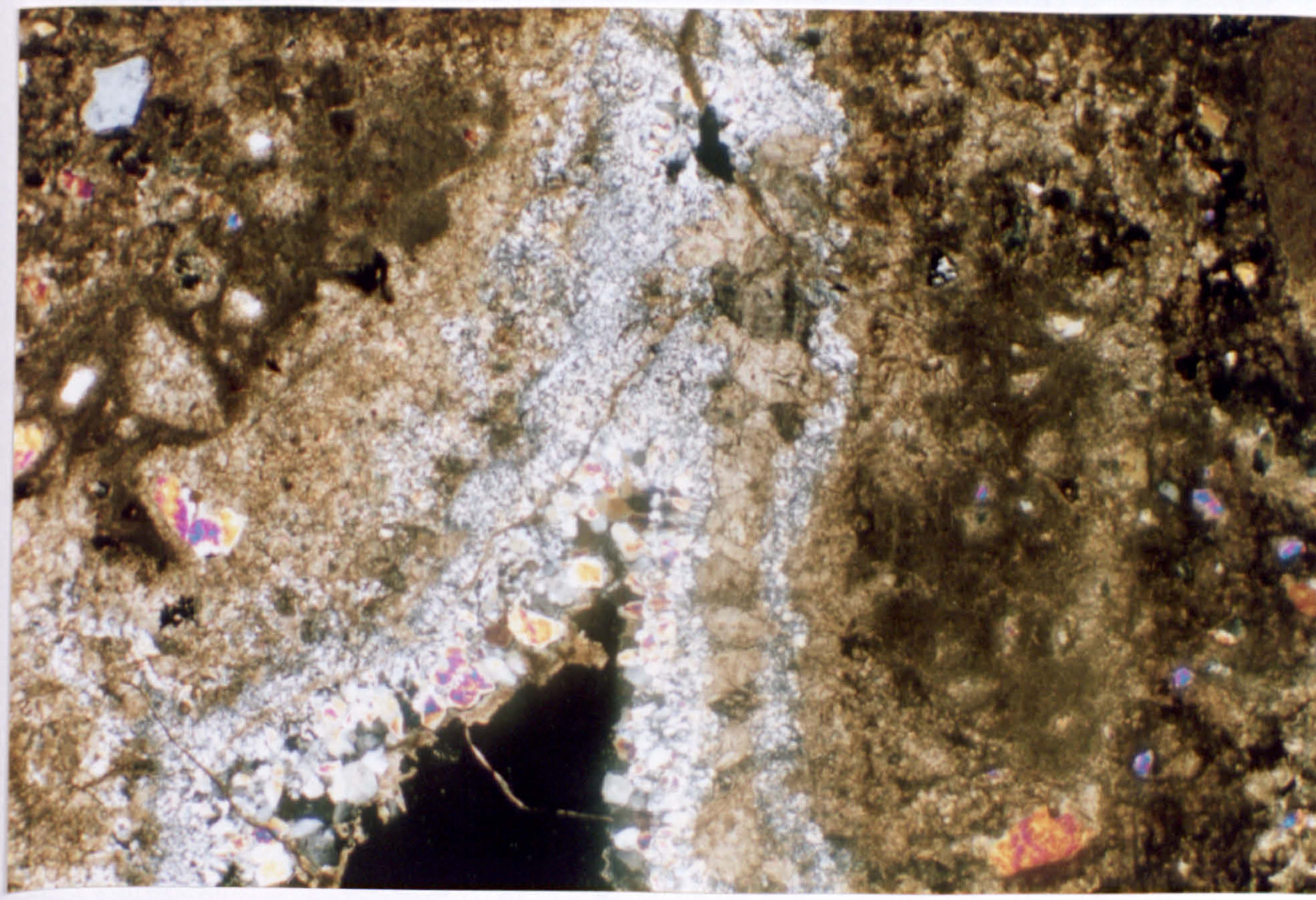


Plate 6.5: Thin-section Auob 115B: M-fabric calcrete with quartz grains in a sparry calcite matrix. Void fill consists of (inwards) disordered chalcedony to sparry calcite to disordered chalcedony to length-fast chalcedony to mega-quartz. (Cross polars. x 5 magnification).

The morphology of most samples is complex, with a mixture of nodular and crystalline material present. Many skeletal grains show sparry calcite grain overgrowths perpendicular to the mineral grain. The matrix consists of both microcrystalline and sparry calcite which shows signs of alteration to disordered chalcedony or cryptocrystalline silica above 7.00 m and below 2.00 m. Alteration appears to be most common in non-nodular areas.

Void fills are more complex, ranging from simple calcite void-linings in areas not subject to diagenetic alteration to sequences of siliceous and calcareous fill in others. Sample C contains the most complex void fill sequence and has a matrix consisting of silica and carbonate in almost equal quantities. Voids contain disordered chalcedony overlain by laminated chalcedony and further disordered chalcedony with megaquartz at a void centre. One linear void shows chalcedony grading into calcite over a distance of less than 0.1 mm (plate 6.5). Also present is one crack filled by chalcedony dissected by another crack with a similar chalcedony fill.

Okwa Profiles 2 and 3

All samples from Profiles 2 and 3 in the Okwa Valley are predominantly massive calcretes and, as such, are considered together. Profile 2 has a skeletal grain content of between 2 and 18% (M- to F-fabric) consisting of quartz grains, opaque minerals and shell fragments. Profile 3 varies between 14.5 and 28.5% (F- to GS-fabric) and in addition to the above skeletal constituents, also contains grains of quartzite, plagioclase feldspar, albite, augite and tourmaline. The shell fragments in both profiles consist of a mixture of bivalve and gastropod molluscs, including complete gastropod specimens up to 2.6 mm in length (plate 6.6).

In both profiles, the matrix consists of microcrystalline calcite, with little evidence of any structure. Void linings are also dominantly microcrystalline calcite. A feature of all samples are sub-circular "glaebule" structures, up to 7.2 mm in diameter, comprising concentrations of quartz in a microcrystalline calcite matrix encircled by Fe-stained laminar calcite. These forms appear to be the result of the infilling of pipes within the calcrete. Pipe walls appear to have been lined by stained calcite, with quartz grains subsequently washed into them and cemented.

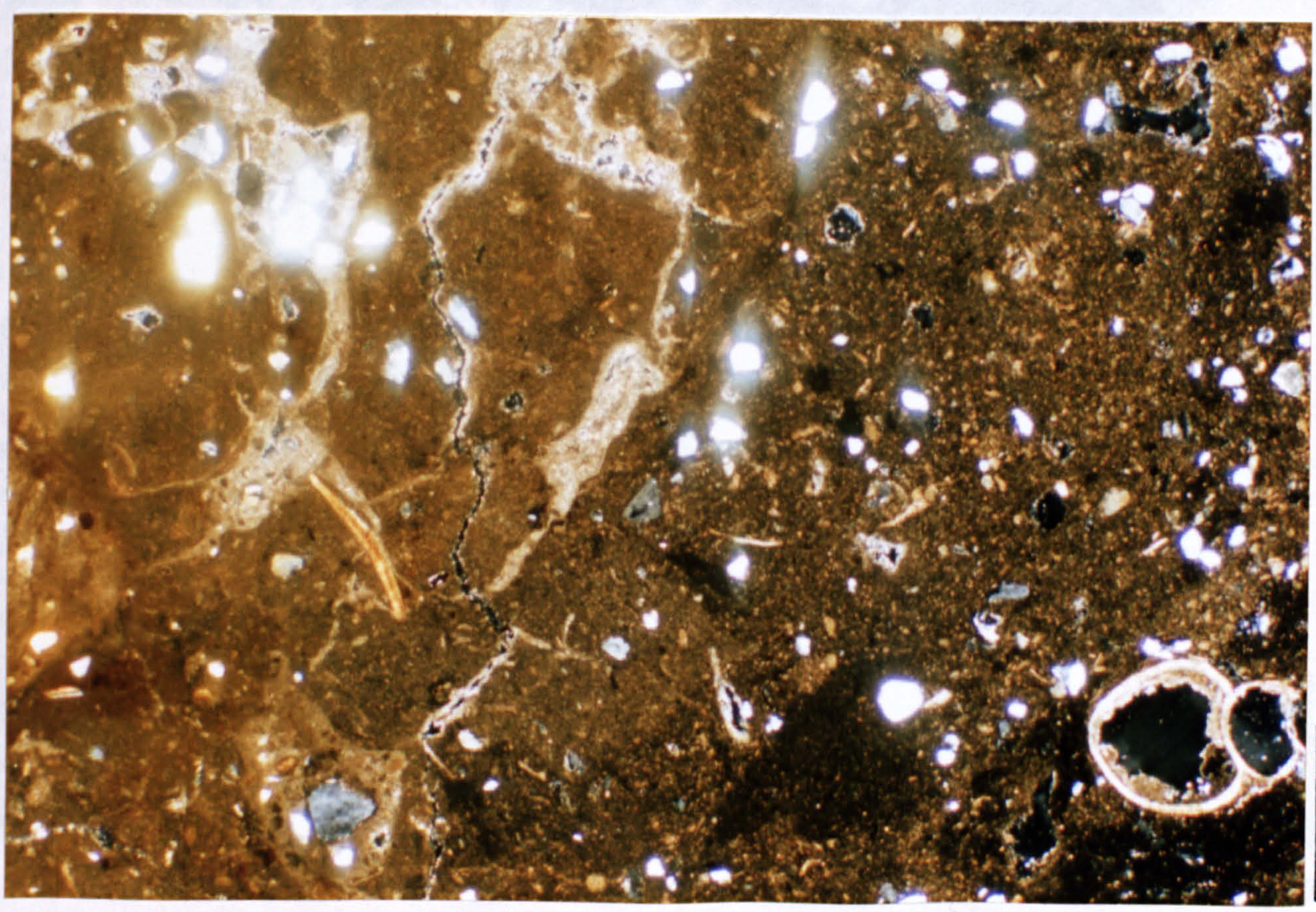


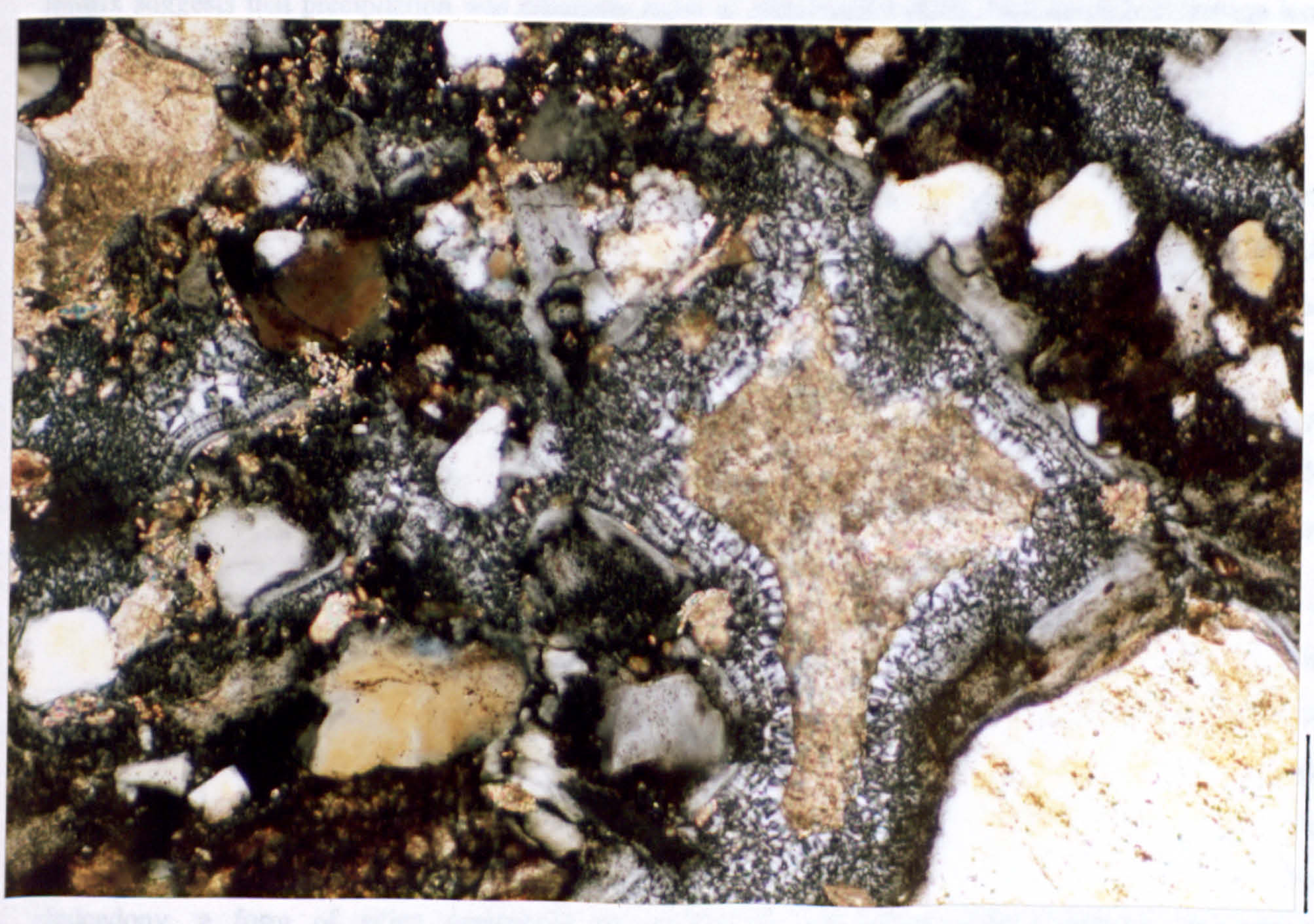
Plate 6.6: Thin-section OKWA 2C: M- to F-fabric calcrete with quartz grains and skeletal shell material. (Cross polars. x 5 magnification).

Okwa Profile 4

Profile 4 contains highly complex composite silcretes with a skeletal quartz content of between 6% at the base (F- to M-fabric) and 35.5% at the top (F-fabric). Other grains present include feldspars, augite, opaque minerals and quartzite fragments in minor quantities. There is little evidence for grain dissolution or fretting.

Upper parts of the profile have a matrix dominated by disordered chalcedony, with patches of cryptocrystalline silica and replacement calcite near to voids (plate 6.7). The F-fabric suggests that the silcrete has developed primarily by the replacement of calcite. This is further suggested by the presence of one glaebule apparently inherited from an earlier calcrete.

The lower samples are composite silcretes, with an Fe-stained cryptocrystalline silica silcrete cut by sub-parallel planar voids in which chalcedony has been deposited. These void fills contain sequences of increasingly well ordered chalcedony and quartz away from the void wall, and calcite at the void centre. One void sequence consisted of disordered chalcedony to crystalline megaquartz (growing perpendicular to the void wall) to coalescing spherulites of length-fast chalcedony to disordered chalcedony.



0.5 mm

Plate 6.7: Thin-section OKWA 4A: Complex F-fabric silcrete, with quartz grains in a disordered chalcedony, cryptocrystalline silica and replacement calcite matrix. Void fill comprises (inwards) opalline silica to chalcedony to well-organised quartz, with calcite centre. (Cross polars. x 10 magnification).

(d) Environmental implications and significance for *mekgacha* development

The results of petrographic studies of duricrusts will now be considered for the Letlhakeng, Auob and Okwa Valley profiles.

Letlhakeng Valleys

The petrographic studies from Letlhakeng Valleys 1, 2 and 3 indicate that duricrusts are just as variable in thin-section as they are in profile. However, clear distinctions can be made between the study profiles from each of the valleys, which have implications for the development of the duricrust suite.

Letlhakeng Valley 1

The silcretes at the amphitheatre valley head of Letlhakeng Valley 1 are densely cemented GS-fabric varieties. These silcretes consist of a self-supporting framework of quartz grains which has been cemented by a variety of forms of silica; they are essentially silicified sands (Summerfield, 1982).

The matrix material is dominated by cryptocrystalline quartz and disordered chalcedony, with isolated areas of microcrystalline quartz. Cementation appears to have been instigated by the initial precipitation of fine layers of opalline silica around quartz grains. The lack of well-ordered crystals of quartz within the matrix suggests that precipitation was relatively rapid or was influenced by the presence of foreign ions (Summerfield, 1982). It is also possible that insufficient time has elapsed since silica precipitation for diagenetic crystal growth to occur (Williams *et al.*, 1985).

Sequences of void fill generally reflect the opalline silica-chalcedony-microquartz-megaquartz pattern identified by Summerfield (1983*c*) and Thiry and Millot (1987), although no single void contained the full suite of silica types. This sequence shows increasing crystallographic organisation of silica and can be attributed to a decrease in the rate of movement of pore solutions as void spaces gradually decline following precipitation. Many voids contained multiple sequences of void fill, with the sequence always restarted with a layer of opalline silica; a maximum of three opalline layers occurred in sample 18 Profile A. This reflects the multiple stages of development of the silcretes, which closely parallel sequences of diagenetic silicification identified in groundwater calcretes by Arakel *et al.* (1989).

Variations in the chemistry of circulating porewaters are indicated by the presence of microcrystalline calcite at the centre of some voids, usually separated from underlying material by a distinct break in precipitation. This would imply either a shift in the pH of porewaters or that silica precipitation occurred over a limited period. Given the prevalence of calcretes away from the valley head area of Valley 1 it is likely that any late-stage diagenesis was dominated by carbonate-rich water. The fact that length-slow chalcedony, a form of silica commonly precipitated in association with carbonate replacement (Summerfield, 1982), is not present in voids where calcite occurs as a late-stage fill suggests a break between calcite and silica precipitation. If alternating precipitation during carbonate replacement had occurred, then both mineral species might be expected to be present.

Profile C, a complex silcrete and altered calcrete exposure, represents a different history of development, although there is also evidence of diagenetic changes due to circulating silica- and carbonate-rich groundwaters. With the exception of the uppermost samples the profile is dominantly a massive calcrete, although authigenic nodules are present in some sections, possibly suggesting a groundwater origin (Wright and Tucker, 1991). The majority of duricrusts have an F- to M-fabric, with dispersed quartz grains sometimes showing evidence of dissolution.

The matrix and void fills throughout the profile provide evidence for diagenesis beneath the water table. Sections of the matrix contain patches of sparry calcite up to 6 mm in diameter in amongst microcrystalline calcite, a feature identified in Australian groundwater calcretes as indicating gradual matrix transformation within the phreatic zone (Arakel *et al.*, 1989). Replacement of microcrystalline calcite by disordered chalcedony, cryptocrystalline silica and microquartz is common throughout the profile but particularly in the upper sections, also indicating dissolution and precipitation beneath the water table (Jacobson *et al.*, 1988; Arakel *et al.*, 1989). In upper sections replacement is almost complete, suggesting that water tables remained at, or fluctuated around, this level for a sufficiently long time to allow total dissolution and replacement.

Void fills are complex, ranging between simple calcite void-linings and opal-chalcedony-megaquartz fills. The siliceous void fills represent similar conditions to those described above, indicating increasingly restricted amounts of movement of silica-rich porewaters. Many voids contain well-formed crystals of calcite perpendicular to the void wall, but overlain by chalcedony and quartz. These fills, as with those in the valley head silcrete, indicate variations in porewater chemistry during precipitation. Void fills normally associated with the groundwater fluctuation zone (section 6.1.5*b*) occur overlying matrix material which would appear to have been partially replaced in the phreatic zone. This may indicate a progressive drop in the water table, although it is possible for chalcedony void fills to occur in the phreatic zone (Arakel *et al.*, 1989).

The origin of the silcretes at the valley head of Valley 1 is of most significance to evaluation of hypotheses for *mekgacha* development. The type of host material is uncertain, but the silcretes do not appear to have developed from the replacement of cements within either pre-existing calcrete or bedrock. This is suggested by three factors; firstly, calcite precipitation in quartz sands to form groundwater or pedogenic calcretes is invariably associated with displacement and, to a lesser extent, dissolution of grains (Summerfield, 1982 p.54; Wright and Tucker, 1991 p.8 & 11). The valley head silcretes have a GS-fabric, although isolated areas with an F-fabric occur, and there is very little evidence for grain dissolution from thin-section studies. Indeed, many grains have a fine Fe-stained coating, indicative of a lack of surface dissolution. Secondly, if the silcretes developed by the passive replacement of cements in pre-existing bedrock (in this case Karoo Ecca sandstones), the process required for the preservation of the self-supported grain framework, then it would be likely that some evidence of remnant bedding structures would remain. There is no evidence of any form of structure within the two sample profiles. Thirdly, replacement silcretes have been shown to exhibit a number of features, including inclusions, irregular crystal boundaries, flamboyant extinction and non-competitive growth fabrics (Summerfield, 1978), none of which are present in the silcretes from the head of Valley 1.

These factors suggest that the silcretes did not develop by passive replacement, which implies formation by direct silica precipitation and cementation within the quartz sand framework. What therefore requires explanation is why over 7 m thickness of silcrete occurs in this particular location. Cementation would require an incursion of silica-bearing porewaters focussed on this particular area. This would imply a local hydrostatic gradient to allow groundwater movement towards the area, which would suggest the presence of a basin or depression. It is also possible that forced upwelling of groundwater occurred in this area, which could be due to the rise in pre-Kalahari topography indicated from borehole logs immediately east of the valley head area (see section 6.2.2*d*).

Letlhakeng Valley 2

The origins of the partly silicified calcretes in Valley 2 are comparatively easy to assess. The presence of bivalve shells replaced by $CaCO_3$ is a probable indication of the deposition of the calcrete host material in fresh- or saline-water environments, since there are no records of terrestrial bivalve molluscs in the Kalahari (Brown, 1980). As the age of the calcretes are unknown it is possible that such mollusca existed in the past, but at the present day such terrestrial species are only found in humid tropical forests; such a radical shift in environmental conditions is unlikely to have occurred. The calcified valves are all intact, suggesting sediment deposition under relatively still-water conditions. Furthermore, the presence of intact shell material throughout the 8.7 m of the profile is indicative of the maintainence of such conditions for a lengthy period. The deposition of these sediments would imply some form of basin structure in the form of a pre-existing valley or depression. Unfortunately, the replacement of the shell material by sparry calcite prevents the exact identification of the bivalve species present which would allow more precise assessment of the conditions prior to and during deposition.

The presence of shells, together with the authigenic nodular to massive micromorphology, displacement F-fabric with little evidence of grain dissolution and constant levels of carbonate throughout the profile (away from silicified areas), suggests formation as a groundwater calcrete (Wright and Tucker, 1991). If this is the case, then downcutting in excess of 15 m has occurred since the initial calcrete formation. The calcrete is relatively unaltered except between 5.65 and 5.90 m where silica deposition and replacement has occurred. Many nodules are coated with distinct brown-stained calcite, presumably due to oxidation of iron in solution. The silica species present and sequences of deposition are diagnostic of the former position of the water table (Arakel *et al.*, 1989). The sequences of void-fill at 5.90 m (sample LETV2 B9) are consistent with silicification at or below the water table (within either the groundwater fluctuation or massive phreatic calcrete zones of Arakel *et al.*, ibid), with calcite void-linings followed by banded opalline silica and chalcedony. Immediately below this (sample LETV2 B11), the replacement of micritic calcite matrix material by chalcedonic silica is suggestive of dissolution and precipitation beneath the water table (in the mottled calcrete zone of Arakel *et al.*, ibid).

Silicification is comparatively limited within the profile, both spatially and in the contribution of silica to the overall composition of the calcrete. This would suggest that either water tables have dropped relatively rapidly, thus preventing extensive diagenetic alteration associated with a fluctuating water table (Arakel *et al.*, 1989), or that circulation of silica-bearing porewaters was temporally limited due to a shift in either pH conditions or groundwater chemistry. The first possibility may indicate that a period of "rapid" incision occurred after the development of the groundwater calcretes, with a time of relative stability before further incision. If the valley floor and associated water table had dropped slowly, it would be expected that evidence of silicification would be more widespread within the profile. The second suggestion is supported by the presence of micrite and microsparite calcite at the centre of some voids, suggesting the dominance of calcite precipitation after an earlier period of silicification. However, as noted above, the precipitation of silica is not simply affected by the amount of silica in solution, but also by the ionic ratio of porewaters (Chadwick *et al.*, 1987*a*). In particular, increases in the concentration of Mg^{2+} in solution may trigger localised silica precipitation. Thus, the precipitation of calcite as pore linings elsewhere in the profile may increase the Mg:Ca ratio and cause silica precipitation (Arakel *et al.*, 1989). A shift in groundwater chemistry may later alter conditions, favouring the precipitation of further calcite.

Clearly, the use of one profile down the valley side is not entirely satisfactory for the assessment of the overall development of the valley, particularly since duricrust composition and diagenetic alteration is spatially variable. Given the variability over short vertical distances, the interval between thin-sections may be too coarse to identify all zones of silicification, particularly if these are texturally controlled (Arakel *et al.*, 1989). However, as suggested above, the general lack of alteration and silicification away from a narrow zone within the profile would appear to suggest a limited period of silicification at or near the water table. This period was not simply associated with the deposition of silica as indicated by the often complex carbonate/silica void fill sequences. The zone of silicification is at a comparable height to the silicified layers identified from borehole logs included in Gwosdz and Modisi (1983), strengthening the probable relationship between diagenesis and water table levels.

Letlhakeng Valley 3

The duricrusts of Valley 3 have similar characteristics to those from Valley 2, exhibiting an authigenic massive to nodular (pisolitic in Profile C) appearance and displacement M- to F-fabrics throughout all three profiles. These similarities, combined with the presence of bivalve shell material (up to 1.1 mm long) in the lower sections of Profile B are probable indicators that the calcretes are again groundwater varieties. As with Valley 2, the mollusc valves are all intact with one specimen consisting of two shells still hinged together indicating relatively still water conditions at the time of deposition of the calcrete host material. Further evidence that the calcretes developed as groundwater types is provided by the mottled appearance of the matrix in lower sections of Profile C, caused by patchy transformation of micritic calcite to microsparite which most probably took place in the phreatic zone (Arakel *et al.*, 1989).

Silicification has occurred in varying degrees within the three profiles, both by replacement and the filling of voids. Profile A shows the most extensive alteration in its upper sections; the matrix of sample 9 contains 56% calcite and 36% cryptocrystalline silica. The upper parts of the profile consist of silcretes which still contain nodular structures indicating the replacement of the original calcitic matrix material by micro-quartz and cryptocrystalline silica. Additionally, partial replacement of calcite nodules is evident in Profile C. A noticeable feature of all siliceous layers formed by replacement is the general absence of displacement associated with diagenesis. The sil-calcretes at the top of Profile B retain the massive and laminar characteristics of unaltered calcretes, a feature also recognised in silicified calcretes from Australia by Arakel (1991).

Replacement appears to have taken place prior to the period of circulation of silica-rich porewaters which led to the precipitation of silica void-fills. In most sections, sequences of void-fill were precipitated over replaced matrix material, indicating that infill occurred at a later stage. Replacement appears to have occurred under similar conditions to those suggested for Valley 2. Sequences of void fill are also consistent with silicification beneath the water table, although some grain and nodule coatings may have developed in the vadose zone, possibly during a time of lower water table. The most common sequence comprises layered opalline silica overlying the calcite void wall, followed by increasingly ordered forms of silica towards the void centre. This sequence reflects that identified by Summerfield (1978, 1983*c*) and Thiry and Millot (1987), indicating declining flow rates of porewaters associated with the progressive reduction in porosity caused by void-filling. Another common feature of void fills is the occurrence of calcite and length-slow chalcedony at void centres. This appears to be a later stage of void fill, and would indicate precipitation in association with the replacement of carbonate (Summerfield, 1983*a,c*).

The sample profile followed the east side of Transect 19, which, as described in section 6.2.1*b (i)*, contains a 100 m long outcrop of cal-silcrete. This outcrop coincides with the silicified upper part of the profile. It was earlier suggested that the silicified calcrete formed in conjunction with lateral groundwater movements associated with the presence of the pan presently occupying the valley floor. Thin-section studies would suggest that fluctuations in the water table were of significance in the silicification of the calcrete, but cannot confirm the role of the presence/absence of a pan in this process.

Okwa Valley

The calcretes from Okwa Profiles 2 and 3 require little interpretation regarding their environmental significance. They are groundwater varieties, containing shell material and detrital mineral grains in an authigenic matrix of nodular to pisolitic calcite, representing a cemented terrace. No diagenetic alteration appears to have occurred in either profile.

Okwa Profile 4 is, however, much more complex. It consists of a composite silcrete containing a siliceous duricrust with some matrix replacement by calcite and mixed silica and calcite void fills. The duricrusts all exhibit M- to F-fabrics, with the presence of a partially replaced calcite glaebule within one section suggesting that the silcrete developed by replacement of a pre-existing calcrete. There appears to have been four distinct phases of development within the profile, starting with the formation of an almost isotropic Fe-stained silcrete generated by disordered chalcedony replacing the matrix material of a calcrete. This silcrete, in turn, appears to have been partially replaced by calcite with the growth of euhedral sparry calcite crystals up to 1 mm long. The calcrete emplacement caused extensive brecciation, leading to the development of sub-horizontal jointing. The calcite crystals were subsequently overlain by disordered chalcedony with siliceous void-fills completing the diagenetic sequence.

This sequence of development is consistent with diagenesis beneath the water table identified by Arakel *et al.* (1989), with alternating precipitation of calcite and silica reflecting changes in the groundwater chemistry. The original calcrete replacement would have most probably occurred within the phreatic zone by dissolution and precipitation of disordered chalcedony. Radially arranged calcite crystal growth could also occur within solution voids in relatively stable conditions beneath the water table, and this could brecciate the silcrete. The calcite shows signs of partial replacement by disordered chalcedony, with a final void fill comprising increasingly well-organised species of silica, again indicating increasingly restricted porewater circulation.

Auob Valley

The profile from Kalkheuval in the Auob Valley is also complex, and consists of F- to M-fabric calcretes which contain little clear petrographic evidence for their mode of origin. The majority of the matrix material consists of sparry and nodular calcite, a fabric which would be consistent with long term alteration in the phreatic zone. There is also partial replacement of the sparry calcite by disordered chalcedony. Void fills contain a variety of silica and carbonate forms, but most commonly sparry calcite. Where siliceous void-fills do occur, they include multiple disordered chalcedony to megaquartz sequences, indicating fluctuations in silica precipitation. There are further indications of variations in the silica/carbonate saturation of the porewater from alternating silica/calcite fills, and even calcite merging into chalcedony in one planar void.

General implications of petrographic studies for mekgacha development

The petrographic studies from Letlhakeng have two major implications. Firstly, the silicification of calcrete and silcrete sections suggests development in close association with groundwater levels. This factor, coupled with the inclusion of shell material within the calcretes implies the presence of a pre-existing depression, presumably a proto-valley, prior to formation. The second observation is the extent to which silica-rich circulating porewaters have contributed to the genesis or silicification of duricrusts. The varying degree of replacement and alteration seen at different heights within calcrete profiles can explain the lack of overall correlation between profiles (as discussed in section 6.2.1*b*). Thus the form of duricrust exposed within valley flanks and boreholes is intrinsically linked to the presence of the valley, and would not appear to represent a pre-existing pedogenic duricrust suite which has subsequently been altered. Furthermore, the majority of silicification appears to have taken place beneath the water table, and presumably occurred prior to valley incision and not within the flanks of an incised valley. Studies of silicification in Australian calcretes by Arakel *et al.* (1989) found that silcretes did not occur in laterally continuous layers, with textural controls within the calcrete identified as a major determinant of silcrete position. This may explain the occurrence of silicified layers at different heights within different parts of the valleys. It is, however, possible that silicification in the uppermost parts of profiles in Letlhakeng Valley 1 occurred as a result of dissolution of overlying Kalahari Sand, although petrographic studies of these silcretes have not been undertaken.

Profiles 2 and 3 from the Okwa confirm their apparent origin based on field observations, namely that they are cemented terrace remnants. Profile 4, however, contains evidence of a complex evolution to produce a composite silcrete by initial replacement of a calcrete in conjunction with water table fluctuations. The Auob profile is the most enigmatic, containing little evidence to confirm its origin. Thin-section studies suggest that partial silicification of the calcrete took place beneath the water table, but a definite groundwater origin cannot be asserted. This, combined with the consistency of duricrust profiles along the valley, indicates that the calcretes probably originally developed within a soil profile, but that alteration has occurred in conjunction with downcutting of the valley.

Finally, it should be noted that none of the diagenetic features in silcretes which Summerfield (1983*a*) suggests as indicative of formation in association with deep-weathering profiles (table 6.4) were identified within the study profiles. In particular, there was no evidence of either authigenic silica glaebules (although inherited microcrystalline calcite nodules were incorporated into some samples) or colloform features. Summerfield (1983*c*) suggests that glaebules represent nucleated precipitation of silica at an early stage of silcrete development, eventually grading into continuous silcrete horizons. Silcrete formation within the study profiles does not appear to have proceeded in this manner. Neither did the matrix material develop by the rhythmic precipitation of silica as would be indicated by the presence of colloform structures. The silcretes, in particular, appear to be consistent with Summerfield's "non-weathering profile" types.

However, the climatic inferences which could be made from this comparison are not necessarily correct. Petrographic evidence of the host material composition from Letlhakeng Valleys 2 and 3 indicates deposition under still or gently flowing water, allowing the preservation of intact mollusc valves. This suggests that the calcretes and silcretes in these valleys developed in alluvial sediments, as opposed to formation by replacement of weathered bedrock and clays in the case of Cape Coastal weathering-profile silcretes or the cementation of desert sediments implied by Summerfield (1983*a*). As such, differences must be expected, particularly with regard to geochemistry, which will be investigated in the following section of this chapter.

6.2.4 X-ray fluorescence analyses

(a) Methodology and sample sites

In order to assess duricrust major element bulk chemistry (with analysis by x-ray fluorescence - XRF), samples were collected from thirteen locations. The number of samples analysed from each site is given in table 6.7, together with the exact sampling locations. In total 32 samples were analysed, including 22 silcrete samples, 5 calcrete, 1 sil-calcrete, 1 cal-silcrete, 2 bedrock and 1 ferro-silcrete. Sample locations were selected where well-exposed profiles of duricrusts occurred, allowing detailed vertical sampling. Additionally, seven samples from isolated outcrops were analysed, including two samples of bedrock associated with duricrust exposures which outcropped in Letlhakeng Valley 1 and in the Okwa Valley east of Tswaane borehole. Where both thin-section and XRF analyses were undertaken (as indicated in table 6.7), the samples used for bulk chemical analysis were from the same rock pieces which had been sectioned and have the same sample number. In this way, the results of thin-sectioning and XRF analysis could be compared to validate the technique of point-counting and to assess mineralogy.

Analysis was carried out at the Earth Sciences Unit of the University of Sheffield. The sample was crushed for 20 seconds using an industrial grinding mill to produce a fine powder. 0.75 grams of the powdered rock was mixed with 7.5 grams of lithium metaborate, this mixture then being fused at 1100 °C for 20 minutes. The fused sample was subsequently cast into a Pt/Au mould, and left to cool. The upper surface of the sample was left untouched and it was this surface which was analysed by x-ray fluorescence (Norrish and Chappell, 1967).

(b) Results

(i) Major element bulk chemistry of duricrust samples

The results of XRF analysis are shown in tables 6.8 and 6.9, the former table including only silcrete samples and the latter a mixture of other duricrust and bedrock samples. A total of eleven major elements were analysed, with the total percentage resulting from the analysis and the loss on ignition also included in the tables. Samples with a high carbonate content have very high weight losses during preparation due to the escape of CO_2 during fusion; as such, the weight loss can effectively be regarded as lost carbonate. The following discussion of results is largely qualitative, with quantitative analysis of silcrete chemistry included in the next section.

Calcretes and intermediate duricrusts

Table 6.10 shows the mean chemical composition of the five calcrete and twenty-two silcrete samples. The calcretes have a mean CaO content of 46.42%, slightly higher than the world calcrete mean of 42.62% reported by Goudie (1973*a* p.18), supporting the absence of clay minerals noted from petrographic studies. Samples from Letlhakeng Valley 2 are also compositionally consistent with analyses from the same location by Gwosdz and Modisi (1983).

Sample			Concentration in Weight %											Total	L.O.I.
Location	No.	Type	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	SO_3		
LET V1	A2	S	95.80	0.12	1.07	0.52	0.01	0.35	0.04	<0.10	0.45	0.03	<0.02	98.39	1.61
LET V1	A4	S	93.47	0.11	1.18	0.55	0.01	0.75	0.07	<0.10	0.42	0.02	<0.02	96.58	3.42
LET V1	A6	S	92.25	0.15	1.45	0.77	0.01	0.20	0.24	<0.10	0.61	0.16	0.15	95.99	4.01
LET V1	A10	S	95.11	0.12	1.05	0.58	0.02	0.20	0.03	<0.10	0.43	0.09	<0.02	97.63	2.37
LET V1	A14	S	93.60	0.15	1.17	0.60	0.01	0.10	<0.02	<0.10	0.52	0.09	<0.02	96.24	3.76
LET V1	A18	S	94.68	0.10	0.89	0.51	0.01	0.34	0.10	<0.10	0.42	0.16	<0.02	97.21	2.79
LET V1	A22	S	93.47	0.13	1.01	0.51	0.01	0.71	0.03	<0.10	0.48	0.05	<0.02	96.40	3.60
LET V1	A25	S	93.80	0.11	0.92	0.37	0.01	0.43	0.11	<0.10	0.41	0.05	0.04	96.25	3.75
LET V1	B1	S	93.24	0.16	1.30	0.65	0.03	1.18	0.03	<0.10	0.56	0.02	<0.02	97.17	2.83
LET V1	B3	S	92.00	0.15	1.40	0.68	0.03	1.44	0.41	<0.10	0.53	0.25	<0.02	96.89	3.11
LET V1	B6	S	93.25	0.13	1.19	0.55	0.02	1.24	0.15	<0.10	0.41	0.07	<0.02	97.01	2.99
LET V1	B10	S	93.42	0.13	1.28	0.56	0.03	1.47	0.31	<0.10	0.50	0.21	<0.02	97.91	2.09
LET V1	B14	S	88.87	0.15	2.35	1.00	0.02	3.24	0.20	<0.10	0.56	0.18	<0.02	96.57	3.43
LET V1	B18	S	92.97	0.15	1.39	0.74	0.02	1.55	0.10	<0.10	0.51	0.03	<0.02	97.46	2.54
LET V1	B22	S	92.38	0.12	1.03	0.56	0.02	1.54	0.86	<0.10	0.42	0.03	0.21	97.17	2.83
Moselebe	108	S	97.94	0.22	1.85	1.32	0.10	0.85	0.07	<0.10	0.64	<0.01	<0.02	102.99	-2.99
Lephephe	100	S	96.46	0.06	0.38	0.35	0.02	<0.05	<0.02	<0.10	0.11	<0.01	<0.02	97.38	2.62
Bosutswe	200	S	91.99	0.24	1.74	0.94	0.04	0.89	0.10	<0.10	0.82	<0.01	<0.02	96.76	3.24
Okwa	4A	S	91.05	0.06	1.09	0.63	0.02	0.60	1.15	0.15	0.67	0.18	0.22	95.82	4.18
Okwa	4C	S	87.05	0.07	1.01	0.56	0.01	1.10	3.78	0.13	0.64	0.04	0.08	94.47	5.53
Okwa	4E	S	90.42	0.12	2.46	1.49	0.02	0.62	0.02	0.11	1.58	0.12	0.03	96.99	3.01
Okwa	4G	S	93.17	0.07	1.31	0.97	<0.01	0.30	0.03	<0.10	0.78	0.45	0.04	97.12	2.88

Table 6.8: Major element bulk chemistry of silcrete samples from the valley head of the Gaotlhobogwe (LET V1), outcrops in the Okwa and Moselebe valleys, and silcrete escarpments southeast of Lephephe and at Bosutswe. Key to sample type: S - silcrete. Analysis by x-ray fluorescence.

Sample			Concentration in Weight %											Total	L.O.I.
Location	No.	Type	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	SO_3		
LET V1	13C	F	68.51	0.09	0.73	26.52	0.02	<0.05	<0.02	<0.10	0.05	0.02	<0.02	95.94	4.06
LET V1	106	SC	35.21	0.07	0.58	0.45	0.02	1.94	33.25	<0.10	0.13	<0.01	0.04	71.69	28.31
LET V1	24A	Bs	45.29	0.40	8.82	30.12	0.67	2.08	0.91	1.16	3.45	0.16	<0.02	93.06	6.94
LET V2	B4	C	1.31	<0.01	0.16	0.10	<0.01	0.52	54.10	<0.10	0.03	<0.01	<0.02	56.22	43.78
LET V2	B15	C	14.03	0.02	0.27	0.15	<0.01	0.73	46.03	<0.10	0.05	<0.01	<0.02	61.28	38.72
LET V2	B25	C	22.92	0.06	0.54	0.25	<0.01	0.29	41.55	<0.10	0.10	0.01	0.04	65.76	34.24
LET V3	C20	C	4.44	<0.01	0.10	0.20	<0.01	0.85	52.24	<0.10	0.02	<0.01	0.02	57.87	42.13
Okwa	68	Bg	63.59	0.29	17.33	4.12	0.02	<0.05	0.26	4.15	8.98	0.03	<0.02	98.77	1.23
Auob	115B	C	23.14	0.15	1.95	1.21	0.06	1.67	38.20	0.19	0.50	0.02	<0.02	67.09	32.91
Auob	115H	CS	51.33	0.39	5.47	3.03	0.04	2.66	17.09	0.62	1.49	0.02	0.03	82.17	17.83

Table 6.9: Major element bulk chemistry of samples from Letlhakeng Valleys 1, 2 and 3 (LET V1, V2 and V3), and the Okwa and Auob valleys. Key to sample type: C - calcrete, F - ferro-silcrete, CS - cal-silcrete, SC - sil-calcrete, Bg - granitoid-gneiss, Bs - Karoo (Ecca) sandstone. Analysis by x-ray fluorescence.

Table 6.10: Mean bulk chemical compositions for calcrete and silcrete samples.

	Calcrete (5 samples)			Silcrete (22 samples)		
Component	Mean %	Maximum	Minimum	Mean %	Maximum	Minimum
SiO_2	13.17	1.31	23.14	93.02	87.05	97.94
TiO_2	0.05	<0.01	0.15	0.13	0.06	0.24
Al_2O_3	0.60	0.10	1.95	1.30	0.38	2.46
Fe_2O_3	0.38	0.10	1.21	0.70	0.35	1.49
MnO	0.02	<0.01	0.06	0.02	<0.01	0.10
MgO	0.64	0.29	1.67	0.87	<0.05	3.24
CaO	46.42	38.20	54.10	0.36	<0.02	3.78
Na_2O	0.04	<0.10	0.19	0.10	<0.10	0.15
K_2O	0.14	0.03	0.50	0.57	0.11	1.58
P_2O_5	0.01	<0.01	0.02	0.10	<0.01	0.45
SO_3	0.03	<0.02	0.04	0.05	<0.02	0.22

The second most important component in all samples was SiO_2 with a mean content of 13.17%, followed by Al_2O_3, Fe_2O_3 and MgO in variable but minor quantities. Variations in the silica content of samples reflects the variability of the skeletal quartz component of the calcrete as discussed in the preceding section.

The intermediate duricrust types include a ferro-silcrete containing 26.52% ferric oxide and 68.51% SiO_2 (sample LET V1 13C), a sil-calcrete with a high MgO content (LET V1 106) and a cal-silcrete (AUOB 115H). This latter sample is unusual as it contains a high SiO_2 content, but also contains above average levels of Al_2O_3, Fe_2O_3, MgO and K_2O.

Silcretes

The mean silica content of Kalahari silcrete samples (table 6.10) is slightly lower than the mean figure of 93.92% reported by Summerfield (1978 p.150) from 13 samples. All other components have similar values to those given by Summerfield, usually within ± 5% of the mean. With the exception of the samples from the Moselebe Valley and Lephephe and Bosutswe escarpments, all other silcrete analyses are from samples taken from vertical exposures. Figures 6.16 to 6.18 graphically represent the chemical compositions of the three main sample profiles, from Letlhakeng Valley 1 and the Okwa Valley. The diagrams include only 9 of the 11 major element components, with the percentages for Na_2O and SO_3 being too low in most samples to be accurately measured by the XRF technique. Two main characteristics of the sampled silcretes can be identified from figures 6.16 to 6.18. Firstly, CaO, MgO and P_2O_5, show considerable fluctuations in their contribution to the silcrete bulk chemistry. Secondly, in all three figures, six of the depicted elements (SiO_2, TiO_2, Al_2O_3, Fe_2O_3, MnO and K_2O) remain in almost constant proportions down the profile.

The fluctuations in CaO and MgO content are relatively simple to explain in light of the discussion of thin-section analyses. Carbonates are a feature of many of the thin-sections, occurring primarily as late-stage void fills within silcretes, and were precipitated from water circulating through pore and void networks. As both CaO and MgO are highly soluble and mobile under "normal" pH ranges (4 to 9), it would be expected that they would constitute a high proportion of the total dissolved solids in such circulating water (Loughnan, 1969). It was impossible to identify whether the carbonate at the centre of the void was dolomite or calcite, but the higher MgO level suggests a high Mg calcite. This may be associated with elevated groundwater salinity levels which favour the precipitation of Mg-rich calcite (Mann and Horwitz, 1979; Watts, 1980). The presence/absence and extent of any carbonate infill is largely determined by the porosity, permeability and interconnectivity of voids within the pre-existing silcrete, with the mineral composition of the infill a function of the chemistry of the circulating waters. The varying levels of CaO and MgO down the profile reflect the extent of carbonate infill encountered in any given sample, and indicate precipitation after the silcrete formation. The variability of P_2O_5 is less easy to explain, but higher levels appear to be associated with the presence of relatively unweathered grains of the heavy mineral apatite (e.g. in sample LET V1 B3). As P_2O_5 is only present in trace quantities, it is likely that even the presence of minor quantities of phosphorus-bearing minerals such as apatite would influence the bulk chemistry of the silcrete sample. Furthermore, given the logarithmic scale on figures 6.16 to 6.18, any slight fluctuation in the concentration of a trace component is likely to be artificially magnified on the diagrams.

The consistency in chemical composition down the profile is more difficult to explain, and requires consideration of both the possible host material from which the silcrete has developed and any subsequent diagenesis. With regard to possible host materials, the bulk chemistry of bedrock samples from Letlhakeng Valley 1 and the Okwa Valley are indicated for comparative purposes on figures 6.16 to 6.18. It should be noted that the position of the bedrock samples on these figures does not imply that the samples immediately underlie the silcrete profiles. This is particularly the case in the Letlhakeng profiles where the Karoo Ecca Group sandstone sample analysed by XRF was taken from the nearest bedrock outcrop within the valley, some 10 km from the amphitheatre head. The granitoid-gneiss sample from the Okwa Valley was taken from an outcrop immediately below the base of a debris slope developed at the foot of the silcrete profile.

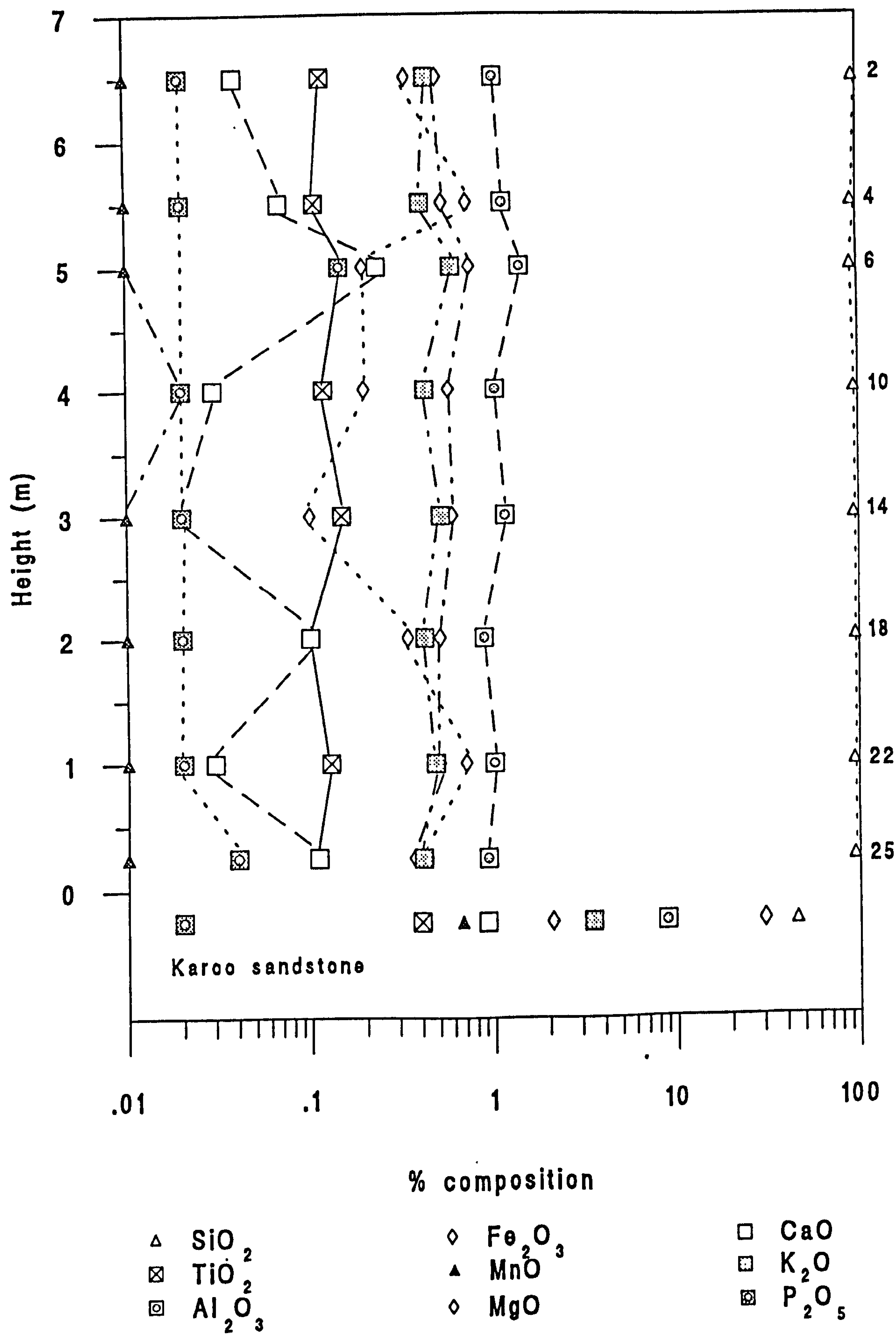


Figure 6.16: Variations in major element bulk chemistry in Letlhakeng Valley 1 Profile A.

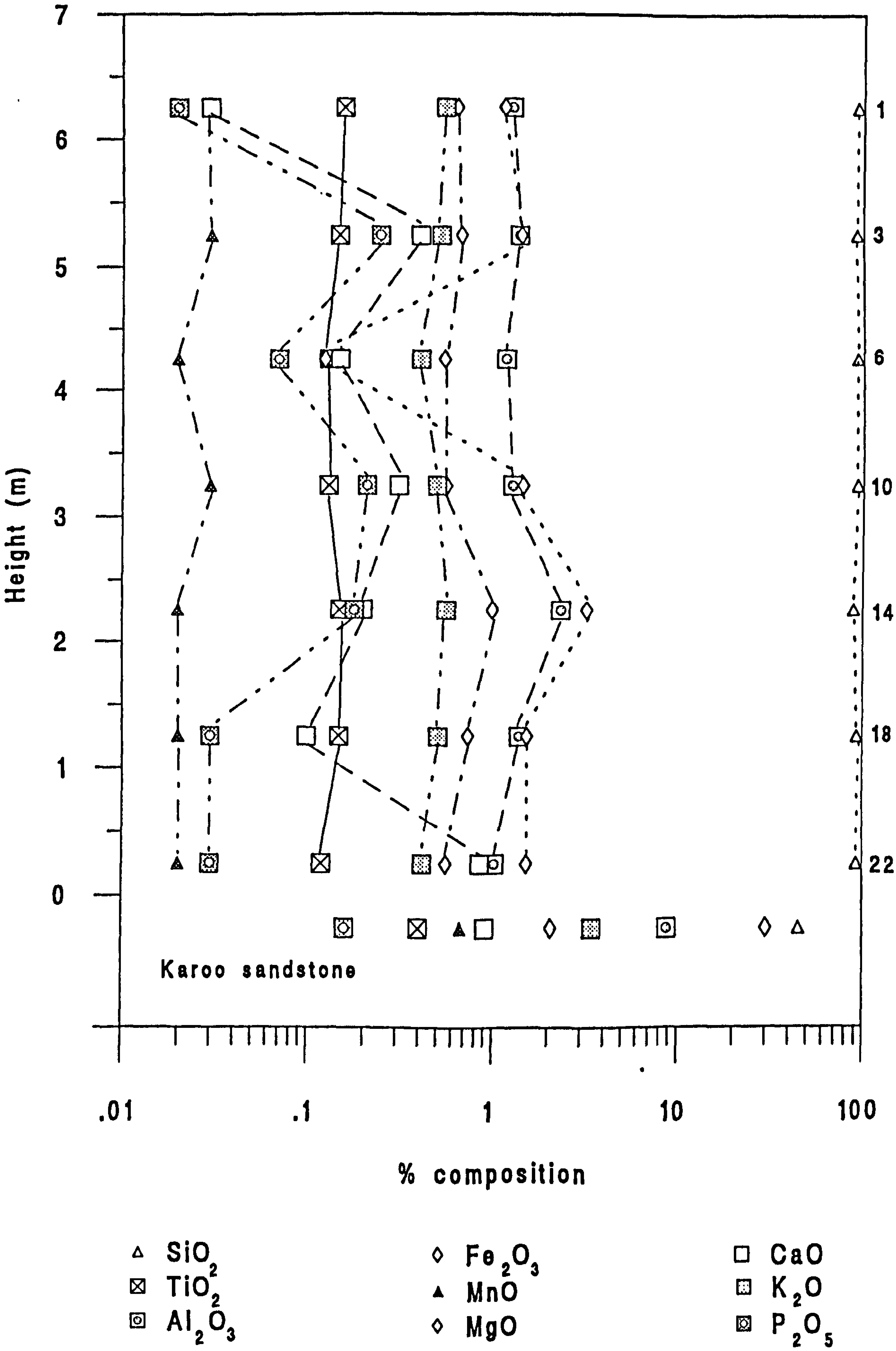


Figure 6.17: Variations in major element bulk chemistry in Letlhakeng Valley 1 Profile B.

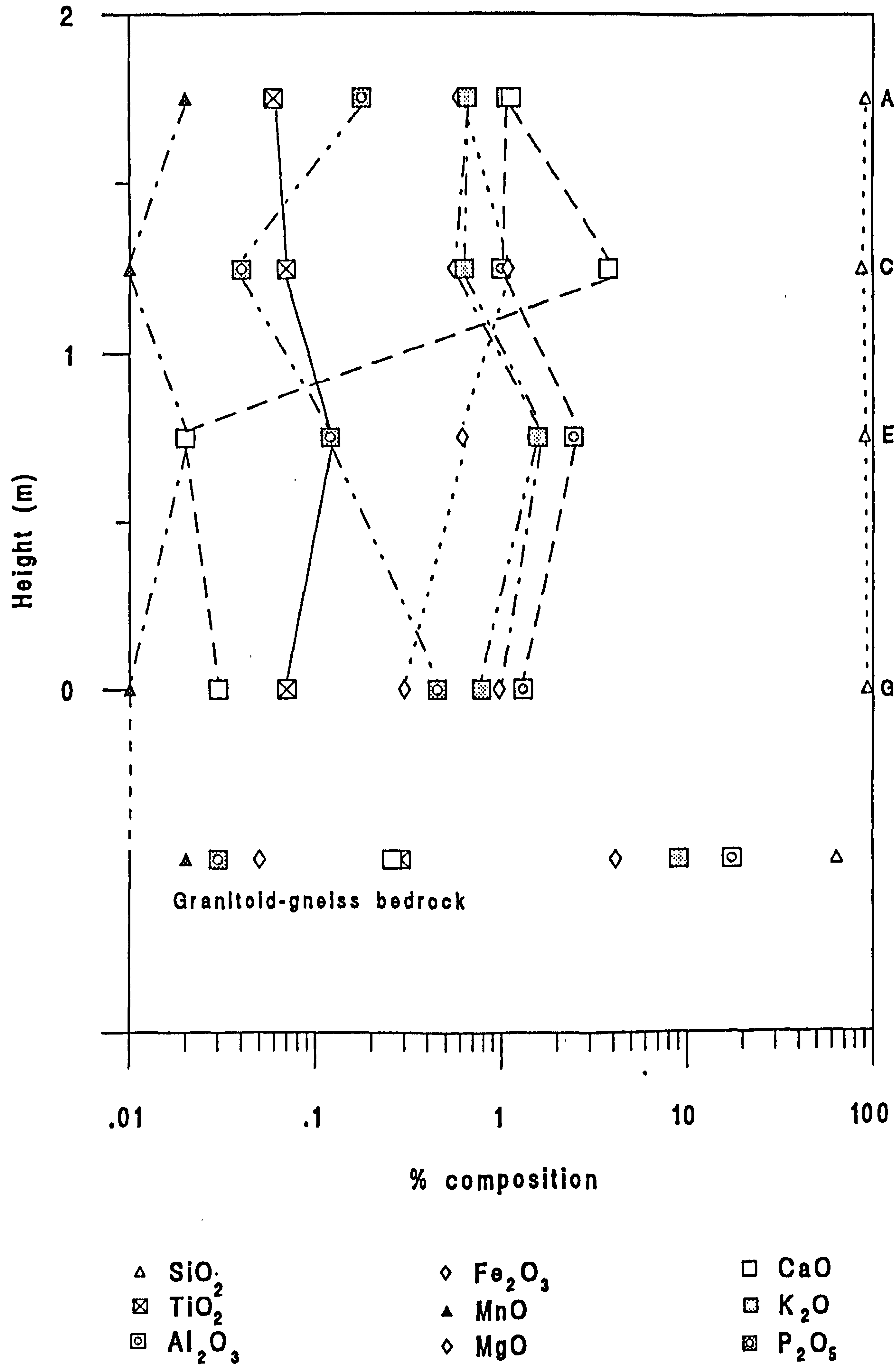


Figure 6.18: Variations in major element bulk chemistry in Okwa Valley Profile 4.

Petrographic evidence from the preceding section suggests that the silcretes from Letlhakeng Valley 1 are most probably derived from silicified Kalahari Group sediments, although Shaw and De Vries (1988) suggest that duricrusts in this area are derived from altered Karoo Ecca Group sandstone. It is instructive to assess the likely environmental conditions if the latter situation were the case. During alteration the original sandstone would need to become enriched in silica (a percentage concentration increase from 45% to over 89%) but depleted in all other chemical components. This replacement would suggest extensive weathering and leaching of the host material, with the breakdown of silicates releasing large quantities of silica. This released silica could then form the cement within the silcrete. Isovolumetric calculations have not been undertaken as part of this study as the sub-horizontal jointing within the silcrete profiles suggests possible expansion during formation.

Such extensive mobilisation of Al_2O_3, TiO_2 and Fe_2O_3 would require significantly more acidic environmental conditions than at present. Al_2O_3 is insoluble in the range pH 4 to 9 (Birkeland, 1974; Loughan, 1969) and would require either highly alkaline or acidic conditions to be mobilised. Decreased pH would appear more likely, since high alkalinity would favour the precipitation of Mg whilst allowing alumina to pass into solution. Titanium is also highly immobile in most surface environments, although mobility does vary dependent upon its valency state. TiO_2 is insoluble at pH values above 2.5, but if titanium is released in the form $Ti(OH)_4$ it is potentially mobile until pH 5 (Loughnan, 1969). This suggests that the pH 3.5 to 4.0 range postulated by Summerfield (1983*d*) for the enrichment of Cape Coastal silcretes in titanium may be more extreme than that actually necessary if Ti were mobilised in the form $Ti(OH)_4$. If Al_2O_3 were also to become mobile, pH would need to be, at least locally, below a value of 4, unless Al_2O_3 were removed by organic complexation (Hesse, 1989). It is unlikely that titanium mobility would increase if released in the form $Ti(OH)_3$ as suggested by Summerfield (1983*d*), as this is even less soluble than aluminium (Loughnan, 1969).

The marked drop in pH required for wholesale mobilisation of normally insoluble elements would be consistent with Summerfield's view of Cape Coastal silcrete development, namely that more humid conditions existed at the time of their formation with an associated increase in acidity due to organic activity (Summerfield, 1983*a*). It is, however, inconsistent with his view of Kalahari silcrete formation under semi-arid conditions, and following petrographic studies of silcretes in thin-section it seems an unlikely scenario for Letlhakeng silcretes. As described above, there is no evidence to suggest extensive replacement and dissolution and none of the diagnostic features of weathering-profile silcretes are present within the study sections. It is more likely that silcretes formed by passive cementation of host materials provided by Kalahari Group sediments. Bulk chemistry of unconsolidated sand is not available, but studies by Baillieul (1975) and Jones (1982) indicate that the surficial Kalahari Sand is predominantly composed of quartz with feldspar in minor quantities. A number of heavy minerals are also present, including garnet, epidote, sillimanite and andalusite, with Thomas (1984*b*) further identifying zircon, rutile, kyanite, apatite, brookite, staurolite, tourmaline and sphene, as well as unidentified opaque minerals. In the samples analysed by Thomas, total heavy mineral percentages ranged from 0.19 to 2.00%, and were dominated by opaques, zircon and rutile. A host material of quartz and feldspar sand plus heavy minerals cemented in a silica matrix could conceivably produce bulk chemistries similar to those presented for the Letlhakeng silcretes without the need for extensive mobilisation. In any event, many of the heavy minerals listed above (e.g. rutile) are unaffected by weathering processes and remain immobile (Loughnan, 1969).

If the silcretes at the valley head of Letlhakeng Valley 1 were developed by passive cementation of Kalahari Group sediments this would explain the down-profile consistency of its major constituents (excluding silica and assuming consistency within the sand). The influence of certain heavy minerals such as apatite upon the percentage composition of trace components such as P_2O_5 has already been commented upon. However, the levels of the more important elements, Al_2O_3, Fe_2O_3 and K_2O, would be less affected by such minor variations in the heavy mineral composition.

In the case of Okwa Profile 4 it is possible that the silcrete has developed as a result of replacement of granitoid-gneiss which immediately underlies the exposure (Don Aldiss, pers. comm. 13th December, 1990), although the contact zone between silcrete and bedrock is obscured by valley alluvium. A more likely situation on the basis of petrographic evidence is that the silcrete developed due to silicification of pre-existing calcrete developed in Kalahari Group sediments. If the silcrete developed by replacement of bedrock there would need to have been overall depletion of TiO_2, Al_2O_3, Fe_2O_3, K_2O, CaO and MgO, and enrichment in SiO_2 and P_2O_5. This would, again, suggest extensive weathering prior to and during silcrete formation, which is not supported by petrographic evidence. Replacement of calcium carbonate by silica in conjunction with fluctuations of the water table is more probable.

The suggestion that Kalahari silcretes developed by the cementation of relatively "sterile" sediments has implications for Summerfield's hypothesis regarding the palaeoclimatic implications of silcrete geochemistry (Summerfield, 1983*a*). If this were the case, then regardless of the extent to which deep-weathering did or did not take place in the Kalahari, there would be little mobilisation of soluble material other than silica and mobile carbonates. Also, any through-flow of water or fluctuation in groundwater level which could provide a lateral or vertical transfer of solutes would have been transmitted through the same barren sediment and would probably be rich only in silica and carbonates. As such, the absence of clay-rich deep-weathering profiles in the Kalahari is not surprising, there being little silicate material except feldspar grains to weather into clay minerals.

There is limited evidence to indicate the circulation of solutes within groundwater stores. Evidence from boreholes in the vicinity of Letlhakeng (section 6.2.2) indicates cave development and decomposition of bedrock at depths of up to 125 m below the surface, with solutional cavities infilled with calcite (Von Hoyer *et al.*, 1985). This may indicate leaching of soluble material from the surface, possibly as a result of groundwater recharge associated with seepage through the floors of *mekgacha*.

(ii) Variations in silcrete bulk chemistry

Qualitative analysis of variations in bulk chemistry

In addition to the assessment of variations within profiles, ternary diagrams of selected major elements from silcrete analyses were compiled to assess variations in bulk chemistry between all samples. These diagrams depict SiO_2 and TiO_2 plus one other component; Al_2O_3 in figure 6.19, Fe_2O_3 in figure 6.20 and MgO in figure 6.21. The variables in figures 6.19 and 6.20 were selected because of their use by Summerfield (1978). The inclusion of MgO in figure 6.21 was made after inspection of the data in table 6.8. The position of a particular sample on a ternary diagram is calculated by totalling the percentages of the three elements for that sample and subsequently expressing each element as a percentage of this total. As a result, the diagrams are misleading, with samples appearing to have much higher percentages of each element than measured in the original XRF analyses. In addition to the data from this study, the ternary diagrams also include samples from the study of Kalahari and Cape Coastal zone silcretes from Summerfield (1978, 1982, 1983*d*).

Considering the samples from this study, there appears to be little distinction between groups of samples on figures 6.19 and 6.20. However, samples from Letlhakeng Valley 1 Profile B have noticeably higher levels of MgO compared to other samples (figure 6.21). This would appear to be due to the higher rate of occurrence of carbonate void-fills evident in thin-section from Profile B, which presumably contain calcite with a high Mg content. The samples from this study plot in similar areas of the ternary diagrams to the Kalahari Group silcretes analysed by Summerfield.

The diagrams also show the same separation of Kalahari Group silcretes from samples from the Cape Coastal zone, primarily on the basis of TiO_2 levels. As discussed in section 6.1.7, Summerfield (1983*a*) has inferred that this regional variation in TiO_2 content is an indicator of palaeoclimatic conditions during periods of silcrete development, with low and high TiO_2 levels suggesting evolution in semi-arid and humid environments respectively.

Whilst there is reasonably convincing evidence that TiO_2 does provide the main difference between Kalahari and Cape samples, the identification of this component was made on a qualitative basis. Inspection of the ternary diagrams shows that two Cape silcrete samples plot very close to the domain of Kalahari samples on figures 6.19 and 6.20. It is possible, although not considered within the scope of this thesis, that selection of other variables for the construction of the ternary diagrams would indicate different relationships between samples. The use of ternary diagrams precludes 6 of the 9 available major elements from analyses, and as such, whilst useful in depicting variations between samples, does not include all the available data.

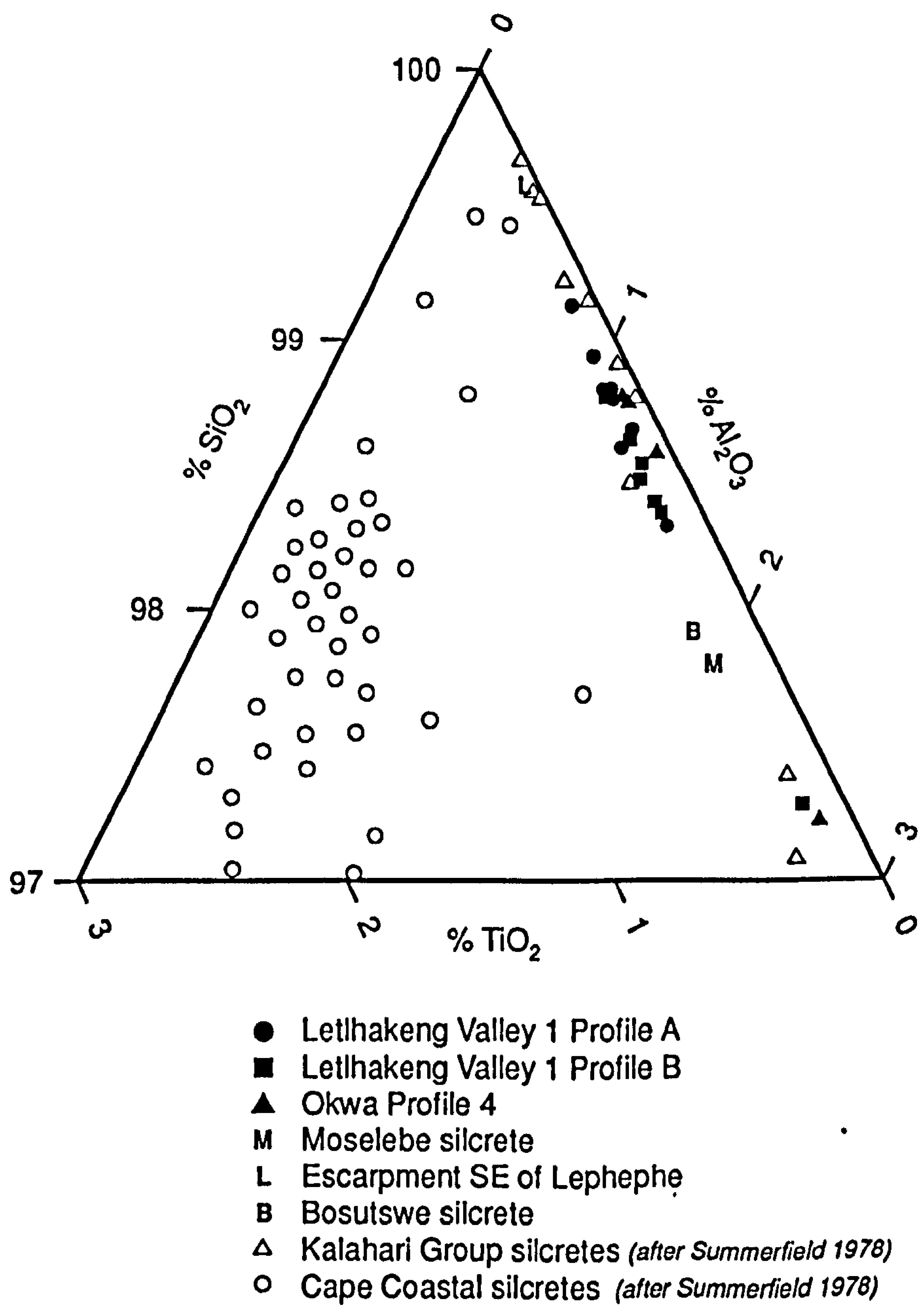


Figure 6.19: Ternary diagram of SiO_2, TiO_2 and Al_2O_3, for Kalahari Group and Cape Coastal silcrete samples.

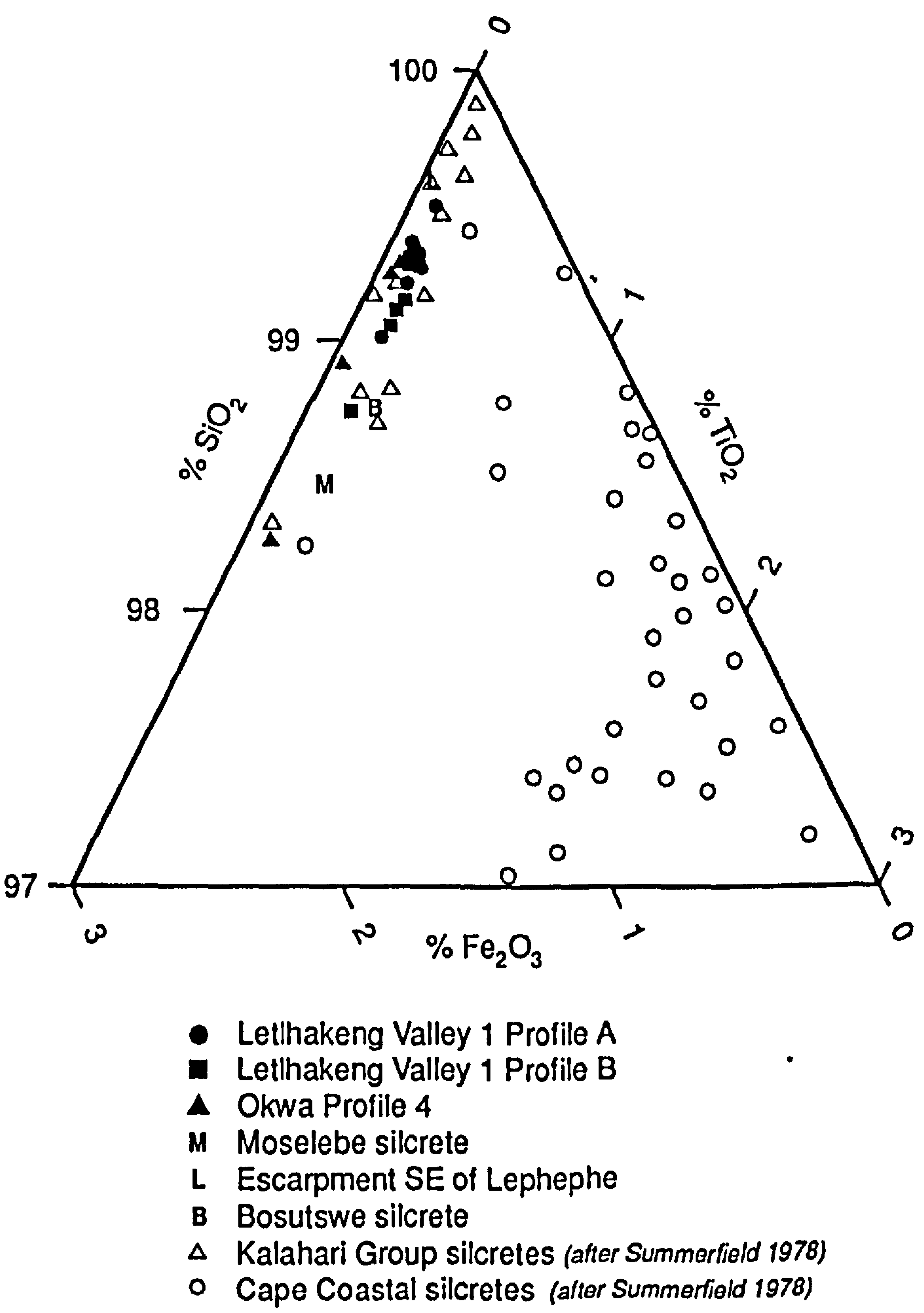


Figure 6.20: Ternary diagram of SiO_2, TiO_2 and Fe_2O_3, for Kalahari Group and Cape Coastal silcrete samples.

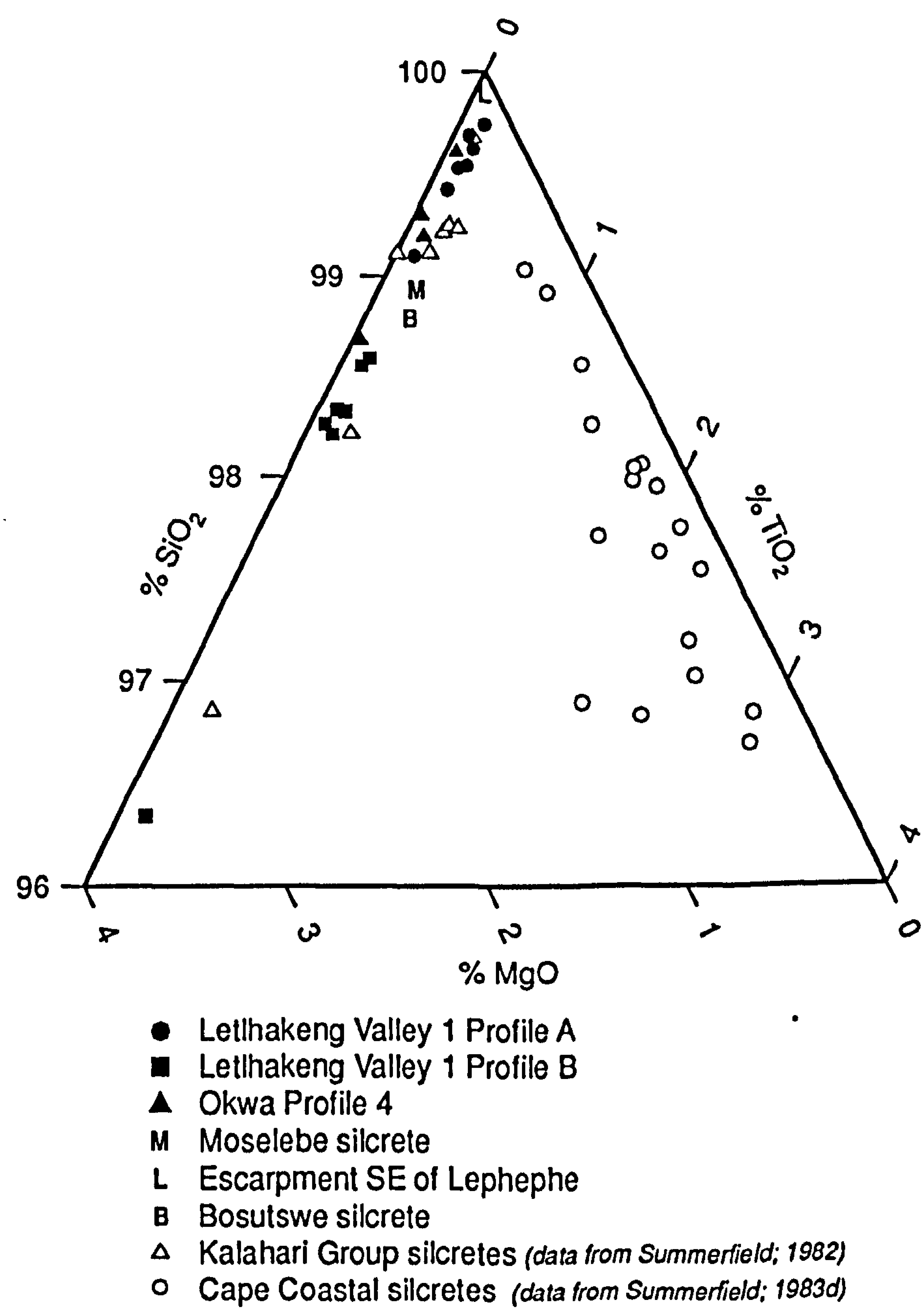


Figure 6.21: Ternary diagram of SiO_2, TiO_2 and MgO, for Kalahari Group and Cape Coastal silcrete samples.

Discriminant analysis of silcrete major element bulk chemistry data

To overcome the problems involved in the purely qualitative identification of chemical differences between silcrete samples from different locations, a statistical approach was used. This involved discriminant analysis of the bulk chemistry data, a powerful statistical method of distinguishing between samples on the basis of different variables. This technique allows all available chemistry data to be taken into account in identifying differences between groups of samples, as opposed to only three variables selected for presentation in a ternary diagram.

Discriminant analysis was used in this study to address three problems. Firstly it was utilised to rigorously identify any differences in bulk chemistry between Kalahari and Cape Coastal silcretes, and particularly to provide quantitative evidence of which element(s) contributed to this difference. In contrast with Summerfield's Kalahari sites, the silcretes from Letlhakeng Valley 1 and Bosutswe have evidence of associated deep-weathering, and yet still exhibit lower TiO_2 levels than Cape Coastal types. As a result, discriminant analysis was subsequently utilised to identify any variations in silcrete chemistry within the Kalahari, primarily to assess chemical variations between silcretes developed in the presence or absence of a weathering profile. Analysis was further extended to assess any variability due to geomorphological associations. The Cape Coastal samples analysed by Summerfield were predominantly from residual surfaces, whilst Kalahari samples were taken mainly from pan or lacustrine environments. Two of the samples from this study were from escarpments at the edges of residual surfaces, whilst the majority were from valley flanks. Thus analyses from a variety of landform associations were included to assess the extent to which differences in silcrete geochemistry could be attributed to geomorphological rather than regional settings.

The remainder of this chapter outlines the technique of discriminant analysis and the application and results of analysis applied to silcrete geochemical data.

The technique of discriminant analysis

The statistical technique of discriminant analysis is discussed in detail by a number of authors (e.g. Johnston, 1978) and, as such, will only be given brief consideration here. The technique is used to distinguish between two or more previously defined population groups (in this case geographically and geomorphologically grouped silcrete samples) on the basis of variations in independent variables (major elements) associated with each group. "Discriminant functions" are statistically derived which maximise the distance between population groups on a linear scale; the number of discriminant functions which may be produced is equal to either (m-1) where m is the number of groups, or v (the number of variables) if there are more groups than variables. In the case of two population groups, only one discriminant function need be produced to separate the groups, with individual cases divided as to whether their discriminant score is higher or lower than the discriminant function. With more than two groups (multiple discriminant analysis) a higher number of functions is often needed to distinguish between groups, although the maximum possible number of functions is not always necessary if particular variables contribute little discriminatory information. In multiple discriminant analysis, the first function derived is the one which produces maximum between-group and minimum within-group variance. The second derived function is orthogonal to the first (Johnston, 1978), with an analysis of variance subsequently performed on the distances of each case from the group mean (the group centroid) in the two-dimensional space formed by the derived functions. Further orthogonal functions are produced until maximum separation of groups in three dimensional space is achieved.

It is possible to assess the success of a discriminant function in distinguishing between groups using statistics including the canonical correlation and Wilks' Lambda, both derived as each function is produced. The former statistic indicates the ratio of total variance estimates along the discriminant function to between-group variance, and is analogous to the squared product moment correlation. Wilks' Lambda indicates the magnitude of the within-groups variance as a proportion of the total, effectively assessing the significance of the discriminating information not accounted for in a function; the larger the value of Lambda, the less successful is the discriminant function at separating groups. Lambda can also be transformed into a chi-squared value enabling statistical testing of the significance of a function. When multiple functions are computed, eigen values associated with each function can be used to measure the relative importance of each function; the eigen value can be expressed as a percentage of the total eigen values for all functions.

Discriminant analysis can be used in two main ways of relevance to this study. Firstly, it is possible to investigate whether a set of variables successfully discriminate between groups of observations defined by hypothetical criteria. This exploratory form of analysis can also identify which of the variables potentially contribute most in distinguishing between groups. The second use of the technique is in the evaluation of the pre-defined classification scheme used to define groups in the original data. The discriminant score for each observation is calculated together with the group centroid. On this basis observations whose scores are closer to another group centroid can be identified.

Applications of discriminant analysis to silcrete bulk chemistry data

Using the SPSS-PC statistical package, discriminant analysis was applied in two stages to combinations of the major element bulk chemistry data for 82 silcrete samples. This data consisted of the bulk chemistry of the 22 silcrete samples shown in table 6.8 and the 60 silcrete samples given in Summerfield (1982, 1983*d*; shown in table 6.11). Data for a total of eleven major elements was available from the analyses in this study, with only nine elements in the analysis of samples by Summerfield. The two stages of analysis grouped samples according to their geographical location (Stage 1) and their geomorphological affinity (Stage 2), with the samples used at each stage shown in table 6.12. At each stage, both uses of discriminant analysis outlined above (testing discriminating variables and evaluating classifications) were employed. Variables were included in the discriminant calculation in a "stepwise" procedure, having been statistically assessed for their contribution to the total difference between groups. Of the various methods available to assess the contribution of individual variables, Rao's Method was selected. This method identifies the variable at each stage of the analysis which causes the largest increase in the parameter "Rao's V", a measure of intergroup distance. Thus, variables are selected for inclusion in the discriminant analysis in a decreasing order of importance based on the extent to which they change the value of Rao's *V*. This effectively identifies the variables which contribute to the greatest difference between groups. The groupings of samples in the two stages (summarised in table 6.12) were as follows;

Stage 1: Grouping by geographical location.

Three separate geographical groupings were identified, at decreasing scales;

a) Sub-continental scale.

Silcrete samples in tables 6.8 and 6.9 were sub-divided according to whether they were from the Cape Coastal region (52 samples), or from the Kalahari Group sediments within Botswana and the Northern Cape (30 samples). Whilst a total of 11 variables are included in table 6.8, only 9 are given by Summerfield, these being SiO_2, Al_2O_3, TiO_2, Fe_2O_3, MgO, CaO, K_2O, MnO and P_2O_5. These 9 variables were entered into the discriminant analysis, the two variables omitted from the analysis (Na_2O and SO_3) only constituting between 0.37% and 0.12% of the sample composition.

b) National scale.

The 22 samples from Botswana within this study (table 6.8) were placed into four groups, with group 1 containing silcretes from Letlhakeng Valley 1 Profile A (8 samples), group 2 the samples from Letlhakeng Valley 1 Profile B (7 samples), group 3 the samples from the Moselebe, Bosutswe and Lephephe (3 samples) and group 4 the Okwa silcretes (4 samples). All 11 major elements included in table 6.8 were used in the analysis.

c) Local scale.

The 15 silcrete samples from Letlhakeng Valley 1 were divided into two groups (Profile A and Profile B samples in table 6.8). These two groups were included in the analysis, with all 11 major elements considered.

Stage 2: Grouping by geomorphological affinity.

Silcrete samples from tables 6.8 and 6.11 were grouped according to their landform association (see table 6.12 for summary). These groups consisted of silcretes from valleys (23 samples), escarpments (4 samples), pans (3 samples) and residual surfaces in the Cape Coastal zone (52 samples). The following nine major variables were used in the discriminant analysis; SiO_2, Al_2O_3, TiO_2 Fe_2O_3, MgO, CaO, K_2O, MnO and P_2O_5.

Table 6.11: Major element bulk chemical analyses (% composition) used in discriminant analysis for silcretes from the Cape Coastal Zone (Sample prefix C-, Summerfield; 1983*d*) and from Kalahari Group sediments in Botswana and Northern Cape Province (prefixes B-, DB- and NC-, Summerfield; 1982).

Sample	SiO_2	Al_2O_3	TiO_2	Fe_2O_3	MgO	CaO	K_2O	MnO	P_2O_5
NC-20	90.09	2.64	0.19	1.50	0.49	0.47	2.04	<0.01	0.11
NC-25	90.70	1.81	0.23	0.94	1.43	0.14	0.51	0.02	0.02
NC-39	87.04	1.44	0.23	0.62	2.61	0.41	0.50	0.02	0.03
DB-1	93.13	0.36	<0.01	0.77	0.82	0.06	0.01	0.70	0.02
B-52	93.00	2.57	0.26	0.96	0.46	0.05	0.49	0.01	0.09
B-51	95.61	0.73	0.12	0.48	0.19	0.06	0.17	0.02	<0.01
B-36	91.96	2.81	0.20	1.01	0.52	0.07	2.33	0.01	0.01
B-37	93.37	2.35	0.18	0.74	0.65	0.04	1.96	0.01	0.01
C-5	96.45	0.36	1.79	0.48	.	0.06	0.02	0.01	0.04
C-18	97.23	0.56	0.70	0.55	0.25	0.02	.	0.01	0.03
C-26	96.16	0.57	1.95	0.61	.	0.01	.	0.01	0.05
C-62	96.83	0.61	1.27	0.17	.	0.02	0.01	0.01	0.02
C-81	95.16	0.70	1.60	1.44	.	0.04	0.02	0.01	0.11
C-90	96.45	0.44	1.72	0.81	.	0.01	.	0.01	0.03
C-123	96.84	0.22	1.75	0.42	0.18	0.01	.	0.01	0.01
C-124	97.14	0.21	1.72	0.30	.	0.01	0.01	0.01	0.02
C-157	97.35	0.23	1.39	0.31	.	0.02	0.01	0.01	0.02
C-163	95.97	0.25	1.71	0.89	0.28	0.14	0.01	0.01	0.01
C-225	98.51	0.17	0.78	0.13	.	0.01	.	.	0.02
C-234	98.03	0.20	0.37	0.46	.	0.05	.	0.01	0.01
C-237	90.86	0.26	3.34	4.96	.	0.01	.	0.03	0.03
C-73	91.74	0.49	1.53	4.48	.	0.02	0.01	.	0.03
C-72	90.86	0.52	1.31	5.56	.	0.03	0.01	0.01	0.05
C-97	91.92	1.18	1.80	2.82	.	0.02	0.01	0.01	0.03
C-98	91.82	0.59	1.78	3.07	0.15	0.01	.	0.01	0.05
C-99	96.17	0.34	1.71	1.12	0.22	0.01	.	0.01	0.01
C-100	96.39	0.27	1.59	1.05	.	0.01	.	.	0.02
C-102	95.52	0.41	1.60	1.34	.	0.04	0.02	0.01	0.06
C-103	96.68	0.24	1.71	0.24	0.57	0.02	.	0.01	0.05
C-104	97.41	0.34	1.31	0.09	.	0.02	0.01	0.01	0.12
C-106	96.86	0.45	1.40	0.51	.	0.03	0.02	0.01	0.07
C-119	96.30	0.30	1.57	1.04	.	0.06	0.01	.	0.01
C-118	91.51	1.03	1.32	3.72	0.31	0.01	.	0.02	0.01
C-117	88.30	1.00	1.65	6.57	.	0.02	0.02	.	0.02
C-116	86.86	0.91	1.85	6.77	0.94	0.01	.	.	0.03
C-115	93.24	0.32	1.96	3.43	.	0.03	0.01	.	0.01
C-114	95.47	0.34	2.26	0.80	0.15	0.02	.	0.01	0.03
C-113	88.98	0.40	2.19	6.04	0.73	0.04	0.03	.	0.03
C-112	91.67	1.00	2.31	2.89	0.34	0.02	0.01	0.01	0.03
C-111	96.54	0.31	2.08	0.37	0.14	0.02	0.01	0.01	.
C-143	95.81	1.04	1.94	0.18	.	0.03	0.01	0.01	0.03
C-141	92.88	1.36	2.81	1.91	.	0.04	0.14	.	0.04
C-140	97.24	0.26	1.19	0.10	.	0.02	0.01	0.01	0.01
C-138	97.50	0.34	1.39	0.14	.	0.02	0.01	0.01	0.01
C-137	96.83	0.48	1.84	0.11	.	0.02	0.01	0.01	0.03
C-136	95.59	0.28	1.59	0.14	.	0.01	0.02	0.01	0.02
C-135	93.48	1.42	2.87	0.29	0.29	0.01	0.23	0.01	0.18
C-176	97.01	0.40	1.76	0.10	.	0.12	.	0.01	0.03
C-174	96.30	0.53	2.41	.	.	0.01	0.01	0.01	0.03
C-173	96.63	0.69	2.54	0.09	0.43	0.03	0.01	0.01	0.03
C-172	95.21	1.09	2.56	0.13	.	0.03	0.01	0.02	0.05
C-184	91.35	1.03	1.93	0.30	0.29	2.07	0.06	.	0.02
C-183	93.40	0.42	2.26	0.07	.	1.74	0.01	.	0.04
C-179	96.64	0.26	2.30	0.06	.	0.03	0.01	0.01	0.04
C-208	94.60	0.69	1.73	0.70	.	0.03	0.01	0.01	0.04
C-207	96.59	1.44	0.85	0.71	0.21	0.02	.	0.01	0.02
C-232	96.41	0.39	2.13	0.46	.	0.02	0.01	0.01	0.06
C-231	94.01	0.57	2.91	1.09	0.31	0.02	.	0.01	0.08
C-230	94.75	0.71	2.86	0.53	0.23	0.02	.	0.01	0.05
C-229	97.46	0.12	1.68	0.22	.	0.02	0.01	.	0.03

Table 6.12: Details of samples used in Stage 1 (analysis by geographical location) and Stage 2 (by geomorphological affinity) of discriminant analysis of silcrete major element bulk chemistry.

Stage	Scale	Groups	No of samples	Samples included in analysis
1 a).	Sub-continent	1. Kalahari Group	30	All samples in table 6.8 plus NC-20 to B-37 in table 6.11
		2. Cape Coastal	52	Samples C-5 to C-229 in table 6.11
1 b).	National	1. Profile A	8	LETV1 A2-A25
		2. Profile B	7	LETV1 B1-B22
		3. Misc.	3	Moselebe 108 Lephephe 100 Bosutswe 200
		4. Okwa	4	OKWA 4A-4G
1 c).	Local	1. Profile A	8	LETV1 A2-A25
		2. Profile B	7	LETV1 B1-B22
2.		1. Valley	23	LETV1 A2-A25 LETV1 B1-B22 Moselebe 108 OKWA 4A-4G NC-25 & NC-39 (Molopo valley) DB-1 (Naledi valley, Jwaneng)
		2. Escarpment	4	Lephephe 100 Bosutswe 200 B-52 & B-51 (east of Orapa)
		3. Pan	3	NC-20 B-36 & B-37
		4. Residual surface	52	C-5 to C-229 (Cape Coastal zone)

Discussion of results of Stage 1 of discriminant analysis of silcrete bulk chemistry, grouping samples by geographical location

a) Sub-continental scale.

The single discriminant function generated by analysis of bulk chemistry data using Rao's method produced a clear separation between silcretes from the Cape Coastal zone and the Kalahari (figure 6.22). Kalahari samples have a mean discriminant function of 3.223 whilst the corresponding value for Cape Coastal samples is -1.859. The histogram of individual discriminant functions in figure 6.22 indicates the degree of differentiation achieved by the analysis, with no overlap between the distributions for the two groups.

The SPSS-PC package produces a standardised canonical discriminant function coefficient for each variable entered into the analysis which indicates the relative contribution of each variable to the function. The coefficients are as follows; Al_2O_3 (0.235), TiO_2 (-0.822), MgO (0.363), K_2O (0.383), MnO (0.257) and P_2O_5 (0.241). SiO_2, Fe_2O_3 and CaO contributed insufficient increases to the value of Rao's *V* during the computation and were not included in the generation of the discriminant function. Of the six variables included in the construction of the function, TiO_2 has by far the highest coefficient value (disregarding the sign) and therefore made the greatest overall contribution. It was also the first variable entered into the stepwise procedure. This supports the qualitative suggestion made by Summerfield that concentrations of TiO_2 are the major distinguishing feature between Cape Coastal and Kalahari silcretes. The negative sign of the coefficients for TiO_2 indicates that levels of titanium were lower for Kalahari Group samples than for those from the Cape, whilst all other constituents were higher.

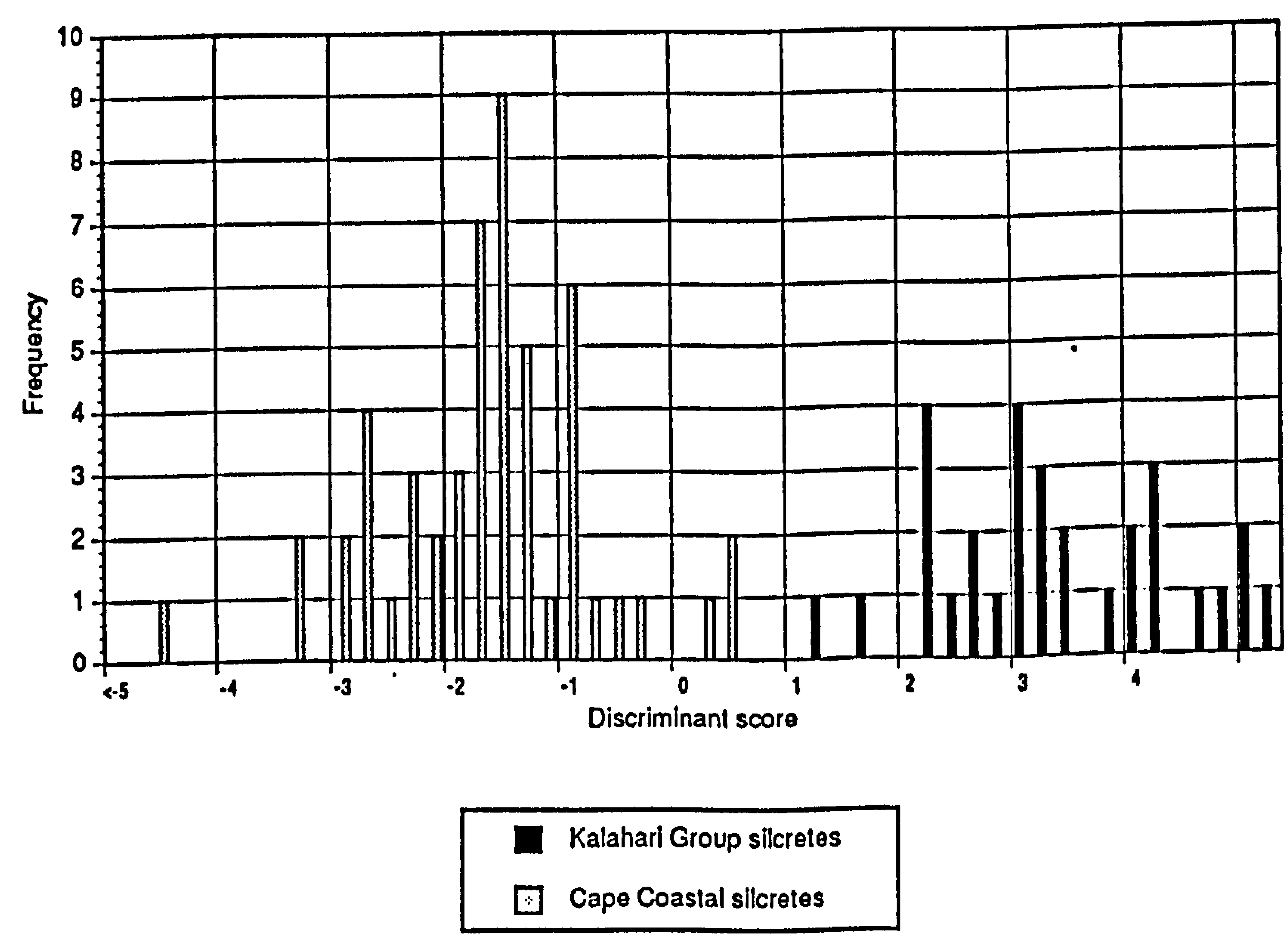


Figure 6.22: Stage 1*a*; histogram of discriminant functions for Kalahari Group and Cape Coastal silcrete samples.

The strength of the group separation produced by the discriminant function is indicated by Wilks' Lambda for the function which has a value of 0.14. This has an associated Chi-squared value of 151.385 with six degrees of freedom, and is significant at more than the 99.9% level. The discriminant function was also highly successful in the classification of individual samples into groups, achieving a 100% success rate.

b) National scale.

The discriminant analysis of four groups of samples from this study produced three discriminant functions. Eleven variables were used in the analysis, of which nine contributed to the calculation of the discriminant functions. The standardised canonical discriminant function coefficients for each of the three functions and nine contributing variables are shown in table 6.13. The major variables contributing to function 1 are Al_2O_3 and Fe_2O_3, with Al_2O_3, MgO and, to a lesser extent, K_2O dominating function 2, and Al_2O_3 and Fe_2O_3 again the major contributory variables to function 3.

The relative importance of each of the three functions can be assessed from the information in table 6.14. The eigen values indicate that function 1 is the most important, accounting for 86.05% of the total variance, with functions 2 and 3 contributing 9.56 and 4.38% respectively. All three functions correlate well with the variables which define group membership, with function 1 having almost perfect correlation.

The position of each sample on the basis of its discriminant score for the first two computed discriminant functions is shown in figure 6.23. From this scatter plot and table 6.14, it is clear that the major group separation is provided by the first function, which splits the samples into two clusters. This suggests that the samples from the Okwa Valley are distinct from the remaining silcretes from Letlhakeng, the Moselebe and escarpment sites. The second function achieves a less distinct separation, but does indicate differences between Letlhakeng Profile B and the remaining samples, primarily on the basis of Al_2O_3 and MgO levels. This plot is slightly misleading as it only contains the first two derived functions. However, as the third function only accounted for 4.38% of the total variance, and is dominated by Al_2O_3 and Fe_2O_3 which were already included in the first function, it is of minor importance.

Clearly, Al_2O_3 is a major variable in all three functions, with Fe_2O_3 also of considerable importance. However, as discussed in the consideration of the results of XRF analysis, both of these constituents are usually immobile under normal pH ranges. As such, the differences between groups of samples are most probably attributable to differences in host material. If the Okwa samples are derived from altered granitoid-gneiss (sample Okwa 68, table 6.8) with a high Al_2O_3 content, this could explain the distinction from the other silcretes. However, as discussed above, this appears unlikely and a more likely explanation is a higher iron content (as indicated from thin-section studies of samples OKWA 4E & 4G) or the presence of more silicate minerals. Petrographic evidence does show slightly higher levels of feldspar within the silcretes, and these may be derived from arkosic bedrock to the west. The importance of MgO to the second discriminant function mainly differentiates between Letlhakeng Valley 1 Profile B and the remaining samples, and will be discussed under Stage 1 c) below.

Table 6.13: Stage 1*b* of discriminant analysis. Standardised canonical discriminant function coefficients indicating contributions of variables to discriminant functions.

Variable	Function 1	Function 2	Function 3
SiO_2	-1.28206	0.61705	-0.17995
Al_2O_3	-4.26075	-2.30946	-3.59319
TiO_2	-1.29340	0.22858	0.68499
Fe_2O_3	4.09406	-0.23162	3.46134
MgO	0.02462	2.10593	0.54619
K_2O	1.53055	1.80135	-0.46059
P_2O_5	0.70987	0.43602	-0.32997
Na_2O	1.60402	-0.07669	0.26770
SO_3	-0.91969	0.10361	-0.26414

Table 6.14: Stage 1*b* of discriminant analysis. Statistics for determining the relative significance of discriminant functions.

Function	Eigen-value	% of variance	Cum. %	Canonical correlation	After Function	Wilks' Lambda	Chi-squared	D. of F.	Signif.
					0	0.0026	86.293	27	0.0000
1	31.6159	86.05	86.05	0.9846	1	0.0849	35.764	16	0.0031
2	3.5136	9.56	95.62	0.8823	2	0.3831	13.911	7	0.0528
3	1.6100	4.38	100.00	0.7854					

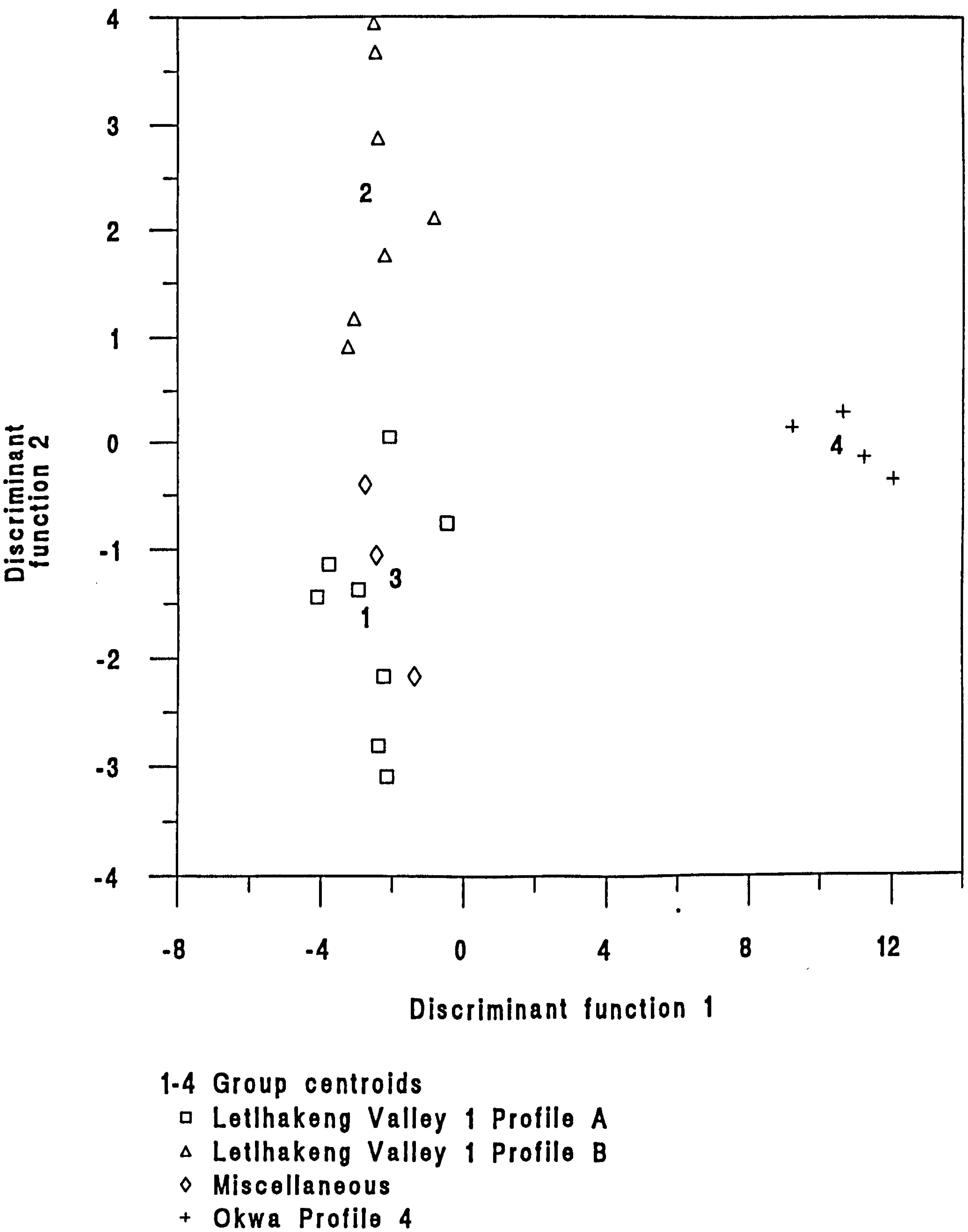


Figure 6.23: Stage 1*b*; scatter plot of silcrete samples on the basis of discriminant functions 1 and 2.

It is interesting to note that on inspection of the raw data for major element bulk chemistry, the differences between groups indicated on figure 6.23 are not immediately clear. This suggests two possibilities; either the maximum inter-group separation produced by discriminant analysis considerably exaggerates real differences, or that discriminant analysis is a powerful technique for identifying subtleties in data sets. However, as the results of classification of samples (table 6.15) indicate, the discriminant functions were 100% successful at predicting group membership.

Table 6.15: Results of classification stage of Stage 1*b* of discriminant analysis

Actual group	No. of cases	Predicted group membership 1	2	3	4
1. Letlhakeng Valley 1 Profile A	8	8 100%	0 0%	0 0%	0 0%
2. Letlhakeng Valley 1 Profile B	7	0 0%	7 100%	0 0%	0 0%
3. Miscellaneous	3	0 0%	0 0%	3 100%	0 0%
4. Okwa Profile 4	4	0 0%	0 0%	0 0%	4 100%

c) Local scale.

The single discriminant function produced a clear separation between silcretes from Profile A and Profile B of the amphitheatre head area of Letlhakeng Valley 1 on the basis of 11 chemical variables (figure 6.24). Profile A samples have a mean discriminant function of -3.570 and Profile B a function of 4.080. The histogram in figure 6.24 indicates the degree of differentiation achieved by the analysis, with no overlap between the distributions for the two groups.

The standardised canonical discriminant function coefficients for each variable entered into the analysis are as follows; Al_2O_3 (-1.244), TiO_2 (1.692), MgO (1.892), K_2O (-1.084), MnO (0.976) and SO_3 (0.547); the variables SiO_2, Fe_2O_3, Na_2O, P_2O_5 and CaO were not included in the generation of the discriminant function. Of the six variables included in the construction of the function, MgO and TiO_2 made the greatest overall contribution, causing the greatest change in Rao's *V*, although MnO and MgO were the first and second variables entered into the stepwise procedure. The importance of MgO in the derivation of the discriminant function supports the observations from the preceding analysis at a "national" level, where MgO was an important variable in the second function also separating Letlhakeng

Valley 1 Profiles A and B. The suggested importance of TiO_2 in distinguishing between the two profiles indicates one of the weaknesses of discriminant analysis, since levels of TiO_2 are only 0.01% to 0.03% higher on average in Profile B than Profile A.

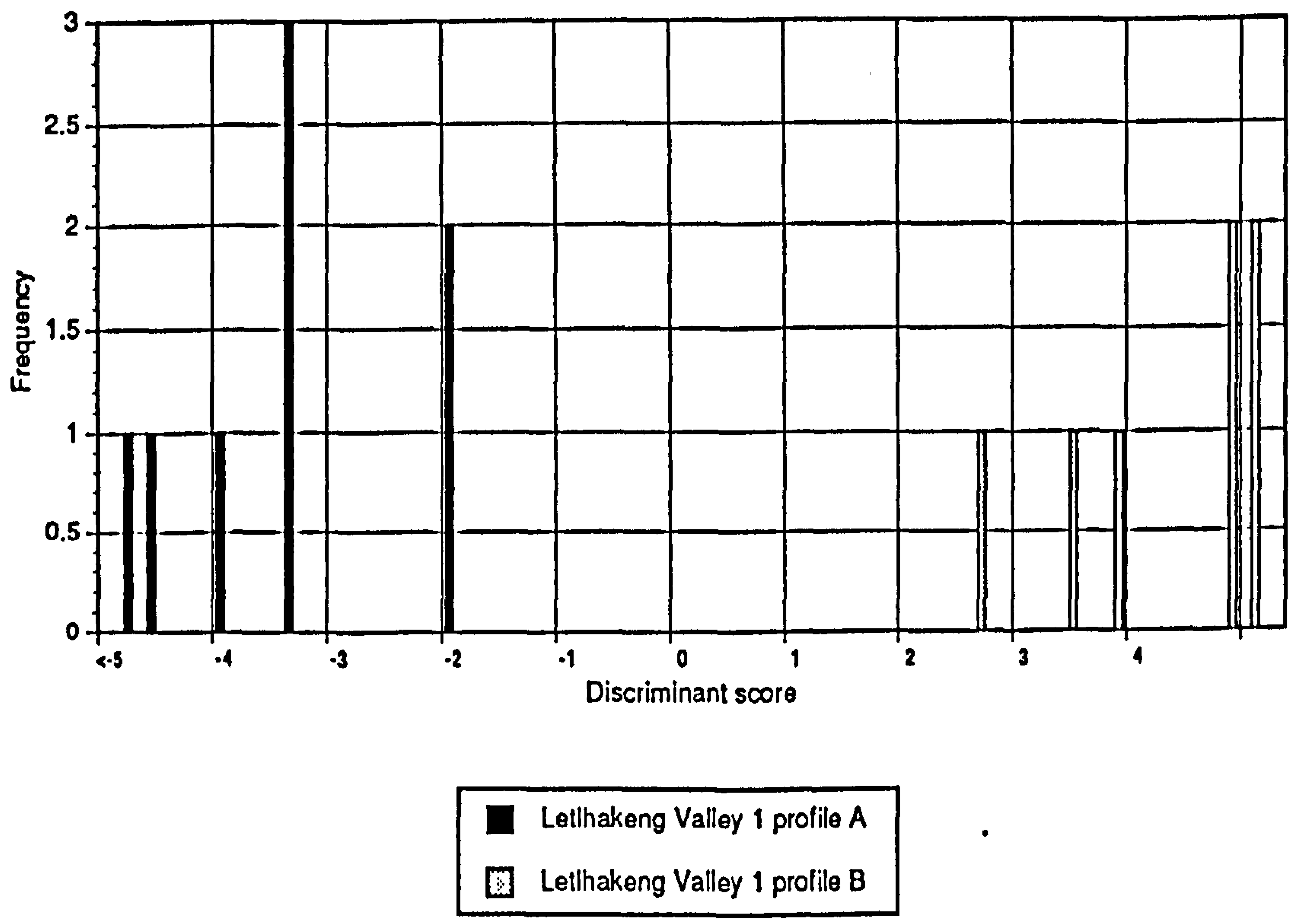


Figure 6.24: Stage 1*c*; histogram of discriminant functions for silcrete samples from Letlhakeng Valley 1 Profiles A and B.

The strength of the group separation produced by the discriminant function is indicated by Wilks' Lambda for the function which has a value of 0.056. This has an associated Chi-squared value of 28.796 with six degrees of freedom, and is significant at the 99.9% level. The discriminant function was also highly successful in the classification of individual samples into groups, achieving a 100% success rate.

Levels of MgO, the main contributor to the differences between Profiles A and B range from 0.10-0.75% and 1.18-3.24% respectively. This difference between profiles can be entirely attributed to the late-stage infilling of voids by carbonates in Profile B evident from thin-section studies. The variable quantity of CaO in both profiles (table 6.8) also suggests the importance of late-stage diagenesis in influencing bulk chemistry. Profiles A and B are situated on opposing sides of the amphitheatre valley head, suggesting that either groundwater chemistries were highly spatially variable or that low porosity in Profile A limited water circulation. The higher MgO levels suggest that the infill is provided by calcite with a high Mg content, although the microcrystalline nature of the void-fill precludes precise mineralogical identification.

Discussion of results of Stage 2 of discriminant analysis of silcrete bulk chemistry, grouping samples by geomorphological affinity

Three discriminant functions were produced from the analysis of four geomorphological groups. Of the nine variables used in the analysis, six contributed to the calculation of the discriminant functions. The standardised canonical discriminant function coefficients for each of the functions and contributing variables are shown in table 6.16. The major variable contributing to function 1 is K_2O, with TiO_2 and K_2O important for function 2 and Al_2O_3 the major contributor to function 3. Inspection of the raw data in tables 6.8 and 6.11 indicates considerably higher K_2O levels in pan silcrete samples which contributes mostly to the first function. In contrast, residual surface samples from the Cape Coastal zone are low in K_2O, whilst valley and escarpment samples can be considered intermediate. Levels of TiO_2 in pan silcretes are also marginally higher than other samples, although it should be noted that this observation is based on a sample of only three.

Summerfield (1982 p.52) notes that the pan silcretes have a green colour and an unusual mineralogy, containing glauconite-illite. He further suggests that the presence of this mineral indicates conditions within the pan prior to silicification of pan sediments; glauconite-illite formation requires supplies of potassium (which would account for the high K_2O content) and iron as well as a favourable oxidation potential. The contribution of TiO_2 to the second function can be largely attributed to the slightly higher levels in pan samples in addition to the differences in titanium content between Cape and Kalahari silcretes already noted in Stage 1 of the discriminant analysis.

Table 6.17 contains information on the relative importance of each of the three functions. The eigen values indicate that function 1 accounts for 85.78% of the total variance, with function 2 contributing 12.92 and function 3 only 1.31%. Functions 1 and 2 correlate well with the variables defining group membership. A scatter plot of discriminant scores from functions 1 and 2 is shown in figure 6.25. The major group separation is provided by the first function, which produces two main clusters of samples. These group the residual surface samples (group 4) and valley and escarpment samples (groups 1 and 2), with pan samples showing little affinity for either cluster. The second function generates a less convincing difference between the two main clusters on the basis of TiO_2 and K_2O levels, with pan samples again separate. The lack of inclusion of a third function is of little importance in this case. The results of classification of samples (table 6.18) indicates the mixed success of the discriminant functions which clearly distinguishes residual surface silcretes from valley and escarpment samples. The functions do not, however, distinguish between escarpment and valley samples. The apparent difference between pan samples and the other 72 analyses is potentially interesting, but as major element bulk chemical information was available for only three samples little further inference can be made.

Table 6.16: Stage 2 of discriminant analysis. Standardised canonical discriminant function coefficients indicating contributions of variables to discriminant functions.

Variable	Function 1	Function 2	Function 3
Al_2O_3	-0.16964	-0.32845	-0.89133
TiO_2	-0.50546	0.68986	0.43493
MgO	0.30783	-0.35697	0.69024
K_2O	1.02236	0.69478	0.38772
MnO	0.38210	-0.04029	0.17079
P_2O_5	0.00843	-0.45805	0.54433

Table 6.17: Stage 2 of discriminant analysis. Statistics for determining the relative significance of discriminant functions.

Function	Eigen- value	% of variance	Cum. %	Canonical correlation	After Function	Wilks' Lambda	Chi- squared	D. of F.	Signif.
					0	0.0206	294.893	18	0.0000
1	12.8240	85.78	85.78	0.9632	1	0.2854	95.286	10	0.0000
2	1.9313	12.92	98.69	0.8117	2	0.8367	13.552	4	0.0089
3	0.1952	1.31	100.00	0.4041					

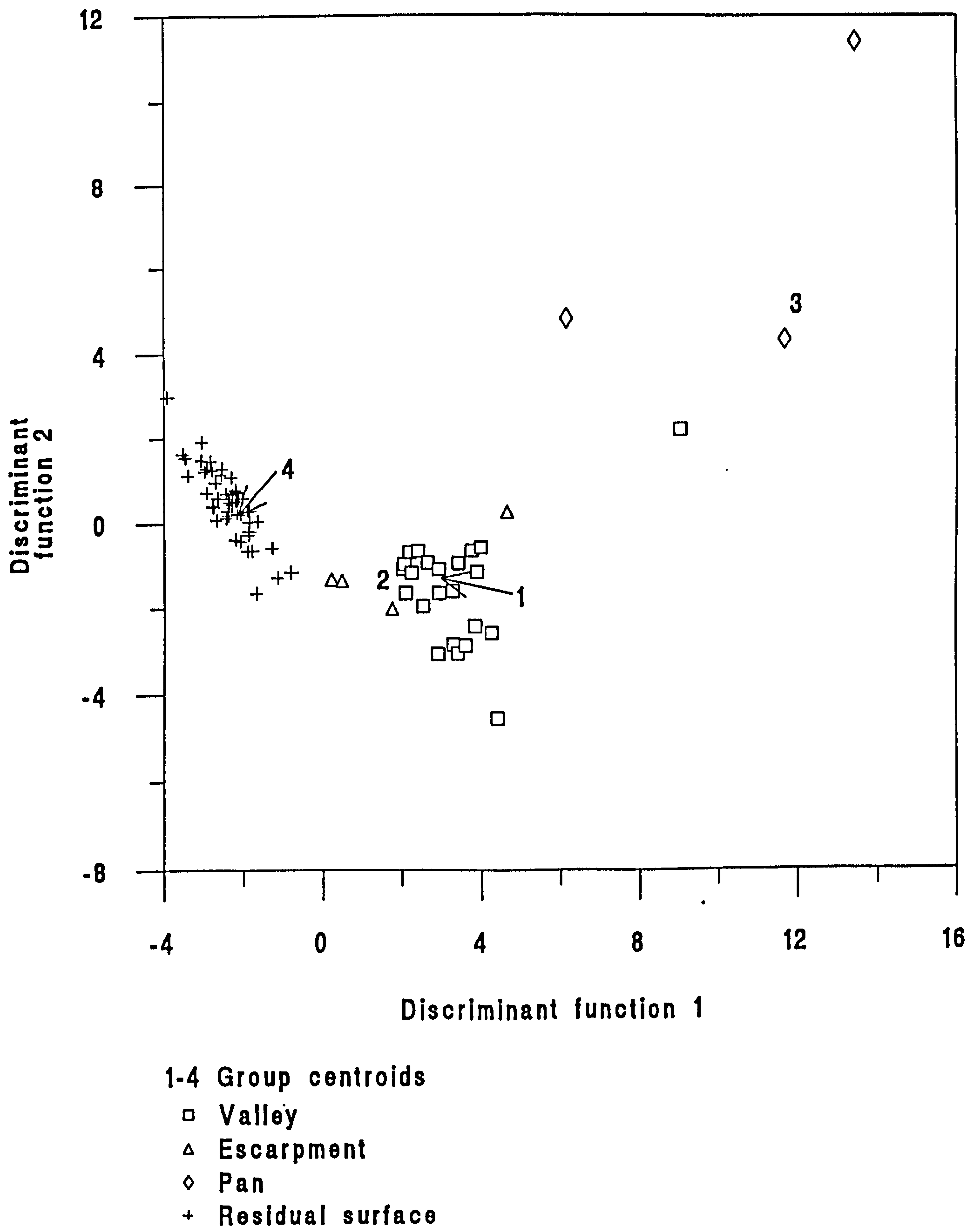


Figure 6.25: Stage 2; scatter plot of silcrete samples on the basis of discriminant functions 1 and 2.

Table 6.18: Results of classification stage of Stage 2 of discriminant analysis

Actual group	No. of cases	Predicted group membership 1	2	3	4
1. Valley	23	16 69.6%	6 26.1%	1 4.3%	0 0%
2. Escarpment	4	1 25.0%	3 75.0%	0 0%	0 0%
3. Pan	3	0 0%	0 0%	3 100%	0 0%
4. Residual surface	52	0 0%	0 0%	0 0%	52 100%

General implications of discriminant analysis

From the results of the discriminant analysis of bulk chemistry data for 82 silcrete samples, three main observations can be made. Firstly, the qualitative observation made by Summerfield (1983*a*) that the major difference between Kalahari and Cape Coastal silcretes is provided by levels of TiO_2 is statistically substantiated. As suggested by Summerfield following a consideration of titanium mobility, this could imply substantially different weathering environments at the time(s) of formation. However, petrographic evidence suggests that the main differences can be largely attributed to differences in host material. The Kalahari silcrete samples from this study and from Summerfield (1982) used in the analysis are comprised of cemented sediment or replaced calcrete which originally developed in cemented Kalahari Group sediments, whilst Cape Coastal samples from Summerfield (1983*d*) were mostly altered bedrock. As such, the climatic inferences made by Summerfield (1983*a*) may well be correct for Cape Coastal silcretes, but the geochemical criteria upon which these inferences are based should not be applied to the Kalahari. The petrographic diagnostic features do, however, appear to be consistently different, and may be a more substantial basis for making palaeoenvironmental assessments of the significance of different silcrete types.

A second factor suggested by the results of discriminant analysis of Kalahari silcretes is that, in addition to variability attributable to differences in host material, mineralogical changes resulting from late-stage diagenesis contribute substantially to differences between samples. This is exemplified by the indication that Letlhakeng Valley 1 Profiles A and B are distinguishable primarily by MgO levels, which are generated by the presence of more extensive carbonate void fills in Profile B. Apart from the extent of void filling, the profiles are otherwise geochemically similar.

The final observation is based upon the results of grouping of samples by geomorphological affinity. The results clearly separated the residual surface samples from Kalahari valley, escarpment and pan samples. However, in the light of the above discussion, it is arguable that the distinction between groups is provided more by geographical location and host material characteristics than by the associated landform. The residual surface samples were all from the Cape Coastal region, and resulted from bedrock replacement. The valley and escarpment samples plot closely together on figure 6.25, but are all formed from cemented sediment or by the replacement of calcrete. The pan samples are derived from cemented pan sediments, a host material of distinctly different chemical and physical characteristics to "barren" Kalahari Sand (primarily due to the presence of glauconite-illite). Thus, the results appear to reflect variations in host material more than landform, although in the case of pan silcretes it could be argued that the difference in host material is a consequence of the alkaline pan environment. In order to further assess variations in silcrete types due to geomorphological affinity, samples associated with different landforms but derived from similar host materials should be analysed.

6.3 Chapter summary

Studies of duricrusts associated with *mekgacha* suggest a number of points of significance to valley development. Firstly, with the exception of the Auob Valley, the formation of duricrusts exposed in Kalahari valleys appears to have been inextricably linked with the presence of some form of valley or depression. This is supported by the results of studies of duricrust profiles, variations exhibited in boreholes and from microscopic analyses.

Studies of variations of duricrusts in boreholes indicates the close association between duricrust development and geomorphology. This is particularly the case for calcretes which appear to have formed due to movements of groundwater beneath valley floors. Petrographic studies also suggest a groundwater origin for most duricrusts, with subsequent diagenesis operating in tandem with fluctuations of the groundwater table. The variability in the morphology of duricrust exposures can be attributed to such diagenesis, which is spatially variable. Significant diagenetic alteration does not appear to have occurred within slope profiles after the incision of valleys, although some silicification of uppermost parts of exposures may be due to dissolution of Kalahari Sand.

The duricrusts exposed in the Auob Valley appear to have developed by a different process, most probably originating due to pedogenesis. This is supported by the consistency within duricrust profile morphology perpendicular to the valley axis and over hundreds of kilometres of the valley length. There is evidence that silicification of profiles may have occurred in conjunction with groundwater movement, but this probably occurred during the course of valley incision.

Finally, geochemical analyses of duricrusts suggest that calcretes and silcretes exposed within *mekgacha* predominantly originated by the cementation of Kalahari Group sediment host materials. Variations in geochemistry between samples can be mostly attributed to variations in the minor components of the original sediment. This can, at least partly, account for differences in bulk chemistry between Kalahari and Cape Coastal silcretes, with a general dearth of minerals other than silica in Kalahari sediments.

Chapter 7

The determination of the relationship between valley orientation and geological structure using network orientation analysis

7.1 Introduction, background and methodology

In evaluating Kalahari *mekgacha* development, a study of the relationships between valley orientation and geological structures was carried out using a technique of network analysis adapted from Abdel-Rahman (1975, 1978) and Abdel-Rahman and Hay (1981). Under the groundwater hypothesis (outlined in chapter 4) a close relationship between geological structures of sub-Kalahari rocks and valley orientation would be expected if either sapping or deep-weathering are dominant processes in valley development. The factors determining this relationship would be expected to vary from network to network, mainly due to spatial variations in geology, but the effects of changes in lithology and features such as faults and joints are of most significance in determining valley orientation.

Fractures, as noted in section 3.3.2, are favoured routes for groundwater flow, especially in non-porous lithologies, because of their secondary porosity (Coates, 1990). Even in porous rock-types such fractures are likely to act as preferential flow paths. Indeed, it is a standard technique in groundwater exploration to locate fracture zones and the intersection of such zones using geophysical methods, as these prove to be most likely sites for groundwater strikes (Peart, 1979; Farr *et al.*, 1981; Coates, 1990; Teme and Oni, 1991; Zeil *et al.*, 1991). Where landforms such as aligned depressions or valleys coincide with lineaments, the chances of groundwater strike are very high (Siddiqui and Parizek, 1971).

Buckley (1984), Buckley and Zeil (1984), Dietvorst *et al.* (1991) and Zeil *et al.* (1991) have substantiated these ideas from studies in eastern Botswana which show that large scale fractures in non-porous pre-Kalahari bedrock act as important aquifers and flow paths for groundwater circulation. Buckley and Zeil (1984) also note that "parasitic fracture zones" associated with doleritic intrusions which extend across much of the northern and middle Kalahari of Botswana (see section 3.2) allow groundwater transmission. The type of fracture most significant in terms of groundwater transmission is uncertain, although results from VIAK Ab (1983) suggest that fractures of tensional origin may be more important than major shear zones. Northwest-southeast trending tensional fractures in the Letlhakeng area of Botswana have been identified as major groundwater aquifers (Buckley, 1984). Dietvorst *et al.* (1991) suggest that present-day regional aquifers in the southeast of Botswana may not be exclusively controlled by fracture systems, with regional tectonic undulations of greater importance in forming groundwater basins. However, despite this alternative suggestion, they do note the role of fractures in locally enhancing permeability and further suggest that aquifer development in synformal structures may be by preferential weathering of fractured bedrock.

The importance of subsurface flows in terms of subterranean weathering of bedrock can be seen from the evidence for deep-weathering recorded in many borehole logs (e.g. in Karoo samples by Von Hoyer *et* *al.*, 1985). Local widening of fissures by groundwater movement has been identified by Buckley (1984 p.29). Such deep flow of groundwater is considered by Shaw and De Vries (1988) to be of importance in terms of valley development by groundwater sapping processes. Fracture zones are also areas of groundwater recharge and subject to preferential weathering, with fractured aquifers in parts of the hardveld in eastern Botswana characterised by surface depressions (Gieske and Selaolo, 1988).

In this section a brief history of network analysis is presented (section 7.1.1), together with a consideration of the relationship between network structure and geological controls. The method used for the analysis of Kalahari valley networks in relation to geological structure is considered in sections 7.1.2 and 7.1.3.

7.1.1 Background to network orientation analysis

The use of statistical methods for the analysis of drainage network composition has a long-standing history as a technique in fluvial geomorphology. The basic qualitative concepts of classifying networks according to their pattern and geometric properties were first suggested in 1899 by W.M. Davis. Since then, the work of Horton (1945) and Strahler (1952) has pioneered the quantification of network geometric properties, with probabilistic approaches to network analysis introduced by Shreve (1966) and Smart (1968).

Despite the considerable interest generated by these studies, very little quantitative research has been carried out on the influence of geology upon network structure. This is despite the long-standing recognition of geological structure as a potential determinant of drainage network orientation. Geological structure is taken here to include lithological variations, folding, faulting, bedding and jointing in rocks, all of which have been considered important in terms of network development, particularly in determining orientation of stream sections (e.g. Bannister, 1980; Laity *et al.* 1980; Pieri *et al.*, 1980). Earlier approaches provided only qualitative assessments of the role of geology on network development. For example, trellis drainage patterns were traditionally considered a result of differential erosion of underlying dipping hard and soft rocks (Knighton, 1984).

Conflicting ideas exist concerning the influence of geology upon drainage networks, and can be illustrated by the work of Mock (1976) and Abrahams and Flint (1983). Testing Shreve's (1966) probabilistic Random Topology Model on ten trellis networks in central Pennsylvania, Mock (1976) concludes that geology exerts little effect upon drainage network structure. Conversely, Abrahams and Flint (1983), testing the same networks, find that the plunging geological syncline underlying the area causes longer stream sections flowing in the downplunge direction.

Whilst many authors have commented on the potential influence of deeper structure on watercourse and valley development (e.g. Egyed, 1957; Ciccacci *et al.*, 1987; McFarlane, 1989) the only attempt to assess the effect of different types of geological structure upon drainage networks was carried out by Abdel-Rahman (1975) based upon studies in the Mula region of southeast Spain. In this study, structures at a variety of scales were taken into account, using data from analysis of aerial photography as well as from geological field mapping. Bannister (1980) also suggests a method for assessing relationships between unit stream lengths and structural data. The methods used by Abdel-Rahman (1975 and 1978) and Bannister (1980) are discussed in section 7.1.3.

7.1.2 The use of network orientation analysis in evaluating modes of Kalahari *mekgacha* development

Under the groundwater hypothesis of valley development discussed in chapter 4, the relationship between valley orientation and geological structure is highly important, particularly in terms of the role of fractures and joints in the transmission of groundwater (Buckley, 1984; Buckley and Zeil, 1984). The significance of structural control can be seen from various investigations beyond the Kalahari (see section 2.2). Studies in the Glen Canyon region of Colorado (Laity *et al.*, 1980; Laity, 1983; Laity and Malin, 1985; Howard *et al.*, 1988) identify certain morphometric properties considered indicative of the influence of groundwater sapping in network development, many of which relate to geological structure. McFarlane (1989) notes the role of fractures as focii for intensive *in situ* deep-weathering in the development of dambos.

At the sub-regional scale, Howard *et al.* (1988) note elongate networks and "pervasive parallelism" of tributaries over a wide spatial extent where structural control is most evident, with asymmetry of tributaries often present. These authors also regard the influence of sedimentological facies changes as important determinants of groundwater movement and sites of emergence. Pieri *et al.* (1980) note that trellis patterns best develop in drainage networks where main valley elongation is perpendicular to the major structural trends. At a more local scale, Laity *et al.* (1980) suggest the importance of joints acting as lines of weakness, for both groundwater and surface flows. Of additional importance for Kalahari valley development is the recognition by Bannister (1980) that structure is likely to have most influence where relative relief and hydrostatic gradients are low.

As already described in section 3.2, the geology of the Kalahari Basin can be broadly viewed as consisting of a veneer of Kalahari Group sediments unconformably overlying older Karoo and Precambrian Basement rocks. In some locations inliers or "islands" of Basement and Karoo rocks outcrop at the surface, but most outcrops and subcrops tend to occur at the fringes of the Kalahari. Any structures in these peripheral areas mainly affect headwater regions of *mekgacha* networks. Thus most structural elements in sub-Kalahari rocks are buried beneath considerable thicknesses of Kalahari Group sediments. With a few exceptions, the Kalahari Group sediments are not displaced by faults or cut by intruded dykes, except where relatively recent earth movements have occurred (e.g. in the vicinity of the Okavango Delta). Where faults and fractures are visible on the surface by means of remote-sensing techniques they generally tend to appear as only faint lineations picked out by the differential growth of vegetation, again with the exception of areas which have experienced neotectonic activity (Aldiss *et al.*, 1989). The fact that fractures which have not been recently reactivated are visible at the surface has been attributed to groundwater circulation in sediments overlying buried fractures, with vegational growth encouraged above the fault zone (Mallick *et al.*, 1981).

For evaluation of the hypotheses outlined in chapter 4, any evidence of structural control within networks may be regarded as significant. However, the type and depth of the controlling geological structure is of most importance in distinguishing the roles of surface- and ground-water in valley development. Thus the depth of bedrock beneath Kalahari Group sediments (where present) is of great significance in the interpretation of any evidence for structural control. The method used for assessing the Kalahari Group thickness is described in section 7.1.4.

As already noted, evidence of close parallelism or perpendicularity of valley sections and geological structure in areas with a covering of Kalahari sediments is likely to indicate the influence of groundwater in valley location. Where bedrock outcrops at the surface the interpretation of any relationship is more complex; geological structure may have influenced the direction of any past surface water flows, but could equally have contributed to groundwater emergence at a site and thus enabled sapping processes to dominate in valley formation or it may have provided a focus for enhanced deep-weathering.

7.1.3 Method of network orientation analysis

The method used to analyse relationships between geological structure and valley orientation is adopted from Abdel-Rahman (1978) and Abdel-Rahman and Hay (1981). In essence, the method uses randomly located sampling circles (as opposed to more commonly used quadrats) which are placed at random locations along a valley course. At each point, the angular difference between valley segments and lineaments (from analysis of Landsat images and other data sources) is recorded for each sampling circle in 10° classes. The pattern of these angular differences is statistically analysed by comparison to the Poisson distribution to assess any deviations from randomness.

(a) Rationale behind network orientation analysis

A variety of methods for the analysis of geomorphological and geological orientation data are available, but few are without limitations. These are primarily due to the directional nature of structures such as joints and faults, which necessitate the use of statistical techniques suitable for the analysis of orientational data. Such directional data may contain a value of dip, and thus need to be plotted over 360°, although plotting over 180° is sufficient where no dip component is included. The major constraint on the adaption of statistical tests for analysing orientational data is the choice of the zero origin, as Abdel-Rahman and Hay (1981) demonstrate. Many conventional statistical tests prove to be highly sensitive to the choice of origin for a circular plot. In traditional analyses, the 0° origin is usually taken as being in a northerly or easterly direction. Should the variable being studied have a distribution in which an orientational peak occurs in the direction of the chosen origin, then this peak is likely to be split and thus appear to be less significant. Handling data with a number of orientational peaks can also present difficulties. If a data-set is unimodal and can be shown to be approximately normally distributed, then parametric statistical tests can be applied; however, multimodality is much more commonplace.

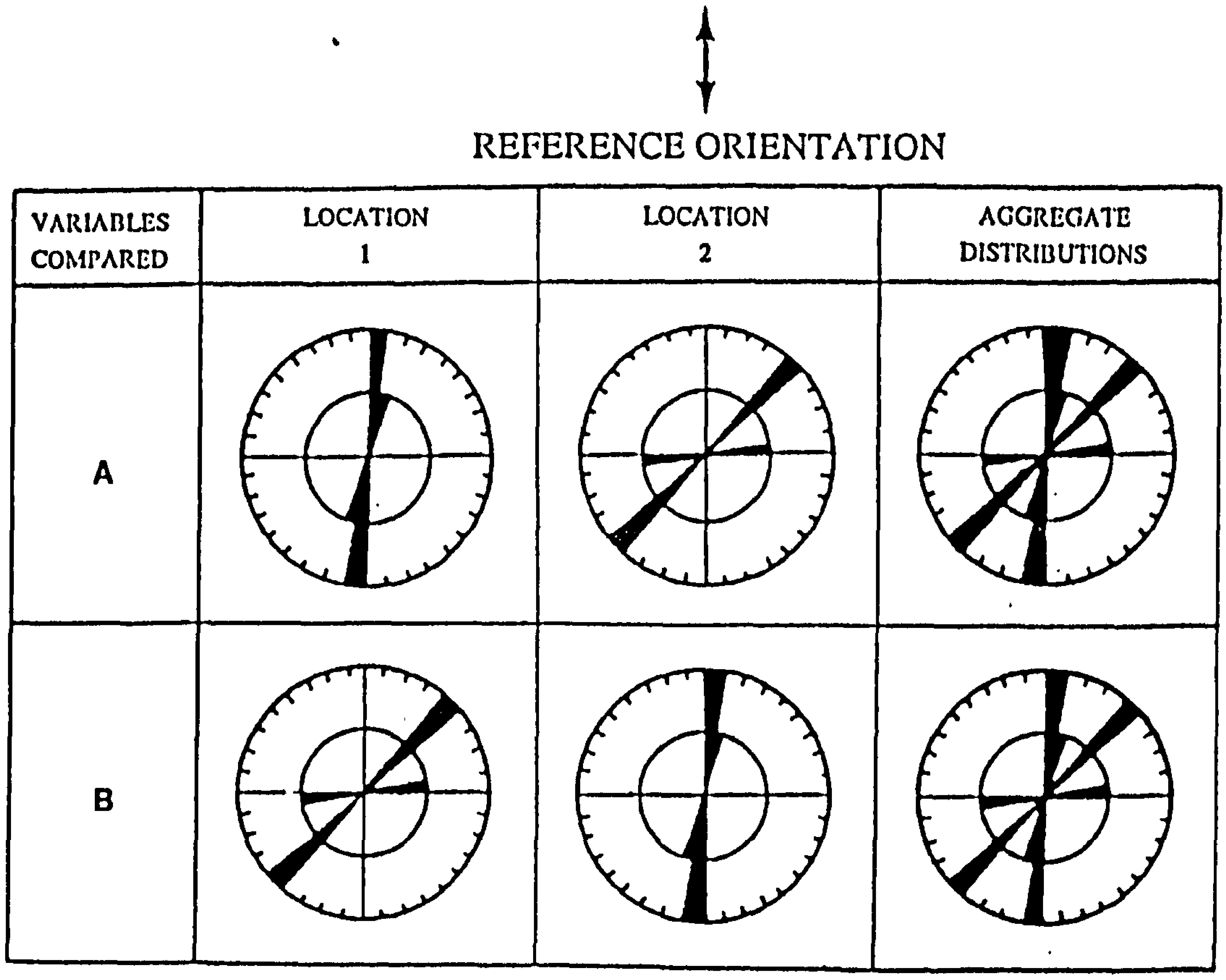


Figure 7.1: Orientational similarity between two variables based on their aggregate distributions. Note that the similarity is spurious since the variables are not orientationally similar in any one sampling location.

Abdel-Rahman and Hay (1981) discuss other difficulties associated with geological orientational data. A basic problem concerns the accuracy of data collection, since errors in the recording of angular observations are common, and data from secondary sources cannot be assumed to be totally accurate. The use of valley networks either taken from secondary sources such as conventional maps or obtained first-hand from remotely-sensed imagery is also prone to errors, especially in establishing the headward extent of networks (Ovenden and Gregory, 1980; Burt and Gardiner, 1982; Knighton, 1984; Burt and Oldman, 1986). As such, statistical techniques which require values with minimal measurement error should not be used with such data; non-parametric tests are therefore preferable.

A further problem is associated with the tendency for fracture lineaments to be extrapolated beyond locations visible on remotely-sensed images. This is especially prevalent where fracture traces are continued across "gaps" between similar-trending fractures, even where the evidence for such continuation is inconclusive. In the event of such extrapolation occurring, any statistical technique taking into account either the length of a lineament or the number of lineaments in a given area will be obsolete. This is a major flaw in the method suggested by Bannister (1980) for comparison of orientational characteristics of unit stream lengths and geological structure. Under this method the stream course is divided into unit lengths which are then classified into orientational classes. This results in a distribution of total unit lengths within each directional category; categories containing a high proportion of the total channel length represent a preferential directional tendency within the network. Structures are classified in a similar way, which may lead to errors if fracture traces or lineaments have been extrapolated across gaps. The distributions of unit stream and structural lengths are then ranked and compared using a Spearman's Rank Correlation. However, unless first-hand data sources are used, this method could well produce an erroneous or misleading result. The more sophisticated analysis of drainage patterns and structurals trend by Ciccacci *et al.* (1987) is based on a similar principle of comparing the azimuth directions of structures and stream segments of different network order, and is as such, similarly flawed.

Another problem with such an approach is noted by Abdel-Rahman (1975). The author argues that the examination of orientational differences between two variables by means of comparing their orientational distributions can lead to spurious results because of problems due to aggregation of information. This is perhaps best demonstrated by figure 7.1 (from Abdel-Rahman and Hay, 1981). In this example, two variables (e.g. fault and valley orientation) are compared by means of their aggregate distributions from measurements made in two different areas. As can be seen in the figure, although there is no correlation between the two variables in the two different regions, the aggregated results suggest similar distributions. The figure illustrates the dangers of attempting analyses of variables such as geological structure over large spatial extents.

(b) Methodology

The method adapted from Abdel-Rahman (1975 and 1978) for network orientation analysis overcomes the problems outlined above by using a random sampling-circle technique rather than taking every structure and valley section within a sampling area into account. The method also measures the angular difference between structures and network sections rather than recording a separate distribution for each and subsequently analysing the two variables.

Analysis took all major endoreic valley systems of the Kalahari north of the Molopo Valley into account, plus the exoreic Moselebe system, with analysis extending as far headward as the hardveld/sandveld boundary at the periphery of the Kalahari. Lack of detailed structural information prevents detailed analyses of the Auob, Nossop and Letlhakane valleys. In order to avoid problems outlined above due to sampling over a very wide area, the analysis was arbitrarily subdivided into quarter degree squares (QDS) of latitude and longitude. Separate analyses were undertaken within each QDS containing dry valley networks, although where valleys were only present in a small part of a QDS the analysis of these sections of the network were combined with an appropriate adjacent QDS (as shown in figure 7.2). The critical length of valley within a QDS was taken as 50 km, below which the square was amalgamated with an adjacent QDS. Analyses were carried out using base maps at 1:250,000, the only scale considered appropriate given the available geological data (see below). Analysis thus took valley orientation into account rather than the larger scale analyses of stream orientation undertaken by Abdel-Rahman (1975 and 1978). The exact method used is as follows;

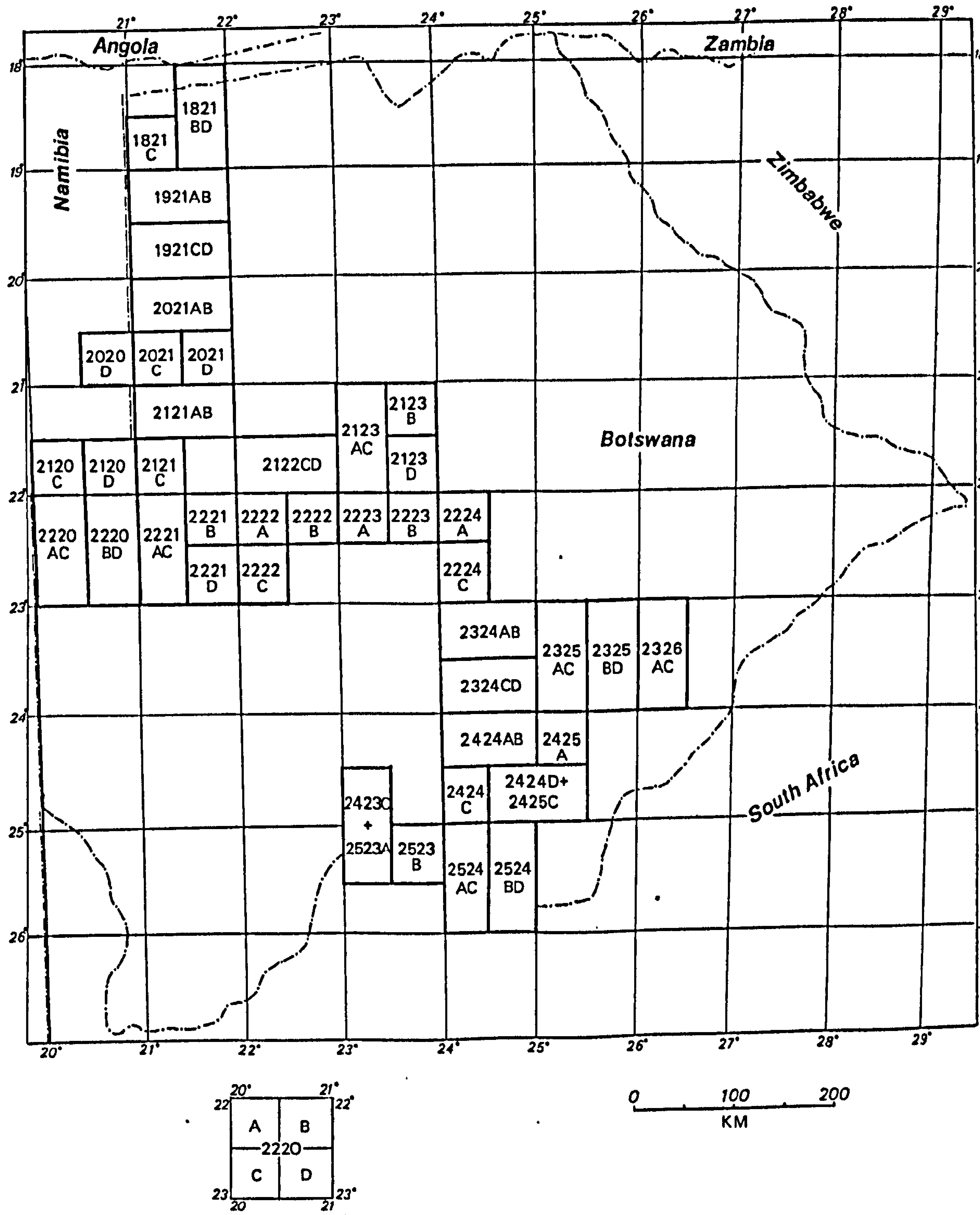


Figure 7.2: Locations of sampling areas (identified by Quarter Degree Square reference number) used in network orientation analysis.

(i) Ordering and numbering of valley networks

Within each quarter degree square, each of the main and tributary valleys is allocated a number, with the main valley designated number 1. Tributary valleys are numbered according to two factors; firstly, their stream order (after Horton, 1945 and Strahler, 1952) within the overall valley network, and secondly, in the case of valleys having the same stream order, their overall length. Thus if the main valley in a QDS (designated number 1) is of stream order 5, then the longest valley of order 4 (if present) was numbered 2, and so on. Valleys are then split into equal segments, each of 10 km in length, to facilitate random sampling. Distances are measured along the central valley course as indicated on maps (see 7.1.4 below) compiled from Landsat imagery by Mallick *et al.* (1981). Segmentation is taken to begin where the main valley crossed the "downstream" boundary of the QDS (in the case of number 1 valleys) and at nodes where a valley joins another valley of higher stream order (in the case of tributary valleys). For each main or tributary valley, the segments are then numbered, starting at the downstream end of the valley.

(ii) Selection of sampling location

Having numbered all valley segments contained within a QDS, a segment is chosen using random number tables, and a point selected at a random distance along its length (measured from the "downstream" end of the segment). In situations where a valley "source" falls within a segment, the segment may still be selected, but if the randomly chosen point falls beyond the valley line indicated on the base map, the point is rejected. In this way, a valley segment has an equal chance of selection regardless of its position within the valley network.

The randomly selected point acts as the centre for a sampling-circle. A sampling-circle is used to avoid the bias effect of using a square in orientational studies. As noted by Abdel-Rahman and Hay (1981), the orientation of the sides of a square quadrat may lead to a bias in the sample. Two circle sizes were selected, both centred on the sample point, with radii 1 cm and 2 cm, representing an actual radius on the ground of 2.5 km and 5.0 km respectively. These two sizes were used to assess the effect of circle size on results, and to identify any scale-dependency in the relationship. Whilst it is clear that a fracture 5.0 km away from a valley is likely to exert little influence on the orientation of that valley, the nature of rock fracturing is such that major fractures tend to have smaller parallel and perpendicular fractures associated with them (Zeil *et al.*, 1991). These fractures are unlikely to be recorded on small scale lineament maps, but their potential significance and effect on bedrock fabric should not be overlooked. Major tensional and shear faults, as occur within the Kalahari basin, are likely to have had great effect on rock fabrics. Densities of structural traces visible on remotely-sensed imagery also vary considerably, mainly dependent upon the thickness of Kalahari Group sediments in an area and upon the occurrence of more recent tectonic activity. Thus, sampling-circles need to have sufficiently large radii to allow analysis even where densities of visible structural traces are low.

(iii) Procedure at each sampling point

At each sample location, the angular difference in orientation between any structural lineament and the valley section within the two sampling-circles is recorded, in 10° class intervals. The value of the angular difference can range between 0° and 90°. Thus if an angle of 55° is measured between a valley line and a lineament, a frequency of one is inserted in the 51-60° class. In the event of no lineaments being present within the sampling-circle, that sample location is discarded for both circle sizes; the number of discarded sample points are noted for each QDS, but not included in the overall sample. The orientation of the valley is taken as being represented by a line joining the two points where the valley crosses the circumference of the sampling-circle. In the event of a channel course crossing the circumference more than twice (e.g. in the case of a meandering channel), the points of initial entry and final intersection are used. Where two valleys are included within the sampling-circle (e.g. at a confluence), only the randomly selected valley upon which the centre of the sampling-circle is located is taken into consideration. If a valley has its source within a sampling-circle then the orientation of the valley is taken as a line joining the source to the point of intersection of the circle boundary. In the event of the sampling-circle overlapping the boundary of the particular QDS under analysis, measurements are carried over into an adjacent square if necessary. Without this rule, areas within 5 km of the edge of a QDS will not be sampled.

Similar rules are applied to the direction of structural lineaments, with the following modification (from Abdel-Rahman and Hay, 1981) if more than one lineament is present within a sampling-circle. In the event of more than one orientation of lineament being present, lineament orientations are plotted in 10° classes on a rose-diagram, with the mid-point of the modal class then taken as the orientational value within the sampling circle. Thus, the angular difference between the modified structural direction and the valley orientation is recorded. In order for a value to be recorded, it is not necessary for the valley and lineament to intersect.

Within each QDS, 50 sampling locations were used. Angular difference observations were recorded for both circle sizes at each sample point, together with the appropriate valley number and segment number for each sample.

(iv) Analysis of data for each QDS

Results for each QDS were analysed using the Minitab statistical package. For each sampling point, Valley Number and Segment Number were recorded together with the Angular Difference values from the 2.5 and 5.0 km sampling-circles. Using Minitab, results for each QDS were classified into ten degree categories ranging from 0-10° to 81-90°, and plotted as histograms. The distribution of angular differences can then be tested for any departures from randomness using the Poisson distribution, as suggested by Abdel-Rahman (1978) and Abdel-Rahman and Hay (1981). Analyses of peaks in the distribution are carried out using the method of Abdel-Rahman and Hay (1981). This technique will not be described in detail, but involves the use of the binomial expansion in the identification of the significant frequency (at a selected probability level) needed within each observational class above which a "peak" frequency can be assumed. Dependent upon the strength of the relationship between valley and structural orientation, peaks can be identified at the 95%, 99% and 99.9% levels.

A peak at the 99.9% level occurs when the number of observations, n, falling within a particular class (from the total of 50 for each QDS) is ≥ 15. A peak at the 99% level is noted when $n \geq 13$ and at the 95% level when $n \geq 11$.

(v) Analysis for individual valleys within a QDS

Having recorded results for each QDS by Valley Number, individual histograms for each valley within a QDS can be produced using Minitab. This allows an assessment of the relationship between geological structure and valley orientation for separate valleys within each sampling area. Histograms can again be produced from results using the 2.5 and 5.0 km sampling-circles.

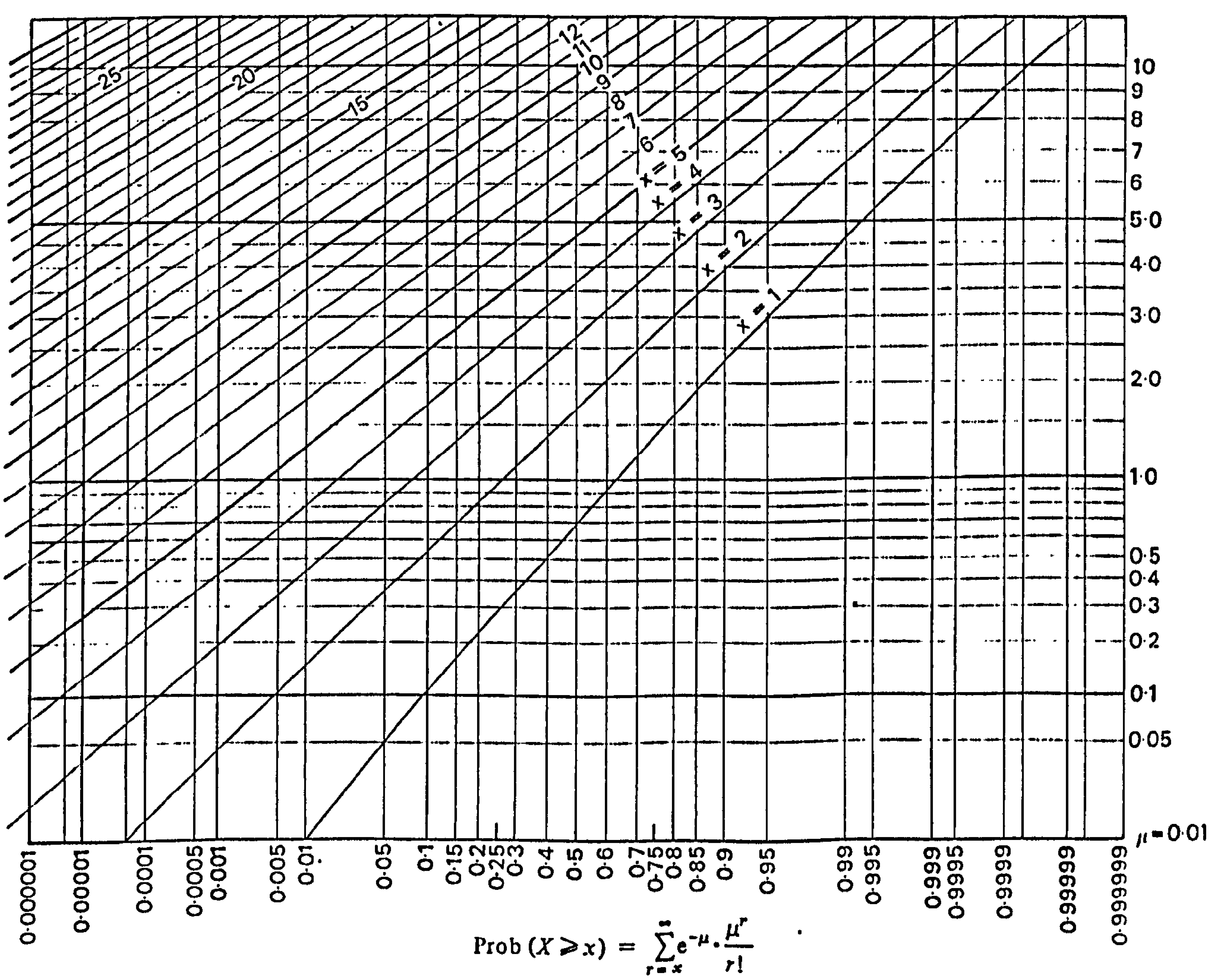


Figure 7.3: Poisson probability chart used in the analysis for individual valleys within a Quarter Degree Square.

To test for significance of results, a Poisson Probability chart showing cumulative probabilities is used, as shown in figure 7.3. To utilise this chart, the mean number of observations which could fall within each of the nine 10° classes on the histogram, μ, is calculated for each valley by dividing the number of sampling locations which fell within a particular valley number by 9. For example, in the case of there only being one valley within a QDS, giving 50 observations between nine classes, $\mu = 5.56$. Reading from figure 7.3, peaks at the 99.9%, 99% and 95% levels occur in this case when the number of observations within a class, n, is $n \geq 15$, $n \geq 13$ and $n \geq 11$ respectively. If 27 of the 50 sampling locations fall within a particular valley, $\mu = 3.00$ and peaks occur when $n \geq 11$ (99.9% level) $n \geq 9$ (99%) and $n \geq 7$ (95%). Only valleys giving a value of $\mu \geq 1$ are included in this stage of the analysis.

Results are discussed in section 7.2.2 for valleys in five groups (figure 7.6); the Northern valleys (the Ncamasere, Xaudum, Groot Laagte and valleys near the Aha Hills), the Okwa/Hanehai system, the Mmone/Quoxo system, the Central Kalahari valleys (Rooibrak-Passarge and Deception) and the Moselebe system.

(vi) Significance of results

If a significant proportion of results fall within the 0-10° class, this suggests a close affinity between valley and structural orientations. Equally, a high frequency of results in the 81-90° class suggests a perpendicularity in the relationship. This may be due to groundwater channelling by *en echelon* fracturing at right angles to the main direction of faulting or fracturing. There are two other possible outcomes of the analysis; a statistically random distribution amongst classes or a peak in a class other than 0-10° or 81-90°. Both of these outcomes suggest no relationship between geological structure and valley network orientation. A random statistical distribution suggests no relationship between variables, whilst a peak in another class may merely indicate a strong regional geological structure. For example, if the 31-45° class contains a peak, it is probable that there is a strong structural trend in an area which the valley orientational trend consistently crosses. This may, however, also indicate tectonic activity after the initial development of a valley.

From these results, quarter degree squares and individual valleys exhibiting a high correlation between valley alignment and geological lineaments can be identified for further examination. Other significant geological factors can then be taken into consideration, such as sedimentological facies changes, which may have influenced valley location.

7.1.4 Determination of Kalahari Group thickness using lithological borehole logs

The depth of bedrock beneath the Kalahari Group also needs to be taken into account in the evaluation of the hypotheses for valley development. Lithological logs from mineral and groundwater exploration boreholes were used to determine the thickness of Kalahari Beds beneath valleys and thus the depth of geological structures in relation to the surface topography. Borehole records (of the type shown in Sekwale, 1984) for all boreholes within 1 km of *mekgacha* courses were studied at the Botswana

Department of Geological Survey in Lobatse. Where sufficiently detailed geological information was included within the borehole record, the thickness of the Kalahari Group was recorded. This enabled the calculation of mean thicknesses (and associated standard deviations) for each QDS.

The results of this study of borehole logs are recorded in figures 7.7 to 7.11 and in table 7.2 of section 7.2.2. Where evidence for deep-weathering of bedrock was identified from borehole logs (see table 7.7), this information is included within the description of results for each of the five groups of valleys in section 7.2.2. In total, 610 borehole records were analysed (790 if those in the Molopo Valley are taken into account) of which 372 contained sufficiently detailed lithological information.

7.1.5 Sources of information used for network analysis

The following sources of information were utilised in the process of analysing network orientation; map and report sources, Landsat imagery and lithological borehole logs. The uses of these are outlined below, together with details of miscellaneous reports used for additional structural information in specific areas.

(a) Map data

Geological structural information was mostly obtained from 1:250,000 maps (held in the Geophysical Section of the Botswana Department of Geological Survey, Lobatse) based on the interpretation of Landsat imagery, which were used in the compilation of the 1:1,000,000 photogeological map of Botswana included in Mallick, Habgood and Skinner (1981). Additional structural information at a scale of 1:250,000 was extracted from maps and reports compiled by Terra Surveys (1978) as a result of the Reconnaissance Aeromagnetic Survey 1975-76. This was utilised to provide sub-surface structural information to supplement the surficial data displayed on maps derived from Landsat imagery.

Maps produced by the Botswana Government Department of Surveys and Lands in Gaborone at a variety of scales were also used for network locational mapping, together with the 1:250,000 interpretations of Landsat imagery by Mallick *et al.* (1981) described above.

(b) Landsat data and aerial photography

Valley networks were mapped from a variety of corrected Landsat images at a scale of 1:250,000 (held at the U.N.D.P./F.A.O. Soil Survey in Gaborone), with additional mapping from aerial photography for precision in headwater areas (at the Botswana Government Department of Surveys and Lands in Gaborone).

(c) Lithological borehole logs

Logs were recorded from borehole completion certificates (as described by Sekwale, 1984) and reports held at the Botswana Departments of Geological Survey and Water Affairs, in Lobatse and Gaborone respectively, with further logs from McDaid (1985) for eastern Namibia, and from Gale (1975), Herbert

(1976), Union Carbide (1977, 1978, 1979 *a* to *d*, and 1980 *a* to *d*) and De Beers Prospecting Botswana (Pty) Ltd (1988) for specific valleys in Botswana.

(d) Miscellaneous structural data

Numerous other sources of structural information were utilised for specific regions where more detailed geological surveys have been carried out. These include reports by Albat (1978), Hegenberger (1982) and Hegenberger and Seeger (1980) for northeastern Namibia, McDaid (1985) for eastern Namibia, Wright (1978), Litherland (1982) and Lüdtke (1986) and for northwestern Botswana, Jennings and Crockett (1961), Crockett and Jennings (1962, 1964 and 1965), Litherland (1982) and Aldiss (1987*b*, 1988), for the Okwa Valley, Botswana, Farr *et al.* (1981), Buckley (1984), Tombale (1986), Timje (1987) and Aldiss *et al.* (1989) for the area around Letlhakeng, Botswana, Kimbell *et al.* (1984) and Gould *et al.* (1987, 1989) for the Molopo and Moselebe valley systems, southern Botswana, and Thomas *et al.* (1988) for the Nossop and Twee Rivieren areas of south-western Botswana and the northern Cape Province of South Africa.

Each of these sources of data have certain limitations, especially where data is drawn from secondary sources. In terms of geological structure the principal limitation is due to the uncertainty regarding the actual structure causing a surface lineament in areas with great thicknesses of Kalahari Beds or lacking bedrock outcrops or subcrops. As Norman (1968) notes, a photo-lineament in an area with superficial deposits may indicate a variety of structural variables. In many of the areas for which more detailed geological reports exist the locations of tensional and shear faults are shown, with other significant structural information also indicated. However, as noted in the introduction to this section, the particular type of fracture most likely to permit groundwater flow is still unknown in the context of Kalahari groundwater studies, and hence the type of structure with most potential for groundwater sapping remains unclear. One of the underlying assumptions of the use of the Poisson distribution in testing for deviations from randomness is a requirement of equal-number density (Abdel-Rahman and Hay, 1981) i.e. there should be equal numbers of each structural type included in the analysis. Given the available geological information, this factor is impossible to ascertain apart from within areas where detailed geological surveys have been undertaken.

A further problem associated with the fracture-lineament patterns used for structural analysis is that not all lineaments visible on Landsat imagery were included on the 1:250,000 base maps for the following reason noted by Mallick *et al.* (1981 p.31);

> "...so many fractures and lineaments were identified from the imagery that only the longer and possibly more significant elements could be selected to be shown on the thirty three compilations at 1:250,000...".

Whilst all available sources of structural data were utilised, the use of other remotely-sensed methods of lineament identification (e.g. studies of NOAA-AVHRR imagery, as used in Australia by Tapley, 1988) may provide even more structural information. The problems of mapping valley networks have already been noted, particularly in headwater areas of drainage networks (Ovenden and Gregory, 1980; Burt and

Gardiner, 1982; Knighton, 1984; Burt and Oldman, 1986). This problem is especially prevalent in the Kalahari where relief is negligible and drainage divides are often not in evidence. Many of the tributaries of the Okwa Valley in the central Kalahari have indistinct headwaters where they are buried by aeolian sand, as do the valleys south of Letlhakeng in the Kweneng District of Botswana. Finally, there are often problems regarding the lithological records included in borehole logs, mainly due to inaccuracies and subjectivities in the interpretation of rock type by the borehole logger. This causes problems in determining the change from sub-Kalahari to Kalahari Beds, necessary for the identification of the thickness of the Kalahari Beds.

7.2 Results of network orientation analysis

The results of the analysis of network and structural orientation are shown in tables 7.1 to 7.6 and in figures 7.4, 7.5, and 7.7 to 7.11. As a result of the method of segment numbering, allowing the identification of individual valley segments, the results may be considered in two parts; firstly, by quarter degree square and secondly by individual valley or valley system.

7.2.1 Results of network orientation analysis by QDS

The results of analysis within whole quarter degree squares, or combinations of squares, are shown in table 7.1. Figure 7.2 should be used in conjunction with the tables as a key to QDS cell locations. The table includes information on the orientational classes in which "peaks" in angular difference between valley and structural alignment occurred, and the associated strength of statistical significance. These results are given for both sizes of sampling-circle, with details of the total number of sample points which yielded no structural information (and were hence disregarded) also indicated for each QDS.

Figures 7.4 and 7.5 give simplified spatial views of the overall results for the 2.5 km and 5.0 km sampling-circles shown in table 7.1. The figures show a tripartite classification of results; a QDS is classified as displaying "No Peak", a "Peak in Other Class" or as having a "Peak in 0-10° Class". The first category indicates squares in which no statistical peak in the orientational data occurred i.e. no relationship between valley and structural orientation occurred. The second and third categories include cells where significant peaks between valley alignment and structure occurred, with the latter category ("Peak in 0-10° Class") of most significance in evaluating valley development hypotheses.

It should be noted that figures 7.4 and 7.5 necessarily show a simplified version of the data in table 7.1. Inspection of the table reveals that it is possible for a QDS to have statistically significant peaks in more than one orientational class. The figures, however, only show the most significant of these classes. In certain cases more than one peak class has the same level of statistical significance; where one of these is the 0-10° class, only this class is shown on the figures. As noted above, the presence of a strong peak within this class (indicating parallelism of structure and valley orientation) is of particular significance in the evaluation of hypotheses for valley development, suggesting strong evidence for structural control.

Table 7.1: Peak classes arising from network orientation analysis by Quarter Degree Square for 2.5 and 5.0 km sampling-circles. "No" indicates no statistically significant peak.

Quarter Degree Square (QDS)	2.5 km Circle Peak Class (°)	2.5 km Circle Signif. Level	5.0 km Circle Peak Class (°)	5.0 km Circle Signif. Level	Samples with no Lineaments
1821BD	No		31-40	95.0%	11
1821C	71-80	99.9%	71-80	99.9%	10
1921AB	31-40	95.0%	41-50	99.0%	12
			81-90	95.0%	
1921CD	0-10	95.0%	No		21
2020D	0-10	99.9%	0-10	99.0%	2
	11-20	99.0%	11-20	99.9%	
2021AB	31-40	99.0%	31-40	99.9%	9
	51-60	95.0%	51-60	99.9%	
2021C	0-10	99.9%	0-10	99.9%	0
2021D	No		0-10	95.0%	2
			71-80	95.0%	
2120C	No		21-30	95.0%	10
			41-50	95.0%	
2120D	0-10	99.9%	0-10	99.0%	0
			31-40	95.0%	
2121AB	41-50	99.0%	41-50	95.0%	17
2121C	0-10	95.0%	No		0
2122CD	51-60	99.0%	51-60	95.0%	1
	61-70	99.0%	61-70	99.9%	
2123AC	71-80	95.0%	71-80	95.0%	5
2123B	0-10	99.9%	0-10	99.9%	0
	51-60	99.0%	51-60	99.0%	
2123D	21-30	99.0%	21-30	95.0%	13
			31-40	99.0%	
2220AC	0-10	99.9%	0-10	99.9%	3
2220BD	No		No		9
2221AC	0-10	99.9%	0-10	99.9%	3
2221B	0-10	99.0%	0-10	95.0%	11
			11-20	95.0%	
			21-30	95.0%	
2221D	0-10	99.9%	0-10	99.9%	1
	11-20	95.0%			
2222A	0-10	95.0%	0-10	99.9%	18
	21-30	95.0%	11-20	95.0%	
	31-40	95.0%	21-30	95.0%	
2222B	0-10	99.0%	0-10	99.9%	32
	11-20	99.9%	11-20	99.9%	
2222C	0-10	99.9%	0-10	99.9%	25
	11-20	99.9%	11-20	99.9%	

Table 7.1 (cont.): Peak classes arising from network orientation analysis by Quarter Degree Square for 2.5 and 5.0 km sampling-circles. "No" indicates no statistically significant peak.

Quarter Degree Square (QDS)	2.5 km Circle		5.0 km Circle		Samples with no Linea-ments
	Peak Class (°)	Signif Level	Peak Class (°)	Signif Level	
2223A	0-10	99.9%	0-10	99.0%	9
2223B	0-10	99.0%	0-10	99.0%	12
	11-20	99.9%	11-20	99.9%	
			21-30	99.0%	
2224A	0-10	99.0%	0-10	99.0%	16
	11-20	99.0%	21-30	95.0%	
2224C	No		51-60	99.9%	40
2324AB	No		0-10	95.0%	10
2324CD	No		No		11
2325AC	0-10	99.9%	0-10	99.9%	2
2325BD	0-10	99.9%	0-10	99.9%	0
	11-20	95.0%	11-20	95.0%	
2326AC	0-10	99.9%	0-10	99.9%	2
2424AB	0-10	95.0%	0-10	95.0%	16
2424D & 2425C	0-10	99.9%	0-10	99.9%	14
2425A	0-10	99.9%	0-10	99.9%	9
2423C & 2523A	51-60	99.0%	41-50	95.0%	30
	81-90	99.0%	61-70	99.9%	
2523B	51-60	99.0%	No		36
2524AC	81-90	95.0%	11-20	99.0%	8
2524BD	21-30	99.0%	11-20	99.0%	6
	41-50	95.0%			

An example of the occurrence of more than one peak orientational class can be seen from the results for QDS 2222A. The results for both sizes of sampling-circle show peaks in three separate orientational classes. In the case of the 5.0 km circle, table 7.1 gives the most statistically significant class as 0-10°. However, for the 2.5 km circle, all peak classes are significant at the 95.0% level, and only the 0-10° class is shown on figure 7.4.

Considering the results in table 7.1, two points should be noted. In general, where more than one peak class is indicated in the table, and one of these classes is 0-10°, the other classes are usually 11-20° and/or 21-30°. This would tend to reinforce the significance of the results. Furthermore, where multiple peaks occur and none are 0-10°, the classes tend not to be adjacent to one another. A second observation is that where "No Peak" is indicated for one sampling-circle size, it is usually also indicated for the other circle size or otherwise a peak other than 0-10° is present.

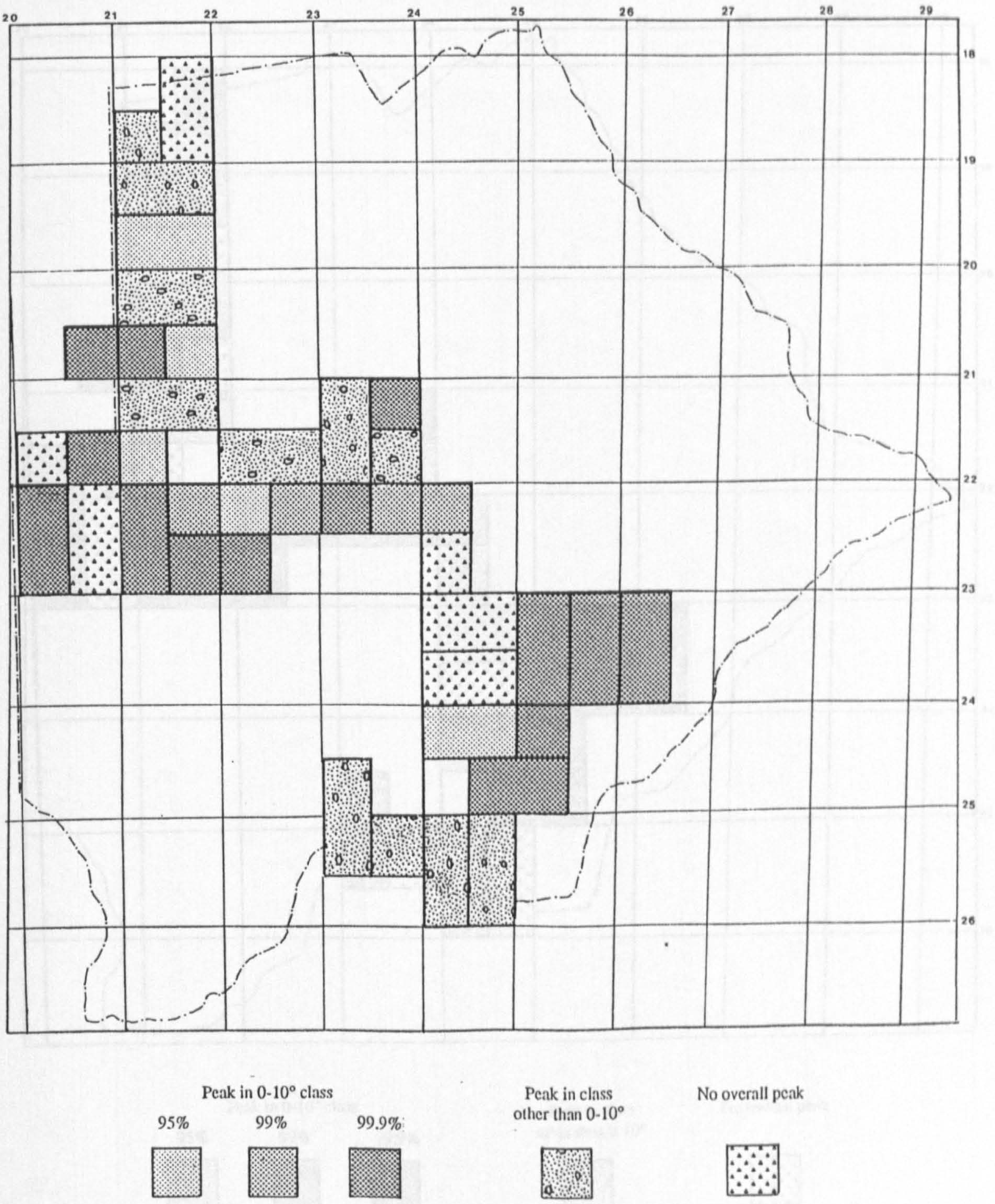


Figure 7.4: Results of network orientation analysis using 2.5 km sampling-circle.

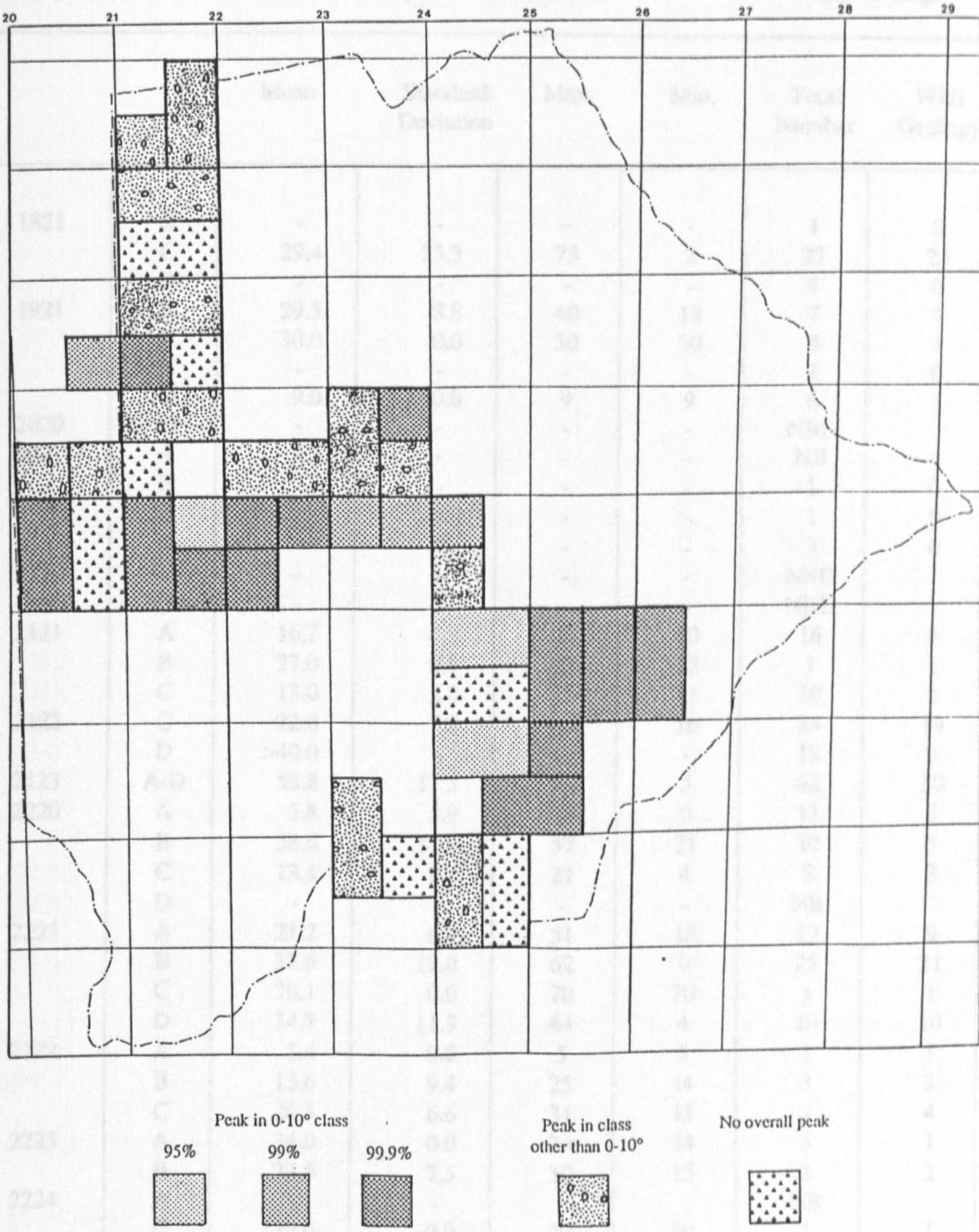


Figure 7.5: Results of network orientation analysis using 5.0 km sampling-circle.

Table 7.2: Thickness of Kalahari Group sediments from lithological borehole logs by Quarter Degree Square.

Quarter Degree Square		Thickness of Kalahari Group sediments (metres)				Number of Borehole Logs	
		Mean	Standard Deviation	Max.	Min.	Total Number	With Geology
1821	B	-	-	-	-	1	0
	C	29.4	23.3	75	2	27	25
	D	-	-	-	-	4	0
1921	A	29.3	8.8	40	18	7	7
	B	30.0	0.0	30	30	8	1
	C	-	-	-	-	3	0
	D	9.0	0.0	9	9	6	1
2020	D	-	-	-	-	NBD	
2021	A	-	-	-	-	NB	
	B	-	-	-	-	1	0
	C	-	-	-	-	1	0
	D	-	-	-	-	3	0
2120	C	-	-	-	-	NBD	
	D	-	-	-	-	NBD	
2121	A	16.7	4.8	24	10	16	6
	B	27.0	0.0	27	27	1	1
	C	13.0	2.0	15	11	10	2
2122	C	22.8	7.2	40	16	24	19
	D	>40.0	-	-	-	12	9
2123	A-D	58.8	17.5	96	5	62	30
2220	A	5.8	5.9	18	0	13	8
	B	38.6	17.6	57	21	10	5
	C	13.4	9.2	21	4	3	3
	D	-	-	-	-	NB	
2221	A	21.2	6.6	51	18	13	9
	B	32.6	19.0	62	0	25	21
	C	70.1	0.0	70	70	1	1
	D	14.7	11.9	44	4	10	10
2222	A	5.4	0.0	5	5	1	1
	B	13.6	9.4	25	14	3	3
	C	26.5	6.6	31	15	7	4
2223	A	34.0	0.0	34	34	3	1
	B	22.5	7.5	30	15	2	2
2224	A	-	-	-	-	NB	
	C	30.0	0.0	30	30	1	1

Table 7.2 Continued:

Quarter Degree Square		Thickness of Kalahari Group sediments (metres)				Number of Borehole Logs	
		Mean	Standard Deviation	Max.	Min.	Total Number	With Geology
2324	A	39.9	22.2	66	11	4	3
	B	44.0	16.0	60	28	2	2
	C	36.0	12.7	45	18	4	3
	D	15.2	11.4	42	2	52	35
2325	A	-	-	-	-	NB	
	B	12.3	5.3	21	6	11	7
	C	24.0	18.4	62	1	28	27
	D	20.8	21.6	82	1	26	13
2326	A	9.3	5.4	13	1	10	7
	C	6.7	4.0	12	3	10	4
2423	C	-	-	-	-	NB	
2424	A	33.0	9.0	42	24	2	2
	B	31.2	4.6	40	24	11	6
	D	47.5	4.2	53	43	3	3
2425	A	15.8	10.4	45	1	46	40
	C	8.2	3.7	12	4	3	2
2522	A-D	-	-	-	MOL	23	23
2523	A	66.6	28.8	105	36	8	5
	B	93.8	44.0	183	36	34	14
2524	A	71.9	26.9	120	22	34	20
	B	46.2	24.2	83	6	18	16
	C	-	-	-	MOL	17	17
	D	94.0	0.0	94	94	3	1
2525	A-D	0.0	0.0	0	0	11	11
2620	A-D	-	-	-	MOL	42	42
2621	A-D	-	-	-	MOL	21	21
2622	A-D	-	-	-	MOL	19	19
TOTAL						790	552

Key to Table 7.2:

NBD; No borehole data available (Namibian Boreholes).
NB; No boreholes situated in valleys within QDS.
MOL; Borehole data available for Molopo valley within QDS, but not indicated in table as Molopo not included in network analysis.

Finally, perhaps most important in the interpretation of the results shown for whole quarter degree squares in figures 7.4 and 7.5 is that there is some loss of information for individual valleys within a QDS. It is possible that one valley alone within a QDS may demonstrate close alignment between valley course and geological structure, but this result is lost by considering the QDS as a whole. However, the analysis of alignments at the QDS level does give a good indication of regional relationships between valley courses and geological structure. The more detailed results of analysis at the individual valley level are discussed in the following section.

7.2.2 Results of network orientation analysis for individual valleys & valley systems

Tables 7.3 to 7.6 and table 7.8 indicate results for individual valleys and valley systems, with table 7.3 giving results for Northern valleys, table 7.4 for the Okwa/Hanehai system, table 7.5 for Central Kalahari valleys, table 7.6 for the Quoxo/Mmone and Serorome systems, and table 7.8 for valleys of the Moselebe system. Reference should be made to figure 7.6 for the location of these five systems. The tables include information for each valley by QDS, and indicate results for main and tributary valleys within a QDS for both 2.5 km and 5.0 km sampling-circles. Again, results are recorded as peak classes with associated levels of significance, or as no peak. In the analysis, valleys or tributaries which were sampled less than 9 times (from the total sample of 50 per sampling area) were considered as too small a sample for statistical inferences to be made, for reasons discussed in section 7.1.3.

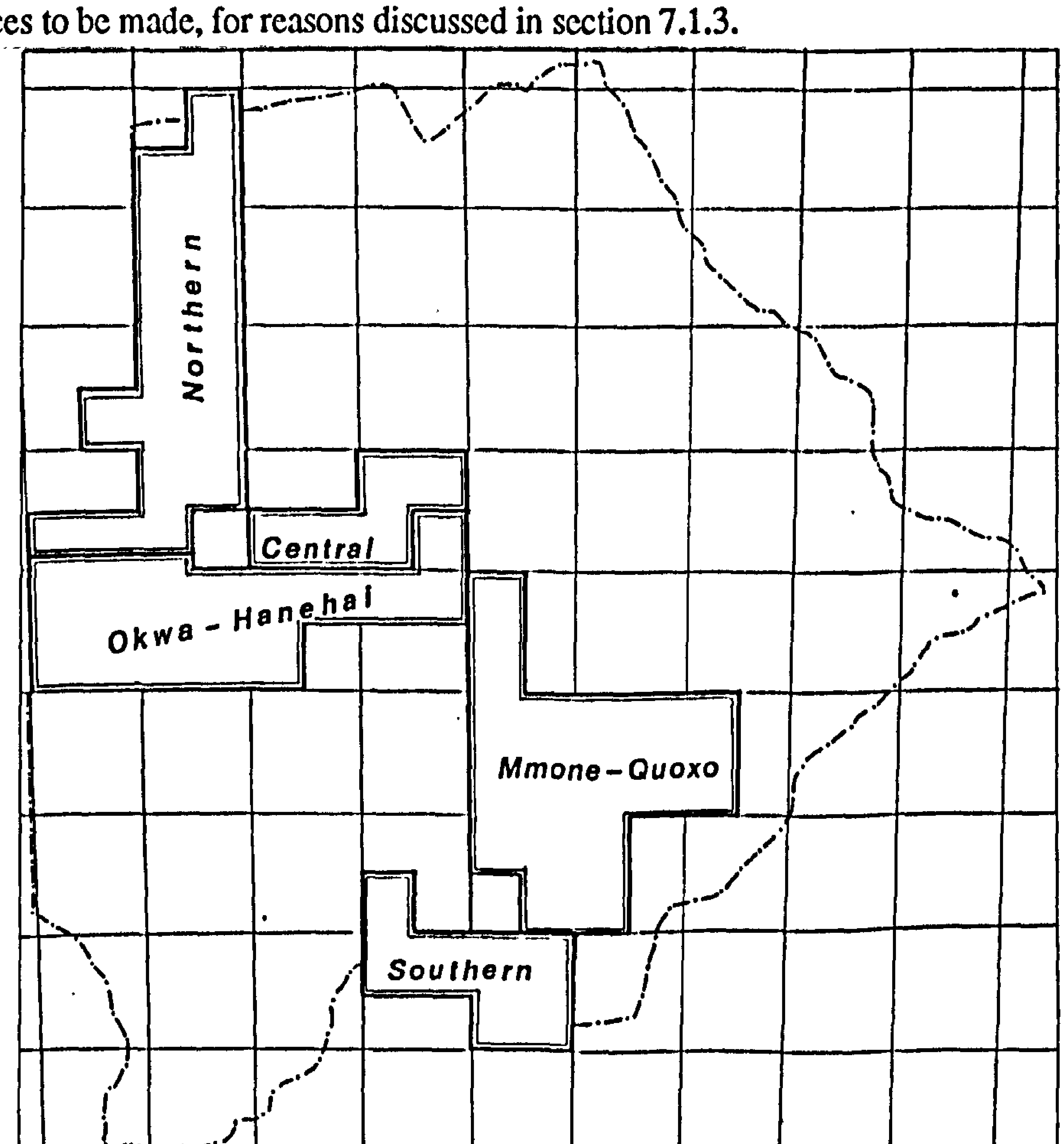


Figure 7.6: Locations of individual systems of Kalahari *mekgacha* used in network orientation analysis.

Figures 7.7 to 7.11 give the results of analysis by individual valley for each of the five groups of valleys shown on figure 7.6. Results are again classified into "Peak in 0-10° Class", "Peak in Other Class" or "No Peak" and shaded appropriately. As discussed in section 7.2.1, it is possible for a valley to show a statistically significant alignment with geological structures from more than one orientational class. Again, the class showing the strongest level of significance is shown on the figures.

Information on the mean thickness of Kalahari Group sediments beneath or in the immediate vicinity of valley courses is also shown on figures 7.7 to 7.11. These results are summarised in table 7.2. Thicknesses are indicated on the figures along with information on total numbers of boreholes and the number of logs containing detailed geological information for each QDS. The results of network orientation analysis for each valley system will now be discussed in turn.

(a) Northern valleys

Results for the Northern valleys and their tributaries are shown in figure 7.7 and table 7.3. The general result from this area is a lack of any strong relationship between valleys and structural lineaments. Of the Northern valleys, the only areas showing a peak in the 0-10° class are the Groot Laagte in QDS 2120D, a tributary of the Epukiro in QDS 2021C and a tributary valley in the Aha Hills area in QDS 1921CD. The dominant structural alignment in the area is due to neotectonic movements associated with the subsidence of the Okavango Delta (Reeves, 1977), with major faults trending approximately northeast-southwest. As most valleys in the region are directed towards the Okavango Delta and trend generally west-east, it is not surprising that the majority of peak orientational classes occur in the range 30-60°. The fact that valleys near major bedrock outliers (where the Kalahari Group sediments are comparatively thin) show the strongest relationship between alignment and structure is of interest, suggesting that valleys may have been structurally controlled prior to the commencement of rifting in the Okavango graben. It is possible that valleys developed on a pre-Kalahari surface (as already discussed for the Moselebe Valley in section 5.3.3*a*) and have migrated upwards as the Kalahari Group sediments accumulated. This would explain the alignment of valleys near bedrock outcrops with structure pre-dating the Okavango graben.

It should be noted that the density of lineaments in the vicinity of the Northern valley systems is relatively low; this can be gauged from the number of sampling-circles within which no lineament occurred (as shown in the final column of table 7.1). With a low lineament density the occurrence of any lineament near a valley course is likely to have a strong influence on the final results of the analysis.

Whether the lack of any significant relationship between valley alignment and geological structure is affected by the thickness of the sediments of the Kalahari Group overlying Basement rocks is difficult to ascertain. Many of the QDS contain either no boreholes or those that have logs available contain no geological information. The most detailed logs are those in the Ncamasere, Xaudum and Groot Laagte, drilled by Union Carbide (Union Carbide, 1980*d*), which reveal mean Kalahari Group thicknesses of around 30 m, with a maximum of 75 m.

Table 7.3: Individual results of Network Orientation analysis by QDS for Northern valleys and tributaries. "No" indicates no statistically significant peak. "Small" indicates valley sampled less than nine times.

Valley Name	QDS	Main/ Trib	2.5 km Circle Peak°	2.5 km Circle Level	5.0 km Circle Peak°	5.0 km Circle Level
Ncamasere	1821C	Main	71-80	99.9%	71-80	99.9%
	1821BD	Main	11-20	95.0%	11-20	99.9%
		Trib	51-60	95.0%	31-40	99.9%
			71-80	99.0%	71-80	99.0%
Xaudum	1821C	Main	51-60	99.9%	61-70	99.9%
	1921AB	Main	31-40	95.0%	41-50	99.9%
					81-90	95.0%
Kaua	1921CD	Main	51-60	95.0%	51-60	95.0%
Aha Hills area	1921CD	Main	51-60	99.0%	51-60	95.0%
		Trib	0-10	99.9%	0-10	99.9%
Gcwihabedum	2021AB	Main	31-40	95.0%	31-40	95.0%
			61-70	95.0%	51-60	95.0%
Epukiro	2020D	Trib	11-20	95.0%	11-20	95.0%
	2021C	Main	No		21-30	99.0%
		Tribs	71-80	99.0%	71-80	95.0%
			0-10	99.9%	0-10	99.9%
	2120C	Main	51-60	95.0%	41-50	99.9%
		Trib	Small		Small	
Groot Laagte	2120D	Main	0-10	99.9%	0-10	99.9%
	2121AB	Main	41-50	99.0%	41-50	99.9%
	2021C	Main	Small		Small	
	2021D	Main	31-40	99.0%	No	
			61-70	95.0%	71-80	99.0%

Figure 7.7: Results of network orientation analysis for individual valleys of the Northern valley systems.

The contour map of Kalahari Group thicknesses produced by Thomas (1988*b*) shows values in the range 51-200 m for this region, which would appear to be an overestimate unless there is significant thinning of the Kalahari Group sediments beneath valleys. Whilst the depth of incision of the valleys into the Kalahari Group sediments is not taken into account when using only those borehole logs drilled in valleys, this factor could only account for at most a further 20 m thickness.

(b) Okwa/Hanehai system

In terms of its length the Okwa/Hanehai system (shown in figure 7.8) represents the largest Kalahari *mekgacha* network, the main channel of the Okwa being in excess of 600 km long. As can be seen from table 7.4, in general the Hanehai Valley only shows close structural alignment in QDS 2121C and 2222A, with other peak classes occurring more frequently. This relates in particular to the regional strike associated with the beds of the Ghanzi Formation which trends across the general course of the valley.

The Okwa Valley shows a close alignment to geological structure along its entire course. There are spatial variations, however, in the elements of the system showing closest geological alignment. For example, table 7.4 shows that in the western part of the network, the main valley has statistically significant peaks in the 0-10° class, whilst towards the central Kalahari it is the Okwa's tributary valleys which are most closely aligned to geological lineaments. There are exceptions to this generalisation; in QDS 2220AC both main and tributary valleys show alignment. Also, within a particular QDS it is uncommon for all tributaries to show structural alignment, as in the case of QDS 2223A. Again, lineament density varies, with the lowest values occurring in the central Kalahari.

Borehole logs show, as would be expected, a general thickening of the Kalahari Group towards the east and south, although significant thinning occurs in the vicinity of the Precambrian inlier in the Okwa Valley near Tswaane veterinary post (QDS 2221B). Thicknesses range from 0 to 70 m in agreement with Thomas (1988*b*), with mean values between 15 and 38 m. There is no apparent relationship between Kalahari Group thickness and valley and structural alignment, with most valleys showing evidence of structural control. As noted above, there is a tendency for the main channel to show the closest structural alignment in the west, whilst tributaries show more significant results in the east. This spatial variation may be connected to the general increase in Kalahari Group thicknesses to the east, but is probably influenced by the presence of major SW and WSW trending faults in this region.

(c) Central Kalahari valleys

The results for the Rooibrak/Passarge and Deception valleys (figure 7.9 and table 7.5) generally show no evidence for structural control, with peak classes falling in the range 30-60°. The exception is within QDS 2123B where the Deception Valley course abruptly swings to the north and east to align itself with a series of minor lineaments. Only a short section (< 30 km) of the Passarge Valley occurs within QDS 2123B, but this short section also shows evidence for strong structural control.

Table 7.4: Individual results of Network Orientation analysis by QDS for the Okwa/Hanchai valleys and tributaries. "No" indicates no statistically significant peak. "Small" indicates valley sampled less than nine times.

Valley Name	QDS	Main/ Trib	2.5 km Circle Peak°	Level	5.0 km Circle Peak°	Level
Hanehai	2120C	Main	No		21-30	99.0%
	2120D	Main	31-40	99.0%	No	
		Trib	61-70	95.0%	71-80	95.0%
	2121C	Main	0-10	95.0%	No	
	2221B	Main	11-20	95.0%	61-70	95.0%
	2222A	Main	21-30	95.0%	0-10	99.9%
					21-30	95.0%
Okwa	2220AC	Main	0-10	95.0%	11-20	95.0%
			31-40	99.0%		
		Tribs	0-10	95.0%	0-10	99.9%
					0-10	95.0%
	2220BD	Main	No		11-20	99.0%
					21-30	99.0%
		Tribs	No		No	
	2221AC	Main	0-10	99.9%	0-10	99.9%
		Tribs	51-60	95.0%	51-60	95.0%
			No		61-70	99.0%
			No		No	
	2221B	Main	0-10	99.0%	0-10	95.0%
					11-20	95.0%
					21-30	95.0%
		Tribs	Small		Small	
	2221D	Tribs	0-10	99.9%	0-10	99.9%
			11-20	99.9%	71-80	99.9%
			71-80	99.0%		
	2222A	Main	No		41-50	95.0%
		Tribs	Small		Small	
	2222B	Main	11-20	99.9%	0-10	99.0%
			21-30	95.0%	11-20	99.0%
	2222C	Tribs	11-20	95.0%	21-30	99.9%
			41-50	95.0%	No	
	2223A	Main	No		21-30	99.0%
		Tribs	0-10	99.9%	0-10	99.9%
		Tribs	No		No	
	2223B	Main	11-20	99.0%	11-20	99.9%
		Tribs	0-10	99.9%	0-10	99.9%
			11-20	95.0%		
	2123D	Main	21-30	95.0%	21-30	99.0%
		Tribs	Small		Small	
	2224A	Main	No		51-60	99.0%

Figure 7.8: Results of network orientation analysis for individual valleys of the Okwa/Hanehai valley system.

Table 7.5: Individual results of Network Orientation analysis by QDS for Central Kalahari valleys and tributaries. "No" indicates no statistically significant peak. "Small" indicates valley sampled less than nine times.

Valley Name	QDS	Main/ Trib	2.5 km Circle Peak°	2.5 km Circle Level	5.0 km Circle Peak°	5.0 km Circle Level
Rooibrak	2122CD	Main	51-60 81-90	95.0% 95.0%	51-60 81-90	99.9% 95.0%
Passarge	2123AC	Main Tribs	No Small		71-80 Small	95.0%
	2123B	Main	0-10	99.9%	0-10	99.9%
Deception	2122CD	Main	51-60 61-70	95.0% 95.0%	61-70	99.9%
	2123AC	Main	41-50	95.0%	31-40	95.0%
	2123B	Main	0-10	99.9%	0-10	99.9%

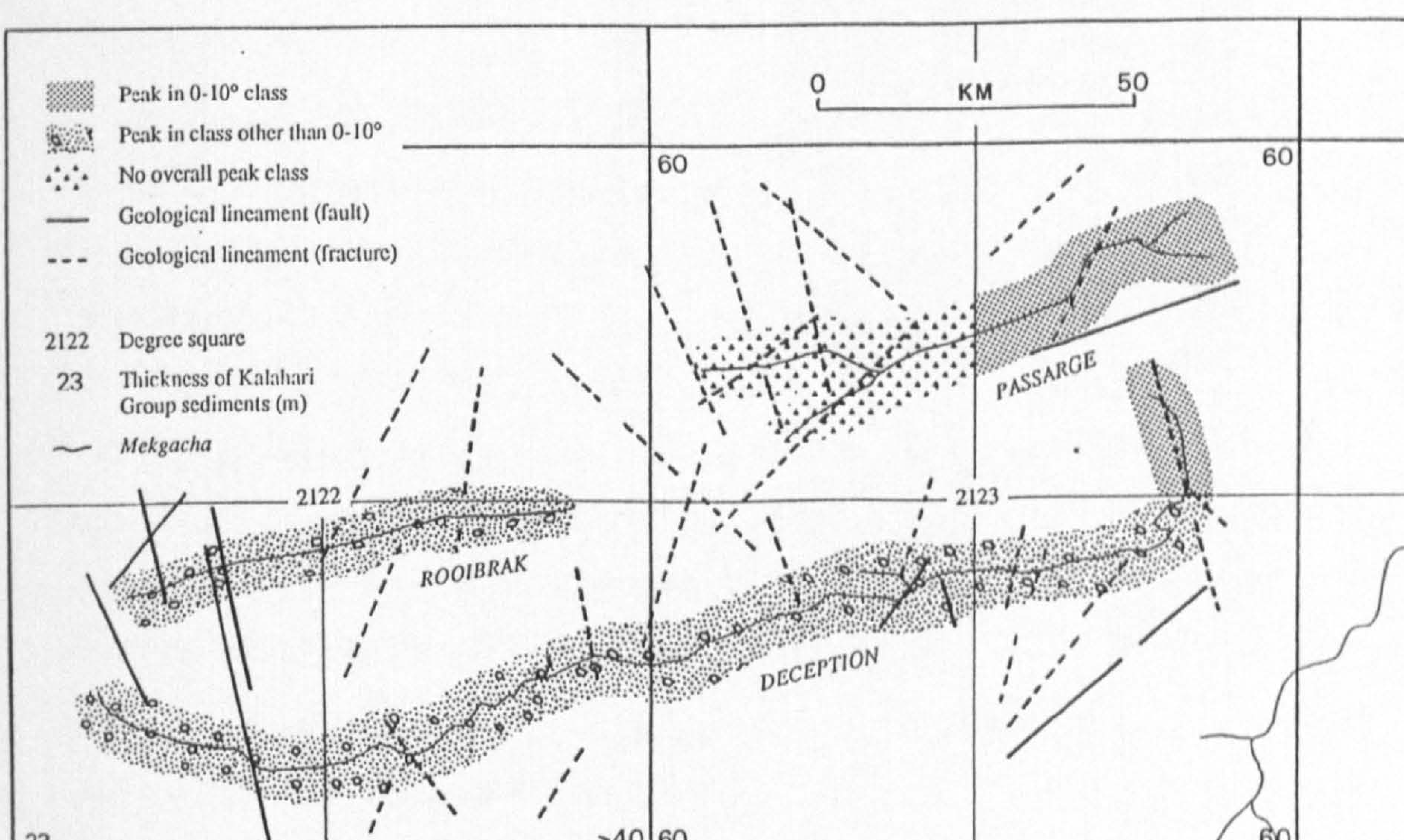


Figure 7.9: Results of network orientation analysis for individual valleys of the central Kalahari valley system.

There appears to be very little correlation between Kalahari Group thicknesses and the spatial distribution of results. The Rooibrak Valley contains a large number of boreholes with detailed lithological information as a result of drilling by Union Carbide (1979*d*, 1980*d*). These show that the Kalahari Group exceeds 40 m in thickness in QDS 2122D, as boreholes fail to penetrate the entire sequence of sediments. The thickness of the sediments in QDS 2123 ranges from 5 to 96 m with deeper values found mainly in the west of the QDS. It is therefore of potential significance that the sections of the valley systems showing strongest structural alignment tend to occur in the eastern part of QDS 2123.

(d) Mmone/Quoxo and Serorome systems

The numerous main valleys and tributaries of the Mmone/Quoxo system make up the second largest Kalahari *mekgacha* network. The system also shows strong structural control (figure 7.10 and table 7.6), particularly in the headwater regions in the vicinity of the villages of Letlhakeng and Jwaneng. The main Mmone/Quoxo or Meratswe Valley only exhibits strong structural alignment in QDS 2324AB and 2224A (which also contains the lowest density of lineaments) in its downstream sections, with the main evidence for structural control being found in tributary valleys. These include the Naledi/Khwakhwe, Dikgonnyane and Kohiye valleys, as well as those in the Letlhakeng area. Strongest structural alignment is shown by the Dikgonnyane Valley, which follows the line of a major lineament in a north-northeasterly direction for over 45 km before swinging west to follow another lineament. The Naledi/Khwakhwe Valley shows similar strong control, following lineaments closely over much of its length before joining the main Mmone/Quoxo channel.

The Letlhakeng valleys of Shaw and De Vries (1988), with the exception of Valley 3, generally show a close orientational affinity with geological structures. This relationship is especially apparent in the case of Letlhakeng Valley 1, the Gaotlhobogwe Valley. Whilst Letlhakeng Valley 3 shows no significant structural alignment, there is evidence for structural control of the two tributaries which enter the main valley around 6 km south of Letlhakeng village.

The Serorome Valley also exhibits marked structural control, showing an abrupt change of direction in its course following the alignment of lineaments associated with the Zoetfontein Fault. The main fault itself does not appear to have any major controlling influence upon the orientation of the valley course, but other fractures presumably associated with movements along the fault-line are almost certainly important.

The Mmone/Quoxo system shows an apparent relationship between the results of network analysis and Kalahari Group thickness. Sediment thicknesses are greatest on average in degree squares 2324 and the northern half of 2424, coinciding with a lack of close correlation between valley alignment and structure. Peaks in the 0-10° class are more common towards the south and east nearer the Kalahari edge where the sediments are generally thinner. Within this system it would appear that structural alignment is most likely where the sediment cover is thin, but not nonexistent.

Table 7.6: Individual results of Network Orientation analysis by QDS for the valleys and tributaries of the Mmone/Quoxo and Serorome valley systems. "No" indicates no statistically significant peak. "Small" indicates valley sampled less than nine times.

Valley Name	QDS	Main/ Trib	2.5 km Circle Peak°	2.5 km Circle Level	5.0 km Circle Peak°	5.0 km Circle Level
Naledi & Khwakhwe	2425C/ 2424D	Main	0-10	99.9%	0-10	99.9%
	2424AB	Main	0-10	99.0%	0-10	99.0%
			31-40	95.0%	21-30	95.0%
	2324CD	Main	0-10	95.0%	31-40	95.0%
		Trib	11-20	99.0%	No	
	2324AB	Main	Small		Small	
Letlhakeng Valley 1	2425A	Main	0-10	99.9%	0-10	99.9%
Letlhakeng Valley 2	2425A	Main	0-10	99.0%	0-10	99.9%
		Tribs	Small		Small	
Letlhakeng Valley 3	2425C/ 2424D	Main	61-70	99.0%	71-80	95.0%
			71-80	99.9%	61-70	95.0%
	2425A	Main	Small		Small	
		Tribs	Small		Small	
Letlhakeng Valley 4	2425A	Main	Small		Small	
Dikgonnyane	2425A	Main	Small		Small	
	2325AC	Main	0-10	99.9%	0-10	99.9%
		Tribs	Small		Small	
	2324AB	Main	Small		Small	
Kohiye	2425A	Main	Small		Small	
	2325AC	Main	0-10	99.9%	0-10	99.9%
			11-20	95.0%	11-20	95.0%

Table 7.6 (Cont.): Individual results of Network Orientation analysis by QDS for the valleys and tributaries of the Mmone/Quoxo and Serorome valley systems. "No" indicates no statistically significant peak. "Small" indicates valley sampled less than nine times.

Valley Name	QDS	Main/ Trib	2.5 km Circle Peak°	2.5 km Circle Level	5.0 km Circle Peak°	5.0 km Circle Level
Mmone/Quoxo	2324CD	Main	61-70	95.0%	31-40	95.0%
		Tribs	Small		Small	
	2324AB	Main	0-10	95.0%	0-10	95.0%
		Tribs	Small		Small	
	2224C	Main	61-70	95.0%	51-60	99.9%
		Tribs	11-20	95.0%	11-20	95.0%
					21-30	95.0%
	2224A	Main	0-10	99.0%	21-30	95.0%
		Tribs	11-20	99.9%	0-10	99.0%
					11-20	95.0%
Serorome	2325BD	Main	0-10	99.9%	0-10	99.9%
		Tribs	Small		Small	
	2326AC	Main	0-10	99.9%	0-10	99.9%
			11-20	99.0%		
		Tribs	0-10	99.9%	0-10	99.9%
		Tribs	Small		Small	

Additionally, the valleys of the Mmone/Quoxo and Serorome systems contain the only evidence for deep-weathering beneath valleys (table 7.7). A total of eleven boreholes contained evidence of bedrock weathering, most notably where boreholes penetrated Karoo basalt. The most common form of weathering within basalt was the alteration or removal of the bedrock, with solution hollows and fractures often infilled with calcite. Two boreholes drilled into Karoo sandstone and shale beneath Letlhakeng Valley 1 also showed evidence of deep-weathering, with borehole 4695 containing solution cavities and calcite infills and borehole 6514 penetrating an 8 m high cave (as detailed in section 6.2.2 *(c)*).

Comparison of the results of network orientation analysis in table 7.6 with table 7.7 is of potential significance in determining the mode of development of *mekgacha*, as this shows some coincidence of valleys exhibiting deep-weathering and close parallelism with structures. Whilst three boreholes (in QDS 2224C and 2324D) in the Mmone/Quoxo or Meratswe contained evidence for deep-weathering, these locations did not coincide with valley sections showing significant structural alignment. However, the remaining borehole locations in table 7.7 all coincide with valleys parallelling geological structures.

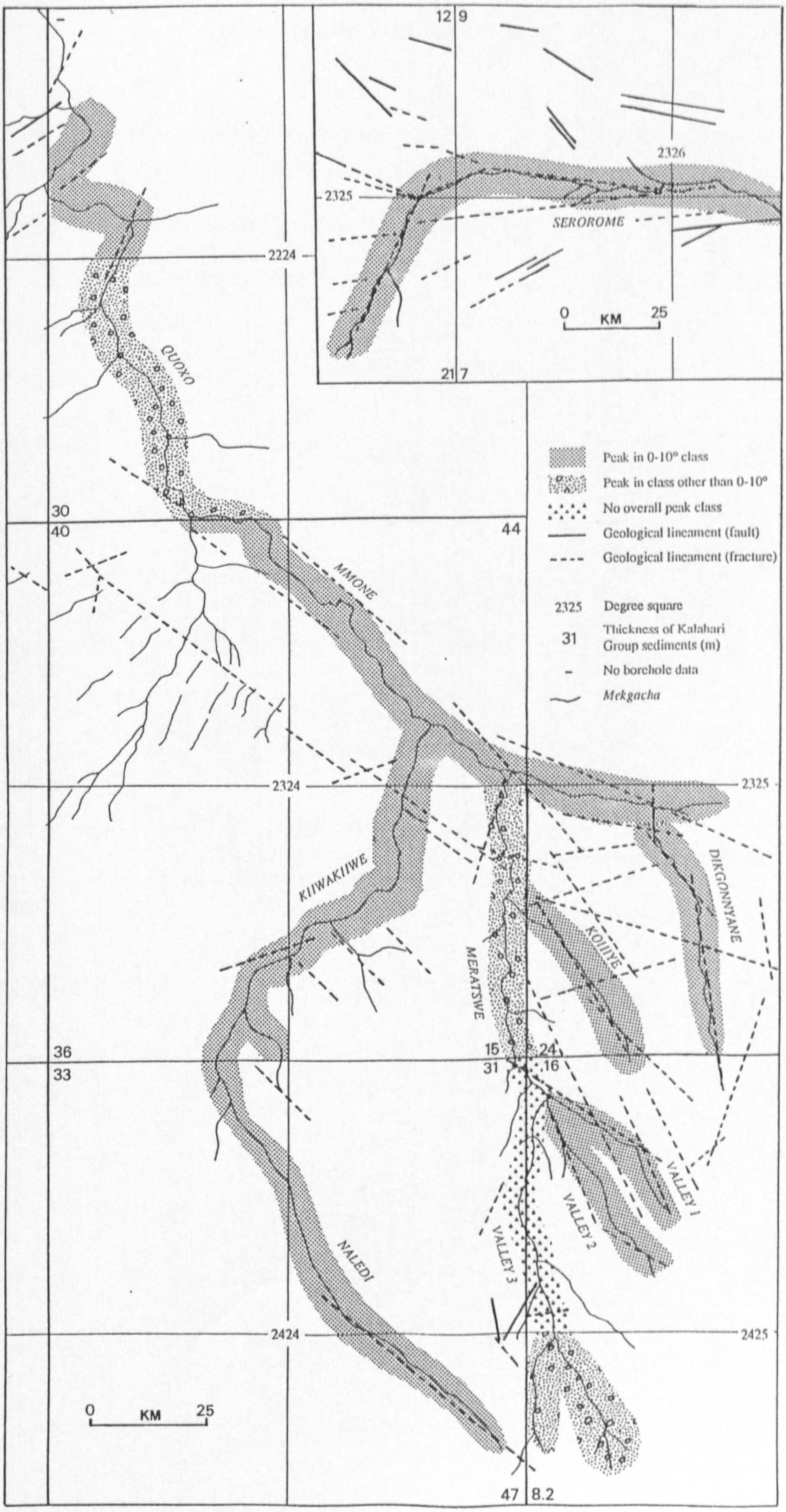


Figure 7.10: Results of network orientation analysis for individual valleys of the Quoxo/Mmone and Serorome valley systems.

Table 7.7: Evidence of deep-weathering from lithological borehole logs drilled in or adjacent to *mekgacha*.

Valley	Borehole no.	Grid reference	QDS	Evidence of deep-weathering
Mmone	Z 4374	22°40'S 24°10'E	2224C	Weathered amygdaloidal Karoo basalt.
Khwakhwe (at Tsia)	4007	23°40'S 24°40'E	2324D	Decomposed Karoo basalt at depths of 15-37 m.
Khwakhwe (at Tsia)	GS10/A6	23°40'S 24°40'E	2324D	Decomposed Karoo basalt with calcite infills between 16-33 m depth, immediately underlying basal Kalahari Group sediments.
Khwakhwe	GS10/A7	23°39'S 24°43'E	2324D	Decomposed Karoo basalt with calcite infilling voids and deposited within fractures.
Meratswe	717	23°50'S 24°56'E	2324D	Decomposed Karoo basalt at depths of 9-39 m.
Meratswe	Z 3813	23°35'S 24°55'E	2324D	Altered Karoo basalt between 29-53 m and 57-105 m depth.
Dikgonnyane	682	23°33'S 25°15'E	2325C	Decomposed Karoo basalt between 76-85 m.
Letlhakeng Valley 1	6514	24°10'S 25°10'E	2425A	8 m high cavity within weathered Karoo sandstone and shale.
Letlhakeng Valley 1	4695	24°15'S 25°15'E	2425A	Weathering and solutional cavities in Karoo sandstone to a depth of 125 m, with calcite infilling cavities in underlying dolerite and shale.
Serorome	Z 939	23°42'S 25°45'E	2325D	Weathered Karoo basalt between 48-58 m, with evidence of weathering and calcite cavity infill at depths of 106-121 m.
Serorome (tributary)	1883	23°40'S 25°48'E	2325D	Calcified Karoo basalt at depths between 39-182 m.

Boreholes GS10/A6, GS10/A7 and 4007 in the Khwakhwe Valley (QDS 2324D) coincide with a peak in the 0-10° class at a 95% significance level. Likewise, boreholes in the Scrorome Valley (QDS 2325D) and Letlhakeng Valley 1 (QDS 2425A) are in valley sections showing structural alignment at a 99.9% significance level. These results suggest a possible relationship between valley location and deep-weathering due to movement of groundwater along preferential flowpaths.

(e) Moselebe system

Of all the Kalahari valley systems analysed, the Moselebe is notable for containing no valleys with any evidence of structural control (figure 7.11 and table 7.8). Indeed, many valleys show no overall peak orientational class. This may be partly due to the low density of lineaments in the western part of the system in QDS 2423C/2523A and 2523B, although this is not the case in the eastern headwaters where an average lineament density occurs.

Table 7.8: Individual results of Network Orientation analysis by QDS for the valleys and tributaries of the Moselebe system. "No" indicates no statistically significant peak. "Small" indicates valley sampled less than nine times.

Valley Name	QDS	Main/ Trib	2.5 km Circle Peak°	2.5 km Circle Level	5.0 km Circle Peak°	5.0 km Circle Level
Moselebe	2524BD	Main Tribs	No Small		No Small	
	2524AC	Main	No Small		No Small	
	2523B	Main	No		31-40	99.0%
	2423C/ 2523A	Main	51-60	99.0%	41-50 61-70	99.0% 99.0%
Selokolela	2524BD	Main	Small		Small	
Sekhutlane	2524BD	Main	41-50	99.9%	31-40 41-50	95.0% 99.0%
	2524AC	Main	No		11-20	95.0%
Ukhwi	2524AC	Main	No		11-20	99.0%
	2523B	Main	51-60	99.9%	51-60	99.0%
Mabuasehube	2423C/ 2523A	Main	71-80	99.9%	61-70	99.9%

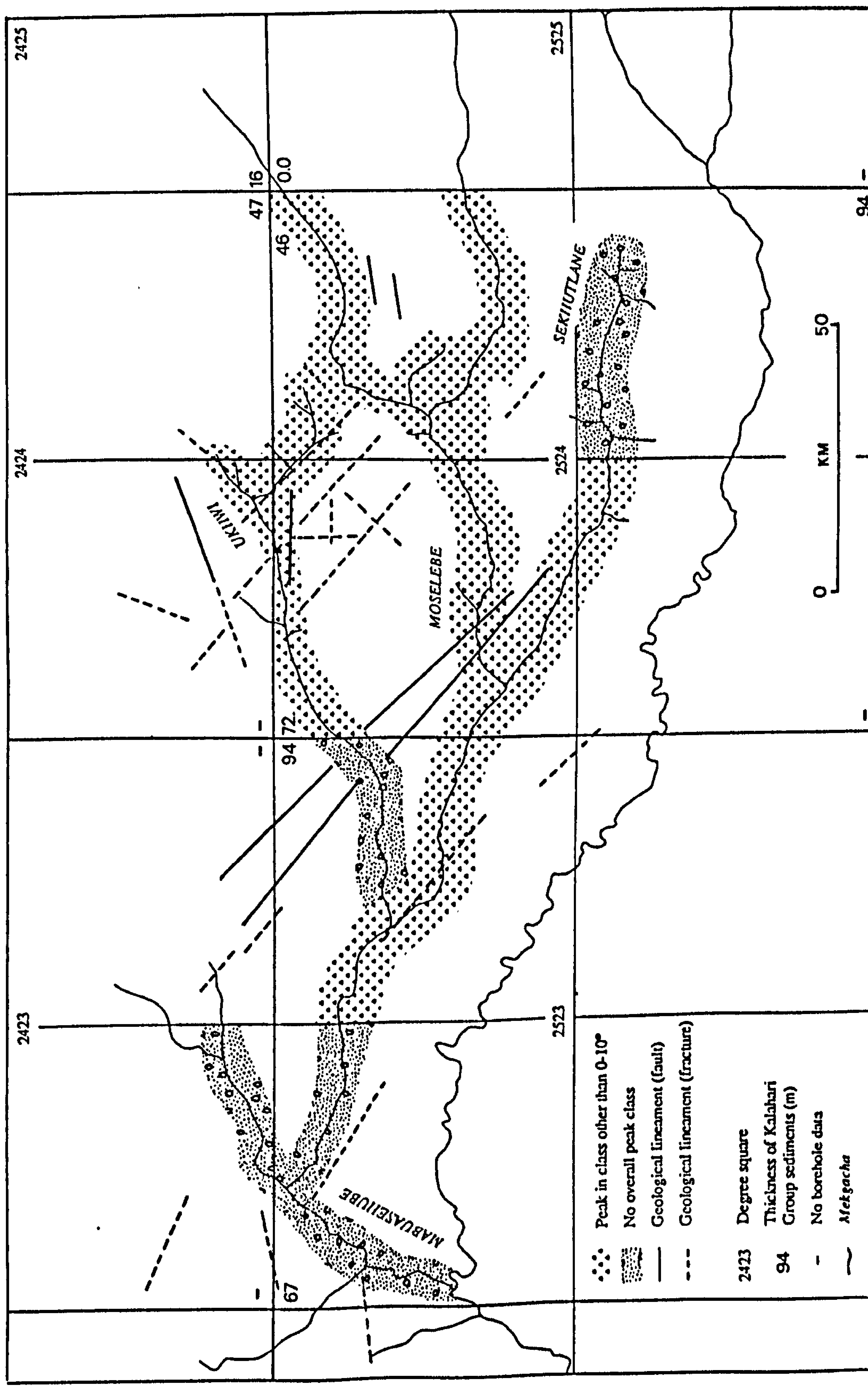


Figure 7.11: Results of network orientation analysis for individual valleys of the Moselebe valley system.

There is a considerable thickness of Kalahari Group sediments in the west of this region, (greatest value 183 m recorded in QDS 2523B) with lower values in the east toward the Kalahari edge. From these results it would appear unlikely that a relationship between valley and structural alignment can occur where thicknesses of Kalahari Group sediment overlying Basement rocks are high.

Unlike the Northern valley system where neotectonic activity has strongly influenced the structural grain of the region, the Moselebe system is comparatively undisturbed. This suggests that the thickness of Kalahari Group sediments is an important influencing factor.

7.2.3 The significance of sampling-circle size on the results of network orientation analysis

As described in section 7.1.3 above, two sizes of sampling-circle were utilised (with radii 2.5 km and 5.0 km) in the analysis of the relationship between valley alignment and geological structure. The underlying reason for the use of two sizes of circle was to assess the effects of scale on the results of the analysis; it would be expected that the different sized sampling areas would detect differing relationships between valleys and structure if the nature of the relationship were scale-dependent. During the course of the data collection, a number of variations in result produced by the different sampling areas were noted. These discrepancies seemed to occur mostly on or near bends or meanders in the valley, most noticeably where the amplitude of the bend fell between the diameters of the sampling-circles. Thus, some scale-dependent variations in the overall pattern of results might be expected if sampling-circle size was a significant factor in the analysis. If this were the case, it may be that any relationship demonstrated by the analysis were purely a function of the sampling method and not a result of a link between valleys and geological structure.

In order to assess the effect of using two sizes of sampling-circle, the results shown in figures 7.4 and 7.5 for each QDS sampling cell were used. These results show the spatial pattern of QDS cells exhibiting various relationships between valley alignment and geological structure. Viewed in terms of the research hypotheses set out in Chapter 4, the five categories shown in figures 7.4 and 7.5 can be effectively grouped into two broad classes depending upon their significance to the possible mode of development of Kalahari *mekgacha*. The classes are thus:

a) "Peak in 0-10° Class", amalgamating the results for the three significance values for "0 to 10° Peaks" shown on figures 7.4 and 7.5,

b) "No Peak/Peak in Other Class", combining the two remaining categories.

The results of this amalgamation of classes are shown on figure 7.12, with Black areas indicating cells which exhibit a "Peak in 0 to 10° Class" and Grey areas those on the "No Peak/Other Class" category.

Figure 7.12: Categorised results of network orientation analysis used in testing for spatial autocorrelation within *a.* 2.5 km sampling-circle and *b.* 5.0 km sampling-circle results.

In order to test for differences resulting from the use of two sampling strategies, a non-parametric McNemar test (after Siegel, 1956) was carried out on the spatial patterns shown in figure 7.12. The stages of this test are described below. However, prior to assessing any differences in the overall results, it was necessary to test the data shown in figure 7.12 for evidence of spatial autocorrelation. As with other non-parametric tests, a lack of autocorrelation is necessary for the validity of a McNemar test (as discussed by Haining, 1990).

Some interdependence between individual cells may be expected because of the nature of the sampling strategy (described in section 7.1.3). The reasons for this are twofold. First, many cells have some common areas at their margins in the vicinity of points where valley courses cross between cells. This was due to the decision to allow sampling-circles to overlap into neighbouring cells if the chosen sampling point fell close to the edge of a sampling cell. This only affects a very small part of the area of analysis. A second, potentially more important, factor leading to possible interdependence of the sampling cells is the fact that valley systems are continuous between cells. Thus, the results of the analysis in one cell may affect the result in a cell immediately adjacent to it, if the cells both contain parts of the same valley or structural element.

(a) Testing for spatial autocorrelation

Methods for assessing evidence of spatial autocorrelation and interdependence in data sets are discussed by Berry and Marble (1968), Shaw and Wheeler (1985), Upton and Fingleton (1985), Goodchild (1986) and Haining (1990) amongst others. To assess the results shown in figure 7.12, a Join-count test (using nearest neighbour "Rook" joins only) was carried out for both 2.5 km and 5.0 km sampling-circles. The methods described by Berry and Marble (1968) and Shaw and Wheeler (1985) were used.

(i) Join-Count test for 2.5 km sampling-circle results

From figure 7.12(a), the number of cells in each category were counted and the probability of a cell being Black (P) or Grey (Q) calculated;

Total number of cells	=		40	
Number of Black cells	=		23	
Number of Grey cells	=		17	
p(Black)	=	0.575	=	P
p(Grey)	=	0.425	=	Q

The pattern of results were then assessed by counting the numbers of adjacent Black-Black, Black-Grey and Grey-Grey cells;

Observed values of Join numbers:

Black-Black	= 20		
Black-Grey	= 23		
Grey-Grey	= 10		
Total (L)	$= \frac{1}{2} \Sigma_k L_k$	=	53
D	$= \Sigma_k L_k (L_k\text{-}1)$	=	200

Where L_k is the number of cells adjoining typical cell k

If the data contained no evidence of spatial autocorrelation, then the following values for the number of joins would be expected;

Expected values of Join numbers:

Black-Black	=	P^2L	=	17.52
Black-Grey	=	$2PQL$	=	25.90
Grey-Grey	=	Q^2L	=	9.58

In order to assess the difference between the observed and expected values of join numbers, the standard deviation associated with each expected value needed to be calculated;

Standard deviations associated with join numbers;

$$\sigma_{BB} = \{P^2L + P^3D - P^4(L+D)\}^{1/2} = 5.28$$

$$\sigma_{BG} = \{2PQL + PQD - 4P^2Q^2(L+D)\}^{1/2} = 3.78$$

$$\sigma_{GG} = \{Q^2L + Q^3D - Q^4(L+D)\}^{1/2} = 4.08$$

The standardised deviance (I) for each join type can then be calculated from the following formula;

$$I = (\text{Observed - Expected}) / \sigma$$

The results were as follows;

$I_{BB} = +0.47$, $I_{BG} = -0.77$, $I_{GG} = +0.11$

If the standardised deviance (I) falls within the range ± 1.96, then there is no evidence of pattern within the data. Positive or negative values of I indicate more or less joins respectively than would be expected if no pattern were present. From this, it can be seen that the pattern of results using the 2.5 km sampling-circle show no statistical evidence for spatial autocorrelation and therefore no evidence for interdependence.

(ii) Join-count test for 5.0 km Sampling-circle results.

The same procedure as above was carried out on the results for the 5.0 km sampling-circle results as shown in figure 7.12(b).

The results were as follows:

Total Number of cells			=	40
Number of Black cells			=	20
Number of Grey cells			=	20
p(Black)	=	0.500	=	P
p(Grey)	=	0.500	=	Q

Observed number of Joins;

Black-Black	=	18
Black-Grey	=	19
Grey-Grey	=	16
L	=	53
D	=	200

Expected numbers of Joins;

Black-Black	=	13.25
Black-Grey	=	26.50
Grey-Grey	=	13.25

Standard deviations;

$\sigma_{BB} = 4.74$, $\sigma_{BG} = 3.64$, $\sigma_{GG} = 4.74$

Standardised deviances;

$I_{BB} = +1.00$, $I_{BG} = 2.06$, $I_{GG} = +0.58$

The standardised deviances indicate significantly fewer Black-Grey joins than would be expected if no pattern were present. Despite the fact that neither the number of Black-Black or Grey-Grey joins is significant, these results suggest weak evidence for autocorrelation when using the larger 5.0 km sampling-circle.

(iii) Join-count Test for Differences between the two Sampling-circle sizes

The results for the two sampling-circle sizes (shown in figure 7.12) were manipulated by the following method to show changes between the two distributions of results. The resulting distribution is shown in figure 7.13. Cells which had changed from "Peak in 0 to 10°" to "No Peak/Other Peak" (or *vice versa*) were shaded Black on the figure, whilst those in which the result had remained unchanged for both size of sampling-circle were shaded Grey. A Join-count test was carried out on this data to assess any spatial autocorrelation, with the following results;

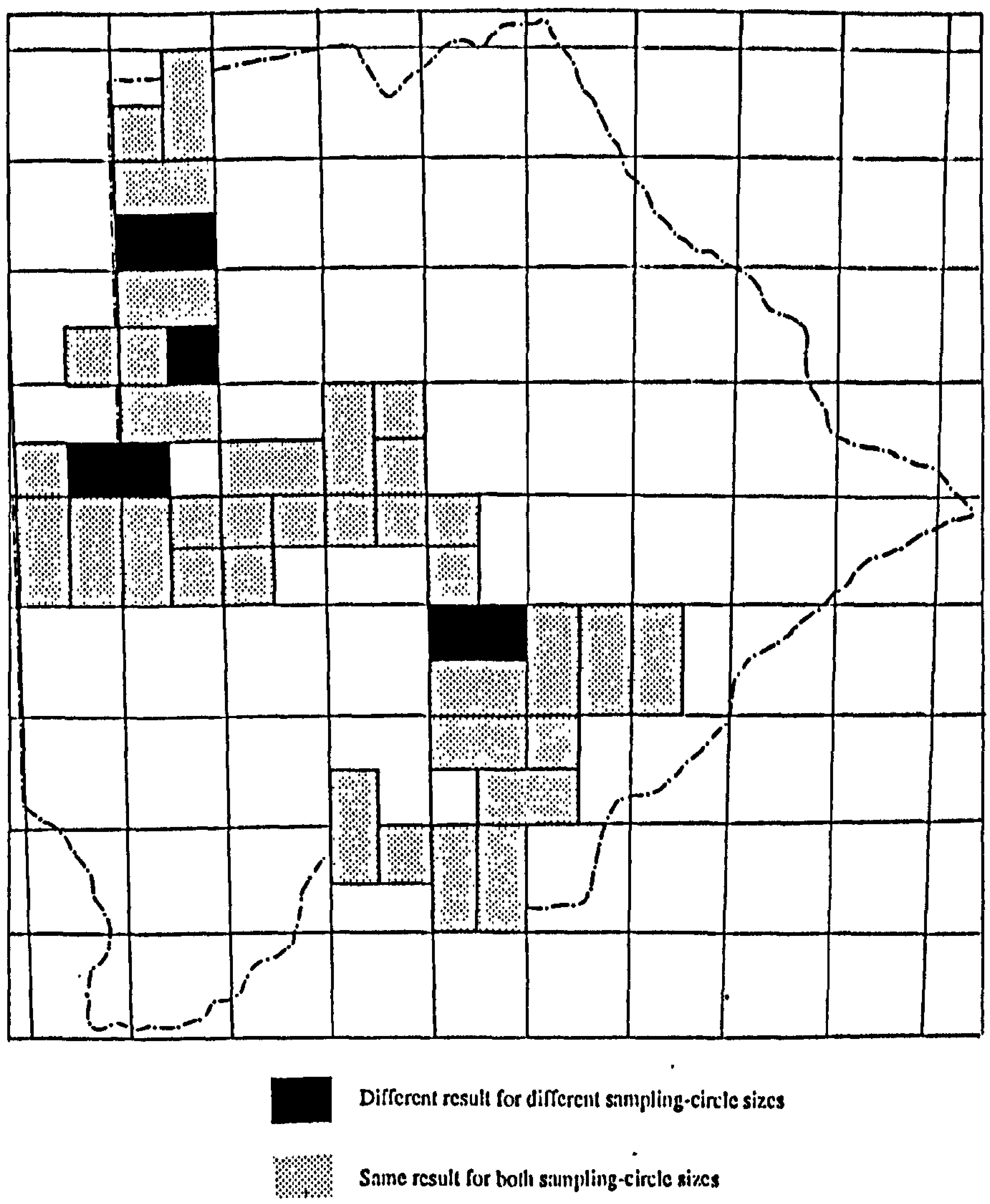


Figure 7.13: Differences in results when using two different sampling-circle sizes. Quarter degree squares giving different results for the two sampling-circle sizes are shown in black.

Total Number of cells			=	40
Number of Black cells			=	5
Number of Grey cells			=	35
p(Black)	=	0.125	=	P
p(Grey)	=	0.875	=	Q

Observed Numbers of Joins;

Black-Black	=	1
Black-Grey	=	11
Grey-Grey	=	41.
L	=	53
D	=	200

Expected Numbers of Joins;

Black-Black	=	0.828
Black-Grey	=	11.594
Grey-Grey	=	40.578

Standard deviations;

$\sigma_{BB} = 1.07$, $\sigma_{BG} = 4.62$, $\sigma_{GG} = 5.12$

Standardised deviances;

$I_{BB} = +0.16$, $I_{BG} = -0.13$, $I_{GG} = +0.08$

The standardised deviances demonstrate that although the results in the north-west area of figure 7.13 seem to show a clustering of cells showing changes, there is no spatial autocorrelation of the overall pattern of results (using nearest cell "Rook" joins only).

(b) The McNemar test to assess variations in results due to differences in the sampling methods

The pattern of results in figure 7.13 show that only five out of the total forty cells analysed have markedly different results for the two circle sizes used in sampling. To assess the significance of these variations, a non-parametric McNemar test (after Siegel, 1956) was undertaken. The data used satisfies all the criteria necessary for the use of the McNemar test; whilst the sampling cells are not all of the same size, they contain the same number of sample points at the same locations, but with a different sampling-circle size in the two cases. Thus, the test can be considered as a "Before and After" study with related samples, with the use of different sampling-circles substituted for the temporal role.

From the data in figure 7.12, a two by two contingency table showing changes in the distribution of results was compiled (table 7.9). The McNemar test result gives a Chi-squared value of 0.8 with 1 degree of freedom, which indicates no evidence for differences between the results produced using the two sampling-circle sizes. This indicates that the overall pattern of results is not significantly affected by the use of two different sampling-circle sizes.

Table 7.9: Contingency table for the McNemar test.

2.5 km Sampling-circle	0-10° Peak	4	19
	Other class/ no peak	16	1
		Other class/ no peak	0-10° Peak
		5.0 km Sampling-circle	

(c) Significance of results

The test for evidence of spatial autocorrelation within the results shown in figures 7.4, 7.5 and 7.12 resulted in there being no evidence for interdependence in the pattern for the 2.5 km sampling-circle, but weak interdependence with the 5.0 km sampling-circle. Had the evidence for spatial autocorrelation been stronger, this would have cast doubt upon the result of the McNemar test, since an absence of spatial interdependence is a pre-requisite for the test. However, as the outcome of the McNemar test yielded such a low Chi-squared value (0.8, when the critical value for a difference in the distributions at the 95% significance level is 3.84), the result is likely to be significant regardless of weak interdependence in one set of data.

This result indicates that the use of two different sizes of sampling-circle had no significant effect upon the overall pattern of results. Indeed, the fact that the two patterns are so similar would suggest that relationships between valley alignment and geological structure found for each sampling cell hold at the two scales of sampling used in this study. The fact that the pattern of results for the larger sampling-circle show evidence for spatial autocorrelation is not surprising, since any overlap of sampling areas would be more pronounced when using a larger circle. Also, since the total number of cells under analysis is relatively small (40 different sampling areas), any change in a cell classification in a "peripheral" setting is likely to significantly effect the pattern of clustering of overall results e.g. in the Northern valley systems.

In terms of the ideal choice of sampling-circle size, a small circle would be better in the analysis to avoid problems of spatial autocorrelation. However, as noted earlier, problems may arise using only a small circle if the nature of the relationship under consideration is scale-dependent and also if the density of available structural information is low.

7.3 Chapter summary

The findings of this chapter can be summarised in two parts: firstly, those results arising from the actual network orientation analysis, and secondly, a consideration of the methodology employed in carrying out the analysis.

(i) Viewing the results of the network orientation analysis for all valley networks, there appears to be no overall relationship between *mekgacha* alignment and geological structure. Whilst many systems do show significant alignment (e.g. the Okwa/Hanehai and parts of the Mmone/Quoxo and Serorome), there is no evidence for structural control in most other systems. From the analysis of Kalahari Group thicknesses, it would appear that the thickness of the sediment cover overlying Basement rocks affects any influence that geological structures may have. In general, a thick sediment cover (as in the Moselebe and parts of the northern Quoxo/Mmone systems) appears to be associated with a lack of structural alignment. Local factors and lithological variations make it difficult to attach a significant threshold figure for Kalahari Group thickness. However, it would appear that where a sediment cover in excess of 25-30 m occurs, there is unlikely to be alignment between valley and structural orientation.

Results from the Northern valley systems suggests that neotectonic activity associated with the Okavango graben appears to "overwrite" previous structural patterns. There is evidence from areas close to Precambrian bedrock outliers that drainage networks may have been controlled by structures formed prior to faulting in the Okavango Delta region.

The coincidence of valley segments in the Mmone/Quoxo system and Serorome showing strong structural alignment with evidence of deep-weathering (table 7.7) is possible support for a control upon valley location. It is, however, unclear why the basalts, sandstone and shales beneath the Mmone and Serorome valleys should show evidence of weathering and not the Precambrian bedrock beneath the Okwa Valley. This may be attributable to the differential weathering resistance of the different rock types, but may also be a result of the greater number of accurately logged boreholes in the southeasterly valleys.

(ii) In terms of methodology, the technique of network orientation analysis employed appeared to be generally successful in establishing broad scale patterns. At the scale of 1:250,000 used for the analysis, a sampling-circle size of 1 cm radius (2.5 km on the ground) was felt to be most appropriate. This size of circle has the advantage of avoiding effects due to spatial autocorrelation, but may not be appropriate for meandering valley systems. Using two sizes of sampling-circle would be recommended for further investigations, to take account of any spatial variations due to scale.

Part 3

Discussion and conclusions

Chapter 8
Discussion and conclusions

8.1 Introduction: scales of study

The three main methods of study utilised in the identification of the mode of development of Kalahari *mekgacha* cover a variety of scales. These range from the mega-scale study of the overall characteristics of valley form using network orientation analysis and inspection of remotely-sensed imagery to the micro-scale analysis of duricrusts in thin-section. At intermediate scales are field studies of valley form and the study of duricrusts in profile.

The results of these studies can also be considered at different scales, which can be separated into "valley" and "intra-valley" scales. The former group includes primarily large scale planemetric and morphometric characteristics of *mekgacha* and *mekgacha* networks, whilst the latter includes landform elements and deposits developed by sedimentary deposition and erosion within the setting provided by a valley.

8.2 Evidence for *mekgacha* development by fluvial processes

Evidence for the role of fluvial processes within *mekgacha* development is best exhibited by intra-valley forms, which mostly indicate the comparatively late stage involvement of flow in modifying superficial valley-floors deposits. This does not preclude earlier involvement, but evidence for this has either been reworked or destroyed by subsequent activity.

Sedimentary evidence of former flows is extensive, although the nature of the flow (i.e. perennial or ephemeral) which deposited sediments is generally difficult to discern. The majority of valley floors contain alluvial and/or aeolian sediments, indicating that *mekgacha* have not been fluvially inactive during their development but that windblown transport is also an important process of sedimentary fill. However, as noted in chapter 1, exposures providing good stratigraphic sedimentary sequences are very limited. The only *mekgacha* containing vertical exposures of fluvial deposits are the Okwa and Kuruman valleys, with absolute dating of carbonates and organic material within sediments in the Kuruman indicating a number of high-magnitude flood events during the Holocene (Shaw, Thomas and Nash, 1993). Without detailed analysis of the stratigraphy and grain surface characteristics of valley floor sediments in all *mekgacha*, it is difficult to assess whether deposits are primarily derived from fluvial or aeolian sources or a combination of the two (Thomas, 1988*b*). This is not assisted by the lack of boreholes containing accurately logged surficial deposits.

Other lines of sedimentary evidence for former flowing water within *mekgacha* include deltaic, gravel lag, shell and lignite deposits in many valleys. Deltaic deposits within the Makgadikgadi Basin associated with the Okwa (Cooke and Verstappen, 1984) cover an area in excess of 350 km^2, and formed below the 920 m lake level. The Groot Laagte delta covers an area of 4,600 km^2 adjacent and perpendicular to the older Okavango Delta alluvial deposits (Thomas and Shaw, 1991*a*). Borehole logs in the Xaudum Valley contain shell deposits near the base of valley fill sequences, which may indicate periodically flowing or standing water at various stages of development. Shallow-buried lignite deposits in these borehole logs also indicate standing water. Radiometrically dated shell deposits from the Xaudum and Okwa valleys indicate long-standing water at approximately 14,000 years BP with reed casts in the Okwa Gorge suggesting swampy conditions around 12,000 yrs BP. The dates from shell material can be considered to indicate a period of alluvial deposition which was followed by calcretisation prior to dissection into paired terraces. This period of incision may be represented in the Okwa by the age of the calcified reed casts. The presence of gravel lag deposits on the inside of a meander bend and signs of water erosion on bedrock outcrops within the Okwa Valley to the east of Tswaane borehole are also indicative of former flow.

Landforms such as terrace levels are also indicative of past fluvial erosion, although they require careful interpretation with regard to base-level changes. Terraces can be identified at various levels within four valley systems; at approximately 2.0 m and 4.0 m in the Okwa Valley, at 0.8 to 1.0 m in the Xaudum, at 2.5 to 3.0 m and 7.5 to 8.0 m in the Kuruman and at approximately 2.5 m in the Moselebe. Anastomosing and meandering channel patterns are present in the Ncamasere and Xaudum valleys, and also in the Auob and Nossop. In the case of the Auob Valley, studies of duricrusts suggest that the valley has incised through a sequence of pre-existing calcretes and silcretes. Evidence of the recent action of water can be seen in both the Kuruman Valley, where a flood in 1988 modified the channel, and the Nossop Valley where linear ridges formed during the 1933-34 flood are identifiable in the field and from aerial photography.

8.3 Evidence for the role of groundwater processes in *mekgacha* development

Evidence for the general role of groundwater processes in Kalahari valley development is less distinct and perhaps more complex than the unequivocal indication of the action of perennial or ephemeral fluvial activity outlined above, although some valleys do show clear signs of formation by groundwater processes. This appears to be primarily a result of groundwater processes operating much more subtly, at slower rates and over longer periods than fluvial erosion. The role of groundwater appears to be primarily manifest at a "valley" scale, as opposed to fluvial activity which mainly modifies intra-valley sediments and landforms.

Deep-weathering processes appear to have been most important in the control of valley location and orientation in particular *mekgacha* networks, primarily the Okwa, Mmone/Quoxo, Serorome and Deception valleys. The orientation of many systems appears to be controlled by geological structures developed in bedrock now buried by Kalahari Group sediments in excess of 20 m thick. Indication of structural control is, in many cases, coincident with evidence for extensive weathering and replacement of bedrock at depth beneath Kalahari Group sediments, usually associated with fracture zones. This suggests that, particularly in the Mmone/Quoxo and Serorome systems, deep-weathering due to circulation of groundwater along preferential flowpaths has been a factor in valley development.

The lack of evidence for deep-weathering from other networks does not necessarily preclude the role of groundwater processes in the development of these valleys. It may merely indicate the lack of sufficiently detailed lithological borehole logs in valleys away from the southeastern quarter of Botswana. The valleys in the sparsely populated areas to the west of the Okavango Delta show little evidence for structural control but also contain few detailed borehole logs. As such, it is not possible to evaluate the role of groundwater processes in their development. In order to ascertain whether deep-weathering is only associated with lineaments, a study of lithological logs from boreholes located on geological lineaments not associated with valleys would be required. However, until detailed data enabling the correlation of exact borehole locations and lineament zones are available for the whole Kalahari (such data are only available for the eastern hardveld) this type of analysis is not possible. As noted in chapter 3, many aquifers and the majority of groundwater recharge within non-porous bedrock in the Kalahari are generally confined to fracture zones (Republic of Botswana/VIAK, 1984*c*; Gieske and Selaolo, 1988). As such, it would be expected that extensive deep-weathering could only occur in association with lineament zones in these lithologies. The borehole logs associated with valleys which did contain evidence for deep-weathering were from boreholes drilled into either sandstones and shales, where groundwater was restricted to zones of secondary permeability, or basalts from the Karoo Sequence. This may indicate preferential weathering of different lithologies, but without more detailed evidence this suggestion is purely conjectural.

A more certain indication of the circulation of groundwater during the early stages of valley development is provided by studies of duricrusts, particularly in thin-section. The calcretes in valley flanks in the Letlhakeng area contain bivalve shell material, indicating deposition of the host material by flowing water, possibly in a proto-valley. The morphology of these calcretes further indicates that formation and later silicification took place beneath a valley-floor water table, possibly when permanent or semi-permanent water was present. However, the timing of calcretisation relative to the deposition of sediment and shell material cannot be easily determined. The presence of shell material within calcretes now exposed in valley flanks, some distance above the *mekgacha* floor means that a period (or periods) of incision took place after their deposition and transformation of the host material into a calcrete, thus isolating the calcretised sediment as hillside exposures. As such, the formation of duricrusts was intrinsically linked to the presence of valleys and valleys have not (with the exception of the Auob and possibly the Nossop valleys) incised through a pre-existing sequence of duricrusts. This casts doubt upon many of the lithostratigraphic attempts at correlating the Kalahari Group sediments based upon exposures in *mekgacha* (which often provide the only surface exposures in many areas). Neither have duricrusts been modified by, or developed in association with, other conduits of water movement such as fractures or lineaments. The distribution of duricrusts on the 1:250,000 interpretations of Landsat imagery by Mallick *et al.* (1981) show no preferred development of duricrusts in association with lineaments away from valleys.

Further indication of the role of groundwater in duricrust formation is provided by the sequences of silicification identified in many calcretes, which are analogous to examples from studies of groundwater calcretes in central Australia (Arakel *et al.*, 1989). The extent of alteration and the multiple sequences of silicification within valley flank duricrusts in the Letlhakeng area highlight the importance of diagenesis in conjunction with fluctuating valley water tables. The sequences of silica and calcite void fill identified from studies of thin-sections implies shifts in groundwater chemistry. It is also likely that water existed beneath valley floors for a long time during diagenesis, particularly where megaquartz is present at the centre of voids, although this may also be attributable to the progressive restriction of porewater flow within infilling voids. There is some indication that silicification may have occurred under conditions of a lowering water table, probably concomitant with valley incision.

The lateral and vertical movement of groundwater beneath valleys is further indicated by borehole logs showing duricrust types varying in a direction perpendicular to the valley axis. In particular, the Rooibrak and Letlhakeng Valley 2 show variations in calcrete lithology and thickness away from the valley axis. This again supports the proposition that the formation and development of duricrusts and duricrust suites are inseparable from valley evolution.

The control of valley orientation and location by deep-weathering and the formation and alteration of duricrusts by groundwater movements indicate the potentially complex interrelationship between groundwater and fluvial activity. Evidence of structural control due to structures now buried by Kalahari Group sediments suggests the predominance of weathering due to preferential flow of subsurface water. Such subsurface flow would lead to gradual valley development by the removal of material in solution, as per the model proposed by McFarlane (1989) for dambo formation (chapter 2). However, the evidence from silicification and alteration of duricrusts suggests that surface water may have existed at times to maintain water tables beneath the valley. If silicification and diagenesis occurred in a similar environment to the studies of Australian groundwater calcretes, then water tables would probably have been at depths of less than 10 m below the valley floor during the period(s) of diagenesis. Water tables in the Australian studies are at depths of approximately 6 m (Arakel *et al.*, 1989), with silicification apparently associated with the presence of seasonal water supplies from sources such as playa lakes. In order for the diagenetic alteration of Kalahari duricrusts to have proceeded, this implies that either exchanges of water between surface and groundwater sources have occurred, as per the Australian studies, or that the presence of a topographic low provided by a valley has led to a lateral transfer of groundwater to maintain relatively shallow water tables within both the valley-floor and hillslope systems. A possible method for identifying the extent of any exchange between meteoric and groundwater supplies may be through the use of carbon and oxygen isotope analyses (Wright and Tucker, 1991; Avigour *et al.*, 1992).

Discriminant analysis of silcrete bulk chemistry data shows a clear distinction between Cape Coastal samples (analysed by Summerfield, 1983*d*) and Kalahari silcretes on the basis of TiO_2 content, statistically verifying the qualitative proposition made by Summerfield (1983*a*). Summerfield suggests that this difference may be due to Cape silcretes developing under conditions of intensive weathering. On the basis of this geochemical criteria (particularly low TiO_2 levels), the silcretes included in this study did not develop under deep-weathering conditions. However, petrographic evidence indicates that differences between Kalahari and Cape silcretes may be largely attributable to differences in silcrete host material. Studies of duricrusts in thin-section combined with analyses of major element bulk chemistry show that silcretes associated with *mekgacha* developed either by the replacement of pre-existing valley calcretes or by primary precipitation of silica in unconsolidated sands. As such, whilst the use of silcrete TiO_2 levels as palaeoenvironmental indicators may be acceptable in Cape silcretes, the geochemical criteria upon which climatic suggestions are based should not be applied to the Kalahari. Low TiO_2 contents in Kalahari silcretes do not, as such, preclude the role of deep-weathering processes in Kalahari *mekgacha* development.

In contrast to the widespread evidence of valley locational control by deep-weathering processes, few locations contain direct evidence for valley development by groundwater sapping. The most probable relict sapping site is the amphitheatre valley head of Letlhakeng Valley 1, where spring lines are present at the base of silcrete cliffs. The spring lines are also associated with zones of sub-horizontal tubular structures, possibly indicative of a zone of former groundwater emergence. Whilst studies of silcrete duricrusts at the valley head do not show clear evidence for the movement of silica-rich groundwaters towards the spring line, they do indicate that thicknesses of silcrete are limited in spatial extent to this area. The silcrete at the valley head does not appear to have developed by alteration of pre-existing bedrock or duricrust, but by the cementation of an unconsolidated sand host material. This cementation would require the incursion of silica-bearing groundwater towards the valley head area, which suggests that the valley may have acted as a focus for groundwater for a considerable time.

For groundwater to converge on a particular location requires either a suitable hydrostatic gradient or forced convergence. If the former were the case in Letlhakeng Valley 1, this would suggest that some form of depression or shallow valley existed at the early stages of valley development to provide the necessary hydrostatic gradient. This would be consistent with the presence of a proto-valley in which the bivalves now included within calcretes in Letlhakeng Valley 2 may have been deposited. The valley location may also have been provided by the former channel(s) from which pebbles now incorporated within conglomeratic duricrusts were derived. These pebbles, although now in a matrix substantially altered due to diagenesis in association with the valley, are part of the basal Kalahari Group and were deposited by ancient drainages, possibly as early as the Jurassic (Thomas and Shaw, 1991*a*). This fact, together with the presence of buried channels beneath parts of the Moselebe system, may indicate that many *mekgacha* are developed within exhumed valleys and that their courses, if not the present valleys, are of considerable antiquity. However, there are numerous examples where geophysical techniques identify buried channels trending obliquely beneath present-day *mekgacha* courses (e.g. the Nossop near Aranos; McDaid, 1985); why one valley would occupy an exhumed channel whereas another would cut obliquely above it is uncertain. It is possible that the combined influence of an aquifer provided by buried channel deposits and the presence of preferential flow along subsurface lineaments has controlled the formation of Letlhakeng Valley 1.

Forced groundwater convergence could be attributed to a number of factors, including the lithological changes identified by Shaw and De Vries (1988) under which the presence of clay-rich shales is suggested to reduce permeability and affect groundwater circulation. There is also evidence of subsurface flow along fractures aligned parallel with the valley, which suggests preferential groundwater flowpaths. The rise in pre-Kalahari topography to the southeast of the valley head suggested by borehole records may have forced the "channelled" groundwater towards the surface at this point.

Away from Letlhakeng Valley 1, Thomas and Shaw (1991*a*) suggested that the abrupt change in valley form from flat headwater areas to a more incised section exhibited by many *mekgacha* may be evidence for the role of sapping processes in valley development. Silcretes are commonly found in "nickpoint" locations, in particular in Letlhakeng Valleys 2 and 3, in the Serorome Valley and in parts of the Moselebe system. If the model of silcrete formation by converging groundwater is correct for Letlhakeng Valley 1, then the association of siliceous duricrusts with other valley nickpoints may indicate the possible convergence of groundwater at specific sites. This would not necessarily identify sapping processes as the primary mode of valley development, since other processes may have operated after the formation of the silcrete. For example, it would be equally possible to argue that the highly indurated silcrete has slowed the headward recession of a migrating nickpoint resulting from the lowering of a regional base level, with the nickpoints preserved after the cessation of fluvial activity. The presence of a spring line at the head of Valley 1 indicates otherwise for this particular valley, but nickpoint recession may be an appropriate for changes in valley form in other locations where evidence to the contrary is lacking.

A further problem exists regarding the suggestion by Shaw and De Vries (1988) that the presence of particular silcrete types indicates the role of groundwater in valley development. As noted in chapter 2, valley development by groundwater sapping processes proceeds by gradual headward extension due to erosion in seepage sites at the valley head. If the silcretes are intrinsically associated with groundwater emergence at spring lines, then they should be present along the entire course of the valley, left on the valley sides and floor as relict deposits indicating the former positions of the valley head. This does not appear to be the case in any of the *mekgacha* studied in the field. It is, however, possible that conditions suitable for silica precipitation were not present during the course of headward erosion, thus precluding the formation of silcretes except under particular circumstances. The observation made by Ollier (1991*a*) should be remembered in this context, namely that two environments need to be considered when accounting for the presence of a duricrust; the locations where dissolution and precipitation took place. Thus, even if silica was available in solution, unless conditions suitable for precipitation were available, silcrete formation could not occur. One possible explanation for the presence of silcrete at the valley head is that a general shift in the environmental groundwater pH occurred, allowing the increased solution of silica and hence silica saturation in porewaters, with localised precipitation due to fluctuations in porewater chemistry. Such a shift may also be implied by the evidence of extensive silicification in calcretes at Letlhakeng.

8.4 Spatial variations in valley-forming processes

The preceding discussion indicates the considerable spatial variability in the relative importance of fluvial and groundwater processes in *mekgacha* development. Of the systems investigated, all exoreic valleys (except the Serorome) contained evidence of fluvial activity, with the endoreic Xaudum, Okwa and Moselebe systems also containing landform elements and sediments indicative of fluvial erosion. The remaining systems contained little or no indication for former flows apart from localised gully development in the Letlhakeng valleys.

The prevalence of fluvial activity in the exoreic systems can be attributed to their source areas. In the case of the Auob and Nossop, both valleys rise in highland areas near Windhoek in Namibia and flow into the Kalahari. Likewise, both the Kuruman and Molopo rise in areas beyond the Kalahari, both having spring sources in the Northern Cape Province of South Africa. Many of the endoreic systems also rise on bedrock, but when contemporary flows do occur in headwater areas, they rarely extend far into the parts of *mekgacha* covered by Kalahari Group sediments before being absorbed into the sedimentary infill.

The evidence for the role of groundwater in *mekgacha* development is less spatially extensive, and on the basis of evidence from network orientation analysis can be isolated to the Mmone/Quoxo and Okwa valley systems, with part of the Deception Valley also controlled by buried structures. Valley networks resulting from sapping processes commonly have a low drainage density. Drainage density was not calculated for individual Kalahari *mekgacha* networks since the true extent of the valley drainage net, and in particular headwaters, cannot be reliably identified beneath the cover of Kalahari Sand. As an approximate figure, the combined Okwa-Mmone system (from 1:250,000 topographic sheets) has a total valley length of at least 2,370 km in a potential catchment of 90,000 km^2, and hence a drainage density of less than 0.26 km of channel km^{-2}. With the future use of remotely-sensed imagery to assess headwater extension beneath the sand cover, this calculation may be possible. At a valley scale, the only *mekgacha* to fit most of the criteria indicative of network development by sapping processes (section 2.3.1*(iv)*) is Letlhakeng Valley 1. In addition to the abrupt valley initiation, alcove development and possible spring sites, Valley 1 has steep walls, a flat floor with evidence of mass wasting, and a long straight main valley with few short tributaries. In contrast, the exoreic systems (again excepting the Serorome) do not appear to show evidence of structural alignment, although in the absence of network orientation data for these valleys this suggestion is purely qualitative. This indicates that no one factor can be identified to account for the formation of all *mekgacha*, and development appears to have occurred by different processes operating in different combinations at various locations. Whilst it is tempting to suggest that the development of exoreic systems has been dominated by fluvial activity, further evidence is required before that statement can be asserted.

8.5 Timescales of *mekgacha* development

In addition to acting at different spatial scales and in varying combinations, fluvial activity and groundwater erosion appear to have operated at distinctly different timescales. Deep-weathering and possibly groundwater sapping processes may have been largely responsible for the location of *mekgacha* and the morphology of many networks, with deep-weathering providing zones of weakness which were subsequently exploited by fluvial activity. However, fluvial activity appears to have been more responsible for intra-valley forms and sediments, and it is likely that flowing water has been a major factor in shaping valley form in many systems. Certainly, the presence of sedimentary infill in most *mekgacha* indicates that valleys have not been fluvially inactive.

Many Kalahari *mekgacha* are clearly of great antiquity, although it is not possible to identify exact ages for any network. Evidence of the probable age of valleys is provided by three systems. Firstly, the presence of Dwyka tillite erratics not buried by other deposits in the Nossop Valley to the south of Gobabis (Hegenberger and Seeger, 1980) indicate that this headwater section of the Nossop has been in existence since the end of the Permo-Carboniferous glaciation. Secondly, the presence of shell beds at a depth of 50 m in a north-northwesterly trending graben which cuts the Xaudum Valley suggest possible permanent water prior to the initiation of this graben. The Okavango "Panhandle" occupies a graben of similar orientation which can be traced as gravity anomalies beneath alluvium in the main Okavango Rift (Mallick *et al.*, 1981), and thus predates rifting. The rift structure occupied by the Okavango Delta existed prior to the division of Gondwanaland (Rust, 1975), which suggests a maximum Late Palaeozoic age for these shells. Thirdly, the control of valleys in the Aha Hills by structures which predate the northeast-trending faults associated with the Okavango Delta may indicate a great age. This can only be regarded as a tentative suggestion since contemporaneous tectonic activity associated with the Okavango Rift also occurs.

There is a general concensus that duricrusts associated with the sides of *mekgacha* are old, although no authors suggest such antiquity as indicated by the examples in the preceding paragraph. Netterberg (1969*a*) and Goudie (1973*a*) both propose a Pliocene age, Boocock and Van Straten (1962) relate duricrusts to the "African" erosion cycle which ended in the early Miocene (Dixey, 1958*b*), whilst Wright (1978) notes formation of duricrusts in the Xaudum predating tectonic adjustment. However, this does not necessarily suggest that all duricrusts associated with *mekgacha* are ancient; the calcretes from terraces within the Okwa Valley are evidence of relatively recent formation, and diagenetic alteration in calcretes from Letlhakeng indicates that formation has occurred over a long time period, not merely restricted to one time period. Establishing the timing of duricrust formation requires the application of dating techniques such as ESR to samples, but with dates obtained for the different microscopic layers of void fill, not for entire crushed samples.

The relationship between *mekgacha* and landforms which have responded to Quaternary environmental changes provide a useful indication of more recent activity within valley networks. For example, many of the valleys to the west of the Okavango Delta (e.g. the Xaudum and its tributaries) cut through linear dune fields, but also have linear tributaries developed within interdune *straats*. This suggests that fluvial activity within these valleys has occurred since the last period of dune extension, but does not necessarily indicate that valleys have formed since that time. In contrast, the Okwa and Mmone valleys both contain extensive infills of aeolian sediments which choke their channels at their down-valley ends, indicative of aeolian activity since the last fluvial event.

Radiometric dates for shell deposits within the Xaudum and Okwa valleys indicate fluvial activity at around 14,000 years BP, which was followed by a period of calcrete formation and subsequent incision to create terrace landforms. The deltaic deposits associated with the Okwa have formed below the 920 m shoreline and probably date to these periods of fluvial activity. However, this relates to only the lower terrace level within the Okwa; the timing of the formation of the upper terrace will remain unknown until other radiometric dating techniques can be applied to older silcretes and calcretes.

As noted in chapter 1, valley development over such long time periods must be viewed within its tectonic context. Geological structures acting as preferential groundwater flowpaths beneath the Kalahari Group sediments must predate these sediments, and are thus at least Mesozoic in age (Thomas and Shaw, 1991*a*), although many lineaments may have been reactivated. Superimposed upon this ancient fault pattern in the northern Kalahari are fractures resulting from neotectonic activity in the Okavango Rift and Panhandle regions. In addition, uplift along the Kalahari-Zimbabwe Axis has occurred since the mid-Tertiary (Du Toit, 1933) and has caused changes in land elevation, particularly along the eastern Kalahari margin.

The influence of these tectonic movements has had a major impact upon drainage patterns in southern Africa and has been implicated as a major factor both controlling and hampering the reconstruction of lake levels in the Makgadikgadi Depression. The effects of changing base levels upon fluvial activity within *mekgacha* are indicated by the presence of terrace levels in many systems, although sediment supplies and climatic changes also need to be considered in the interpretation of these features. Uplift along the Kalahari-Zimbabwe Axis is perhaps of greatest importance because of the impact it would have had upon regional water tables. As has been noted, higher water tables than the present day would be required in order for sapping processes to operate in the Kalahari. It is possible that regional uplift along the Kalahari-Zimbabwe Axis once provided increased hydrostatic gradients, encouraging greater flow along sub-surface lineaments. This would have greatest impact upon valleys such as Letlhakeng Valley 1 developing above fractures oriented approximately perpendicular to the uplift axis. It is also possible that inputs of groundwater from regions beyond the Kalahari assisted in maintaining higher water table levels. The depth of present day water tables is a combination of a lack of significant widespread recharge since 12,500 years BP (De Vries, 1984) and depletion due to anthropogenic effects (Thomas and Shaw, 1991*a*). The fact that contemporary water tables are at great depths below many valley floors does not indicate that this has always been the case, and climates would not necessarily have to have been considerably wetter compared to the present day to produce higher water tables if regional groundwater transfers and tectonic changes have affected present levels.

The regional influence of uplift along the Kalahari-Zimbabwe Axis may explain why evidence for the role of groundwater sapping and deep-weathering is mainly confined to headwater regions in the Mmone/Quoxo system. These headwater areas would be most affected by any regional uplift, not only in terms of the impact upon water tables but also the effect upon valley gradients. Although not investigated in the field, a similar scenario may be envisaged for the valleys to the west of the Okavango Delta but with the change in regional gradients provided by downwarping in the Okavango Rift.

In general terms, the influences upon valley development can be viewed at three scales. Firstly, in the long term (10^7 to 10^8 years) geological structural lineaments are likely to have influenced valley location in many systems by deep-weathering processes, although parts of valleys such as the Nossop may be located within ancient valleys. At an intermediate scale, climate and neotectonic changes together with drainage capture may have acted upon valley location, gradient and the height of water tables, whilst comparatively recent (10^1 to 10^5 years) climatic and neotectonic changes, including the action of contemporary floods, have influenced intra-valley forms.

8.6 Summary conclusions

In summary, the most pertinent points which have been raised regarding the development and environmental significance of Kalahari *mekgacha* are outlined below.

1. It was recognised that two main groups of processes are important for the development of world valley systems, as opposed to the concensus opinion that most valleys form as a result of erosion due to fluvial activity. In addition to fluvial erosion by either ephemeral or perennial flow, groundwater processes (sapping and deep-weathering) have been identified as major, if poorly understood, mechanisms of valley formation. These processes are not mutually exclusive, and groundwater and fluvial activity should be regarded as end-members of a process spectrum.

2. From a methodological viewpoint, this study considers *mekgacha* evolution from a mainly geological (but also geomorphological) perspective. Two other possible approaches were recognised; a palaeohydrological approach utilising evidence from sedimentary sequences and morphological data, and an investigation of changes in network density, structure and composition. These were not adopted due to practical reasons, in particular the lack of suitable exposures within *mekgacha*, and the presence of Kalahari Sand infill in valley headwaters which makes accurate network delimitation impossible. More importantly, these alternative approaches consider only channel and sedimentary, not valley, environments. Finally and fundamentally, neither approach enables the recognition of the role of groundwater proceeses in valley development.

3. The technique of network orientation analysis employed to identify the relationship between structure and valley alignment was largely successful at establishing broad scale patterns. The use of a sampling circle of 1 cm radius (2.5 km on the ground) was found to be most appropriate for the study of valleys, but may not be so in meandering channel systems where meander wavelengths are similar to the sampling circle diameter.

4. The two main sub-groups of Kalahari *mekgacha* identified by Thomas and Shaw (1991*a*) were used as a basis for study; exoreic systems directed towards the Atlantic Ocean via the Orange River, and endoreic drainage towards the interior Okavango Delta and Makgadikgadi Depression. Field investigations and analyses of remotely-sensed imagery indicate the great variability in the form of Kalahari *mekgacha*, both within and between different systems. Studies indicate that the morphological characteristics of *mekgacha* networks identified by earlier studies are generally correct, although many systems show major deviations. There is a major difference in valley morphology between endoreic and exoreic *mekgacha*, with the latter systems exhibiting a much more clearly defined form. This is attributed to the occurrence of contemporary flow within externally directed systems.

5. Field studies indicate the role of fluvial activity in shaping valley floor sediments and morphology. Evidence includes paired terrace levels in the Okwa, Xaudum, Kuruman and Moselebe systems, anastomosing channel patterns, abandoned meanders and channel lag deposits. These features have developed within pre-existing valleys, and hence indicate fluvial activity in *mekgacha* after the formation of main valley morphology. This does not, however, preclude the role of fluvial activity at an earlier date; studies of borehole logs and duricrusts in thin-section also indicate the role of flowing water in *mekgacha* at earlier stages of development.

6. Radiometric dates of shell material within terrace calcretes and valley floor sediments provide the first indication for the timing of former flow within the Xaudum and Okwa valleys. Dates indicate the probable deposition of shell material at approximately 14,000 years BP, indicating comparatively recent periods of flow. Previous dates from *mekgacha* have been restricted to the Dobe, Gcwihabe and Nossop valleys. Evidence suggests that headwater sections of the Nossop have existed since the Permo-Carboniferous. Shell deposits in the Xaudum occupy a graben structure which may predate rifting in the Okavango Delta region, which itself existed prior to the breakup of Gondwanaland.

7. Evidence for the action of groundwater in *mekgacha* is primarily indicated by larger scale morphological features of networks. The relict springlines at the valley head of Letlhakeng Valley 1 suggest development by groundwater sapping processes although more recent fluvial activity has eroded the back wall of the amphitheatre head. The valley also exhibits steep sides, a flat floor, alcove development, has a very straight course and few tributaries, features identified as indicative of development by sapping by Howard *et al.* (1988).

8. Investigation of the relationship between valley and geological structural orientations indicates a strong structural control of many valley networks, particularly the Okwa, Mmone/Quoxo, Deception and Serorome systems, although within these systems there is spatial variability in the degree of

valley and structural alignment. The degree of structural control appears to be related to the depth to which pre-Kalahari bedrock is buried beneath Kalahari Group sediments. Thus the lack of structural control in the Moselebe system is attributed to the thickness of Kalahari Group sediments within which the valley is developed.

9. Deep-weathering within fractures zones can be identified from boreholes drilled in or adjacent to structurally controlled valley segments. This indicates subsurface flow along preferential flowpaths to depths of over 125 m beneath valley floors. The presence of calcite infills at depth also indicates that groundwaters are (or were) rich in solutes. These factors implicate the role of deep-weathering processes in the control of valley location and development.

10. Valley systems to the west of the Okavango Delta show little evidence of alignment with structures, except near areas of bedrock outcrop such as the Aha Hills. This suggests that the valleys developed prior to the emplacement of the Kalahari Group sediments and are controlled by old structures not visible at the surface. Neotectonic activity associated with rifting in the Okavango Delta region has strongly influenced fracture patterns (and hence the lineaments included in network orientation analysis) and appears to have overwritten previous structures.

11. With the exception of the Auob and possibly the Nossop Valley, the majority of duricrust exposures on the floors and flanks of *mekgacha* appear to be genetically related to the presence of a valley or depression. This is supported by;

    a) Results of studies of duricrust profiles, which indicate that duricrust morphology varies both along and perpendicular to valley axes, except in the Auob where the valley has incised through a pre-existing duricrust stratigraphy.

    b) Studies of lithological borehole logs from areas with boreholes sited along transects perpendicular to the valley axis, which show duricrust type and thickness varying with distance from the valley axis.

    c) Thin section analyses of duricrusts which exhibit both textures and extensive diagenetic alteration, particularly in calcretes, which have developed in association with a valley water table.

    d) The presence of shells in many valley flank duricrust samples from the Letlhakeng area, indicating deposition of the calcrete host material by flowing water.

12. Attempts to lithostratigraphically correlate the Kalahari Group sediments on the basis of duricrust exposures within *mekgacha* will be hampered by the variability in samples because of the close link between duricrusts and valleys. Whilst lithostratigraphic relationships can be identified from exposures in the Auob and Nossop valleys and from borehole logs, such techniques should not be applied to surface exposures in endoreic *mekgacha* unless only the host material is considered in a stratigraphic context.

13. Geochemical and thin-section analyses of silcretes from the Letlhakeng area and from the Okwa Valley suggest that many silcretes associated with *mekgacha* developed by cementation of Kalahari Group sediments, or by replacement of pre-existing calcretes, themselves developed in Kalahari Group sediments.

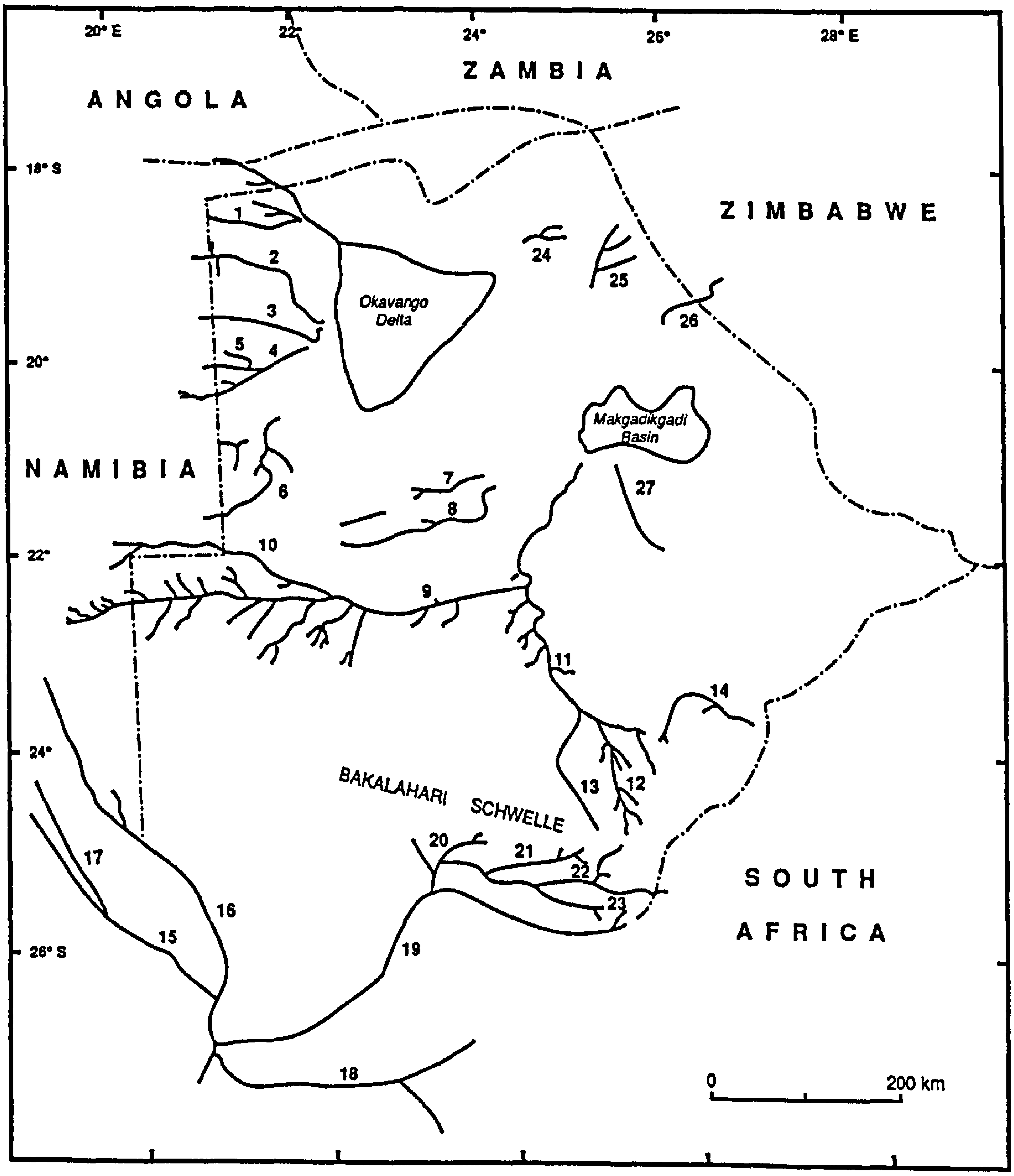
14. Discriminant analysis of silcrete samples from the Kalahari and from data included in Summerfield (1983*d*) for samples from the Cape Coastal zone statistically supports Summerfield's (1983*a*) qualitative suggestion that the main distinction between silcretes from these two regions is in TiO_2 content. Summerfield suggests that Cape silcretes developed by relative enrichment in TiO_2 under humid climates. However, on the basis of the preceding conclusion, the use of such geochemical distinctions to imply palaeoenvironmental differences at the time of formation of the two regional groups of silcretes may be erroneous. The Kalahari Group sediments are known to be predominantly composed of quartz and occasional feldspars together with trace heavy minerals, whilst Cape Coastal silcretes developed in weathered bedrock. As such, the differences between silcrete types may be purely due to differences in host materials.

Establishing a chronology for the development of *mekgacha* will potentially contribute further information to the overall picture of landscape development emerging from the Kalahari. The most obvious approach to future chronological studies will be to consider the evidence from valley fill sequences relative to other landforms. However, any such consideration of valley evolution based upon such a methodology will require extensive excavation or coring of sedimentary deposits, and will only provide spatially limited information for particular valleys. Additionally, such methods will only indicate periods of erosion and deposition which have occurred within a valley setting, not the overall development of the valley *per se*.

At present, the formulation of such a chronology is potentially limited due to both the availability of radiometrically datable material and the timescales over which valley development has taken place. With the exception of the dating of cave speleotherms using U-series techniques, the chronology of the Kalahari has, thus far, mainly considered changes which have taken place within the 100,000 year timespan accessible by radiocarbon dating. Potentially, the most useful information for identifying the timescales of *mekgacha* development will be gained from further studies of duricrusts, applying dating techniques such as ESR to silcrete and calcrete samples. However, as cautioned above, any such development must take into account the multiple stages of development which may be present within any one sample.

From these summary conclusions it is apparent that the development of Kalahari *mekgacha* has occurred over considerable timescales by a complex combination of groundwater and fluvial processes, with the relative importance of each process varying both spatially and temporally. The identification of the relative role of each process is problematic, particularly in the Kalahari where fluvial activity appears to have occurred comparatively recently, thus overwriting evidence for the role of groundwater processes.

A further problem in the identification of the past operation of groundwater processes, particularly sapping, is that, unlike erosion by fluvial activity, landform evolution by sapping does not leave a sedimentary signature. This factor alone highlights the importance of using a multi-disciplinary approach in order to give a greater indication of the operation of past processes. The identification of the role of deep-weathering in valley development requires detailed information from boreholes, which are presently extremely limited in the Kalahari in both their accuracy of logging and spatial extent. However, it is possible that with the further development of boreholes throughout the Kalahari the importance of the role played by groundwater processes in the landform development of the region will become increasingly recognised.



MEKGACHA NETWORKS

1 NCAMASERE
2 XAUDUM
3 QANGWADUM
4 EISEB
5 GCWIHABEDUM
6 GROOT LAAGTE
7 ROOIBRAK/PASSARGE
8 DECEPTION
9 OKWA
10 HANEHAI
11 MMONE/QUOXO
12 LETLHAKENG
13 NALEDI
14 SEROROME
15 AUOB
16 NOSSOP
17 ELEPHANTS
18 KURUMAN
19 MOLOPO
20 MABUASEHUBE
21 UKWI
22 MOSELEBE
23 SEKHUTANE
24 GHAUTAMBI
25 NUNGA
26 LEMEMBA
27 LETLHAKANE